Geophysical Studies in Southern and Central Rockall Trough, Northeast Atlantic

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DEDICATION

To my father
Colin Dicker (1926-1963)
For whom I have achieved all this.

EPIGRAM

" In a word, new matter offers to new observation, and they who write next, may perhaps find as much room for enlarging upon us, as we do upon those that have gone before ".

Daniel Defoe, 1724

ABSTRACT

This study is based on the interpretation of approximately 8600 line km of underway marine geophysical data over southern and central Rockall Trough. The data consist of seismic reflection profiles, and gravity and magnetic anomaly observations. Some of this data was newly acquired for this work; the remainder was made available from other sources. This work represents the first time that most of the geophysical information from this area has been inspected and integrated as a whole.

The bathyal bathymetric deep that is Rockall Trough is bordered by rifted passive continental margins. These are backed by continental crust of fairly typical thicknesses - about 30 km. Rapid attenuation of this crust occurs across the steep continental margins over a distance of 50-60 km. Beyond the margins previous gravity and seismic refraction studies, and the free-air gravity modelling presented here, define a 200-250 km wide zone of thin crust flooring Rockall Trough. This thin crustal layer varies in thickness from 5 km (locally) to 8 km but it is generally fairly uniform at about 7 km. The Moho typically models at 14-15 km depth.

Conflicting interpretations have been proffered for the thin crustal layer. This research supports the balance of recent opinion which favours an oceanic origin and evolution for Rockall Trough, though perhaps under less than normal conditions of accretion. But the possibility of highly stretched continental crust being present cannot be excluded. The size and geometry of the Trough, the oceanic layer 2 appearance on many seismic profiles, and the structures defined by the gravity models are most readily interpreted in terms of an oceanic origin. The absence of a coherent oceanic magnetic anomaly pattern is accounted for by advocating the lack of an ordered spreading ridge - transform system and a diffuse style of accretion, perhaps under high sediment accumulation rates.

Two previously unreported features are identified, mainly from the seismic reflection data. Firstly, the Barra volcanic ridge system is a large zone of arcuate volcanic ridges in southern Rockall Trough. The ridges attain heights of 3-4 km and widths of up to 40 km. This volcanic province is related to the last pulses of spreading in Rockall Trough, possibly in the E.Cretaceous. It also seems to be associated with a broad upwarp (c.2 km) in the underlying mantle, a feature defined by a conspicuous 60 mgal free-air anomaly. The Barra volcanic ridges are associated with curved magnetic anomalies having amplitudes of up to 1000 nT. Secondly, a number of sill and ?lava complexes are mapped out within the thick sedimentary fill to the Trough. These igneous events are dated as L.Palaeocene - E.Eocene from a consideration of the regional seismic and volcanic stratigraphy. It is likely that they are coeval with the Thulean igneous episode and the onset of spreading between Greenland and Rockall Plateau.

The continuation of the Gibbs F.Z. eastward into the Clare Lineament and thence to the base of the continental margin is proven for the first time. Detailed mapping of the seismic basement here has clarified the nature and significance of the Clare Lineament as an ocean-ocean or ocean-continent fracture zone. This combined fundamental discontinuity marks the southern tectonic limit of Rockall Trough and separates it from the mature Late Cretaceous ocean crust to the south in Porcupine Abyssal Plain.

The seismic stratigraphy in the Rockall area is reviewed and a stratigraphic/reflector framework developed for the Trough based on slight modifications of previous schemes. The R4 reflector frequently discussed in the literature concerning this area is reassigned a R2 label and given a new E.-M. Miocene age. There does not appear to be a well developed L.Eocene -E.Oligocene reflector (i.e. R4) in Rockall Trough. A distinct regional reflector R5, dated roughly as earliest Eocene, marks the onset of differential deposition in Rockall Trough and the construction of Feni Ridge drift. A conspicuous reflector R7 (? M.Albian) defines the top of a deep, seismically well layered sedimentary sequence that is present in the Trough but absent over oceanic crust south of the Gibbs F.Z. - Clare Lineament.

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ACKNOWLEDGEMENTS

My warmest thanks are extended to my supervisor, Dr Roger Scrutton, who for over three years maintained a high level of interest, commitment and encouragement toward this research project. In a working environment where the eccentricities and idiosyncracies of a scientific department seemed only too prevalent it has been a great pleasure to work alongside someone who still appreciates the personal and leisure aspects of academic life, and who values the art of everyday communication. In addition, his astute and thought-provoking conversations concerning matters geological were a constant source of direction and are, I hope, reflected in this piece of research.

I am also greatly indebted to Doug Masson, Lindsay Parson and Pete Miles at the Institute of Oceanographic Sciences, Surrey for their interest and enthusiasm for this project. Their combined knowledge and expertise on passive continental margins, especially those bordering the North Atlantic, has been of enormous benefit to me. I am particularly grateful to Doug and Hilary Masson for their marvellous open hospitality on two occasions when it was necessary for me to work at I.O.S. for extended periods.

I thank the Master, Officers, technicians and crew of R.R.S. "Challenger" for their professionalism at sea from 6 to 21 June 1984, during which scientific cruise data was collected for this study. Dr Martin Sinha and his colleagues from the Bullard Laboratories, Cambridge University uncomplainingly provided scientific assistance throughout the cruise. And Mick Geoghegan, acting as Irish observer, entertained all aboard with his contagious leprechaun humour.

Dr Ian Hill and Rob Young of Leicester University kindly invited me on their scientific cruise around the Cape Verde Archipelago, off West Africa during December 1982 and January 1983. I heartily thank them and the Master, Officers, scientists, technicians and crew of R.R.S. "Shackleton" for making the cruise so enjoyable and educational. As a first-hand introduction to the methods, collection and processing of underway geophysical information it enabled me to prepare better for my own data collecting cruise in June 1984.

I am grateful to the staff and technicians at N.E.R.C. Research Vessel Services base in Barry, S.Wales, in particular Doriel Jones, for their able and concerned assistance both at sea during the two aforementioned cruises and also at later times on dry land.

I gratefully acknowledge the staff of the British Geological Survey, Murchison House, Edinburgh who unhesitatingly provided their services for, or access to, special reproduction and automatic digitising facilities, and also unusual items of literature in their excellent library. Special thanks are due to Mr Dave Smythe of B.G.S., Hydrocarbons Unit for his interest in this research, his numerous valuable comments and suggestions, and for allowing me both to look at the product of his own investigations in the Rockall area and to inspect a number of important multichannel seismic profiles in my own study area.

I wish to thank Dr Alan G Nunns and Dr Graham Westbrook, formerly of Durham University, for their assistance with and permission to use the FORTRAN magnetic anomaly inversion program (Chapter 7) written by the first-named person.

Edinburgh Regional Computing Centre, Edinburgh University provided computing facilities and frequent related advice, without which it would have been impossible to perform a large proportion of the manipulation and interpretation of the gravity and magnetic anomaly data (Chapters 4,6 and 7).

In addition to the underway geophysical data collected during the "Challenger" 1/1984 cruise in Rockall Trough I was fortunate indeed to be able to inspect or utilise a considerable amount of multi-channel seismic reflection, gravity anomaly and magnetic anomaly information held by the Institute of Oceanographic Sciences, Surrey and also the Irish Department of Industry and Energy, Dublin. Without the incorporation into this thesis of the good quality material from the former source, and the accompanying discussions with and foresighted advice of Doug Masson, this study would have been a far sadder and simpler affair.

During my short residence at Edinburgh University I have benefited from conversations with many people: notably, Professor M.H.P.Bott of Durham University; Dr Roger Hipkin of the Geophysics Department here for valuable advice and access to computer programs regarding matters gravitational; and the teaching staff and postgraduates at the Grant Institute of Geology.

I owe a large vote of thanks to Colin Chaplin and Diana Baty for their superb conscientious work in the photographic and drawing departments, especially when often faced with large strangely-shaped material! Mrs Thea Grieve provided assistance on countless occasions when trying to locate difficult references in the library. And Peder Aspen kindly helped me to locate a variety of bathymetric charts in the map collection.

A special thank you is in order for the secretaries at the Grant Institute, Marcia Wright, Patricia Stewart, Denise Wilson and Heather Hooker, who have always dealt with my enquiries, difficult-ies and complaints with a smile. Their constant light-heartedness and jocularity have undoubtedly served to create a more pleasurable working atmosphere. I am particularly in the debt of Denise Wilson for her rapid and accurate typing of this thesis, a task she performed with splendid flexibility.

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Finally I would like to express my gratitude to the residents of the City of Edinburgh, a remarkale place in so many respects, for making my three years and more stay there such an enjoyable, fulfilling and learning time...thank you.

1. INTRODUCTION: LAYING THE FOUNDATIONS

1.1 Scientific curiosity or future harvest?

Why study the geology of Rockall Trough? In reply to this prompt one could argue simply that the acquisition of further scientific knowledge is in itself grounds for justification. However, an investigation of this small ocean basin - ocean being used in the broadest sense - would seem to hold more import than just academic curiosity. The main objective of this work has been to understand the general, large-scale geological evolution of the southern and central Rockall Trough (some people prefer to call it Rockall Channel) and its adjacent continental margins and shelf during much of the long Phanerozoic period. The nature of the data used in this research and their interpretation do not allow the finer details of the area's history to be resolved with confidence. It has been a high priority to step back and view the Rockall Trough in its broader regional context; namely to consider its significance in the development of the North and, less obviously, the Central Atlantic Ocean since Late Palaeozoic times. This is important when one recognises that the true origin and composition of Rockall Trough, and its evolutionary position within the framework of North Atlantic continental drift, are as yet poorly understood.

In addition, the importance of the considerable thicknesses of sediment in certain areas of the trough as potential sources of hydrocarbons cannot be overlooked, despite the large technological difficulties which would have to be surmounted in order to produce oil and gas from a sedimentary basin where water depths are an order of magnitude greater than the hydrocarbon-producing areas of the North Sea. In fact there has been much interest shown recently in the hydrocarbon prospectivity of the Rockall Trough, particularly to the north-east, and with the arrival of the tension-leg drilling platform some exploratory work may be witnessed in the near future. However, it seems that development and production in such a deep-water area is a long way off yet.

Thus in Rockall Trough we see a large, deep basin whose geological history remains controversial; indeed considered together with its north-easterly continuation, the Faeroe-Shetland Channel, it is distinct in being the only large basin on the Northwest European continental margin and shelf whose crustal geology has yet to be convincingly accounted for and understood. Viewed in these terms, and bearing in mind the economical and political implications of any significant hydrocarbon discoveries, it is easy to recognise the value of an improved understanding of thecrust and sediments constituting Rockall Trough.

In order to realise the aims set out above a programme of data acquisition was undertaken involving marine geophysical remote sensing . The three standard geophysical survey methods employed were continuous seismic reflection profiling, gravimetry, and total intensity magnetic field measurements, and these are described in Chapter two. The newly acquired data have been supplemented by existing geological and geophysical information from a variety of sources; but the limitations of these and the new results, and also of the modelling and interpretation methods available, has meant that no special emphasis has been achieved in any one field of the research. Rather, an integrated approach has been sought whereby information from many branches of marine geology has been considered together to establish a hopefully coherent and tenable geological picture of the Rockall Trough.

1.2 Geography and Bathymetry of Rockall Trough

The south-west to north-east trending Rockall Trough lies beyond the continental slope to the west of Ireland and northern Scotland. The region covered by this work falls within the area defined by the 51°N and 56°N parallels and the meridians 12°W and 20°W (roughly 310,000 sq. km, Fig. 1.1); within this window detailed geological and geophysical coverage extends over approximately 230,000 sq. km (Chart 1), an area equivalent to that of England, Scotland and Wales combined (c. 229,500 sq. km). Hence we are concerned with a research area three to four orders of magnitude greater than that of most land-based studies, although considered within the realm of marine earth sciences it is by no means an unusually large exercise.

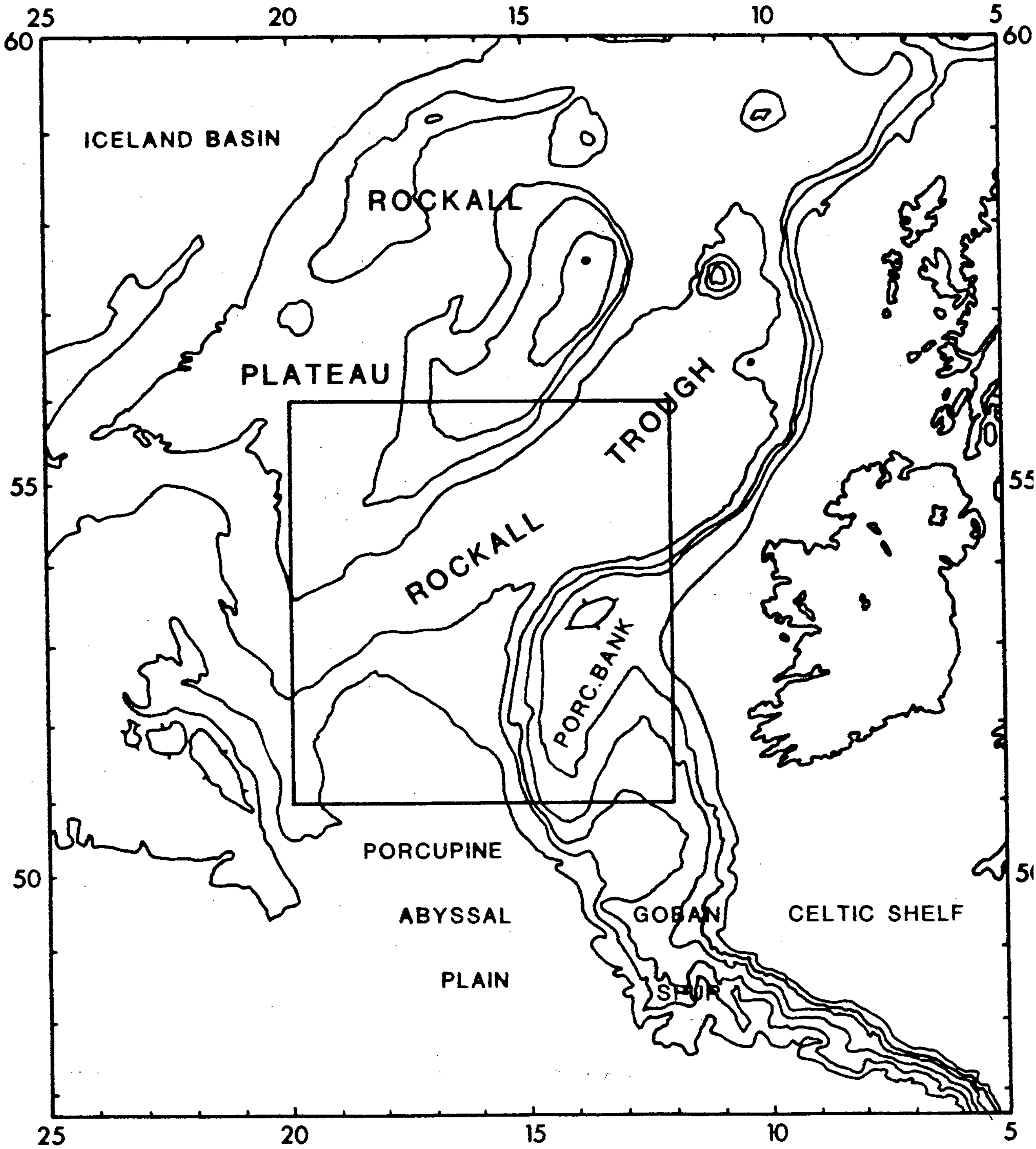


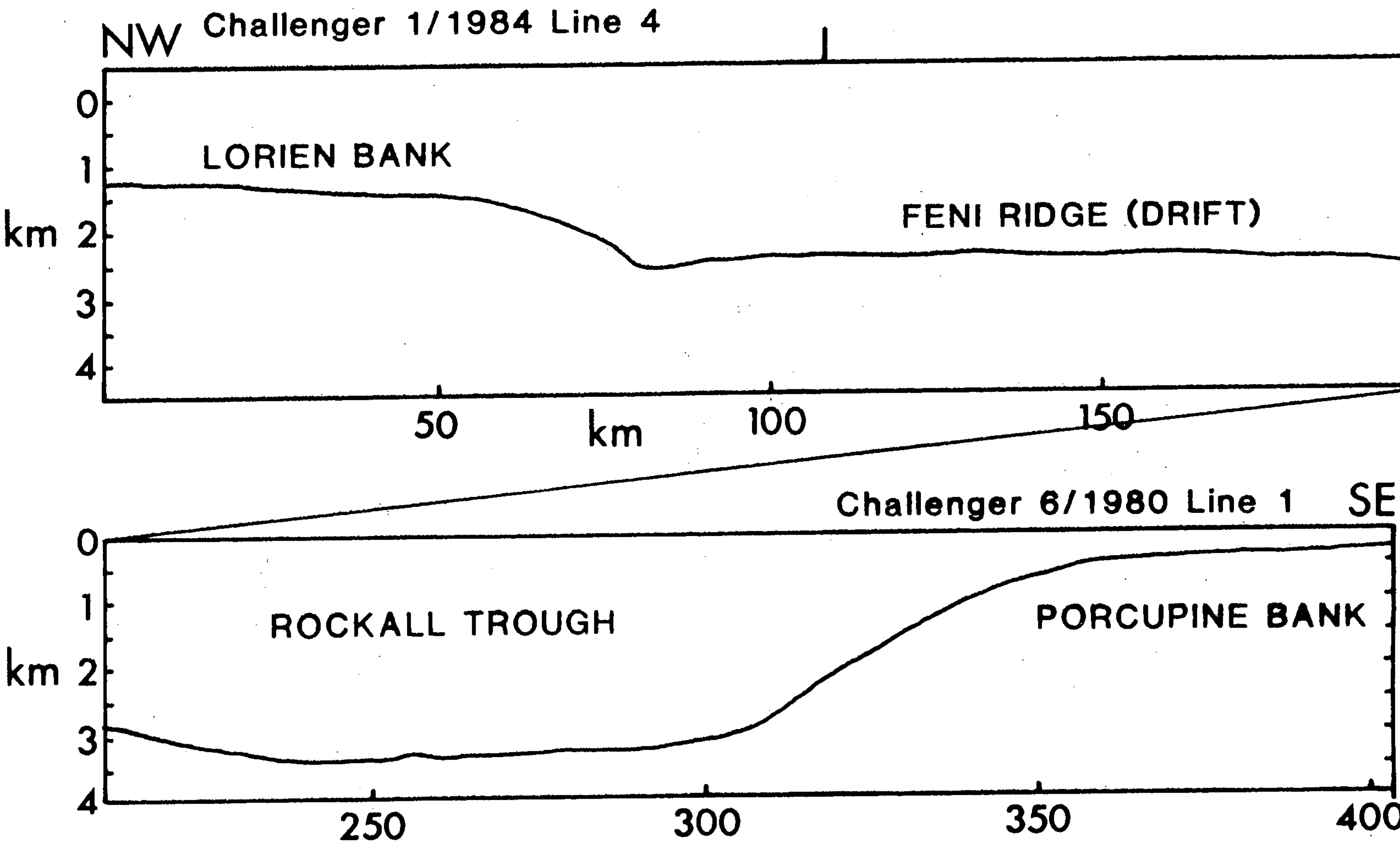
Figure 1.1 General bathymetry around the west UK continental margin and Rockall Plateau. Inset box marks the area covered by this study. 200, 500, 1000, 2000, 3000, and 4000 m isobaths.

The history of hydrographic surveying around the Rockall Trough region has been sporadic. The earliest indications - albeit slender - that a deep channel was present came to light following bathymetric surveys associated with the search for trans-Atlantic telegraph cable routes during the second half of the nineteenth century. A long period of relative oceanographic inactivity ensued and it was not until the 1960s that a renewed international effort to chart the waters of the Northeast Atlantic was witnessed. These surveys, particularly of the German and UK Hydrographic Departments, were responsible for detailing most of the bathymetric features we know of to-date (Figs 1.1 and 1.4).

Thus we observe that the Rockall Trough separates the shallow continental shelf west of Ireland and Scotland from the submerged banks and basins of the Rockall Plateau - Faeroe Shelf region. Rockall Bank, Hatton Bank and the intervening Hatton-Rockall Basin each has the same trend as Rockall Trough, a trend which is continued by the line of George Bligh, Lousy, Bill Bailey's and Faeroe Banks and thence to the Faeroe Shelf (Fig. 1.4). Excepting the north-west to south-east oriented Wyville-Thomson Ridge the Rockall Trough finds a north-easterly continuation in the Faeroe-Shetland Channel and ultimately the Norwegian continental margin. Consequently it is pertinent that the latter two provinces should be taken into account in any discussion of the Rockall Trough. The Porcupine Seabight, which lies to the south-west of Ireland, is seen to have elements of both a N-S and a NE-SW trend, attains comparable depths to the Trough (down to 4000 m), and as such should provide clues to the development of the region as a whole.

Rockall Trough itself is bounded by distinct steep continental slopes that are developed over a breadth of some 50 km. On the south-eastern margin the shelf extends to depths of 200 m except in the south-west over Porcupine Bank where it is poorly developed and the shelf break occurs at approximately 400 m. The slope has a gradient mostly varying within the range 5° to 15° (Roberts 1975; measurements on Edinburgh profiles, Fig. 1.2); local flattening or

Figure 1.2 Bathymetric profile across Rockall Trough along tracks C 80-1 and C 84-4 (see Chart 1 for location). Vertical exaggeration is roughly 10 times.



steepening of the slope results from down-slope failures, numerous examples of which have been detected by sidescan sonar studies (Kenyon 1985) and conventional seismic reflection profiles. On this SE margin the continental slope usually passes abruptly into the gently undulating continental rise or sedimentary apron which floors the Trough (Fig. 1.2). This sudden change in gradient is not seen north-west of Ireland where the Barra and Donegal Fans extend towards the axis of the Trough around the Hebrides Terrace Seamount.

The north-western margin of the Trough shows a continental slope whose gradient is usually less than 5°, the main exception being that length of margin immediately south and east of Rockall Island where gradients may exceed 15° (Roberts 1975). The large re-entrant between Rockall and Lorien Banks is discussed in chapter 5.

The shelf is poorly developed on Rockall Bank, its only surface expression being Rockall Island (first recorded in 1871; Sabine 1960). The shelf break appears at about 300 m, beyond which the continental slope, devoid of canyons, descends to 2-2.5 km where it passes into a reasonably well developed continental rise - at variance with the SE margin. The continental rise here results from the presence of the sinuous Feni Ridge which can be traced along the north-western side of Rockall Trough and, in diminished form, as a broad curve up to the Wyville-Thomson Ridge (Fig. 1.4). At its south-western termination it is deflected to the NNW and SSE as the Isengard Ridge. On the north-western side of Rockall Plateau the Hatton Bank is outlined by the 1000 m contour, although it shoals to less than 500 m over a small area. The Hatton-Rockall Basin deepens from roughly 1200 m in the NE to 1500 m near Fangorn Bank in the SW.

Within the Trough and away from the Feni Ridge, whose crest lies for the most part at 2300 m, a progressive increase in depth is observed from 2000 m around Anton Dohrn Seamount in the north-east, to 4000-4500 m in the Porcupine Abyssal Plain to the south-west. There is no bathymetric feature present which clearly defines the southern limit of the Trough, merely a gradual slope of its floor towards abyssal depths; depths in the Trough are comparable to those in the Iceland Basin to the north-west of Rockall Plateau, so justifying its label of a small ocean basin. The only perturbations to the fairly gentle topography in the Trough are the three crudely circular seamounts to the north-east: the Hebrides Terrace, Anton Dohrn and Rosemary Bank Seamounts, all of which shoal to depths less than 1000 m and have planated tops.

1.3 History of Research Activity

As in so many other studies around the globe which are encompassed under the umbrella of marine geology and geophysics, strong interest in the origin, structure and evolution of Rockall Trough and its surrounds did not develop until a rational basis had evolved around which projects could be coordinated and set in context; such a basis - the theory of plate tectonics - was propounded in the mid-and late-1960s, and it was not until then that a concerted investigative effort was witnessed in this region. This effort increased throughout the 1970s consonant with the refinements that were being made both to the new global theory and, equally as important, to the techniques of geophysical remote sensing at sea. The first half of this decade has seen a somewhat reduced, although steady, interest maintained in the marine geology of this part of the European margin.

The proposal and confirmation of Rockall Plateau as a fragment of continental material largely separated from the main UK continental shelf has been an important aspect in our understanding of this area; important because it implies that there has been motion across the Trough. Miller and Mohr (1965), in obtaining an age of 60 ± 10 Ma for aegirine-acmite samples from the Rockall aegirine granite, placed the Island in the context of the Early Tertiary intrusive activity manifested in NW Scotland and NE Ireland (as had Cole as early as 1897). In addition they proposed that "the trough between Rockall Bank and the Hebrides could be a result of crustal separation of an originally unified landmass". And further. "The possibility of the bank being a continental fragment cannot be ruled out". Bullard et al. (1965), in the same year, incorporated these ideas in their reconstructions of the continental masses around the whole Atlantic. Their retaining the Rockall Plateau to fill in a gap between the pre-drift Greenland and Europe margins added weight to the argument that the feature was a submerged continental fragment.

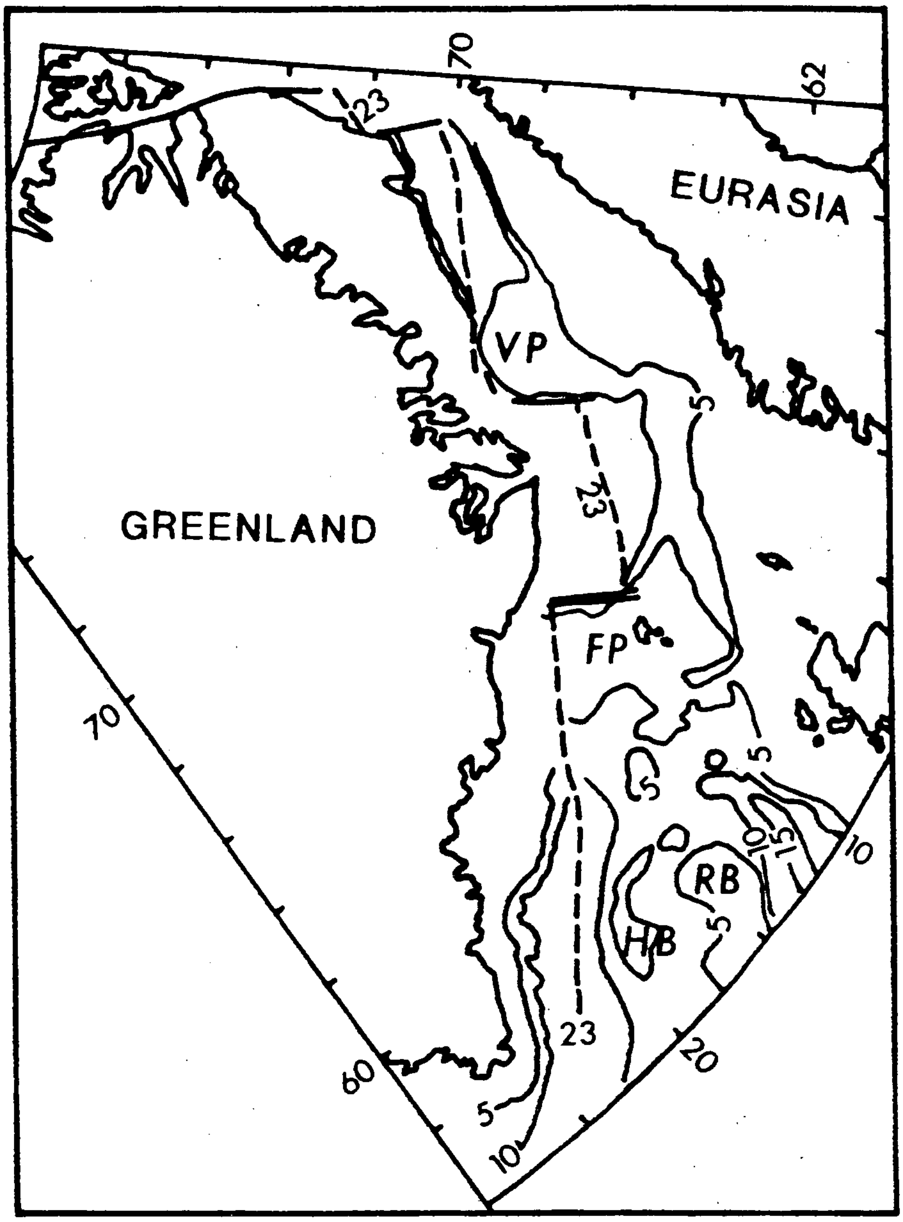


Figure 1.3 Anomaly 23 (52 m.y.B.P.) reconstruction of Greenland and Eurasia showing the extent of the proposed Rockall-Faeroe microcontinent. Bathymetry in 100s of fathoms. FP, Faeroe Platform; HB, Hatton Bank; RB, Rockall Bank; VP, Voring Plateau. Redrawn from Nunns (1983).

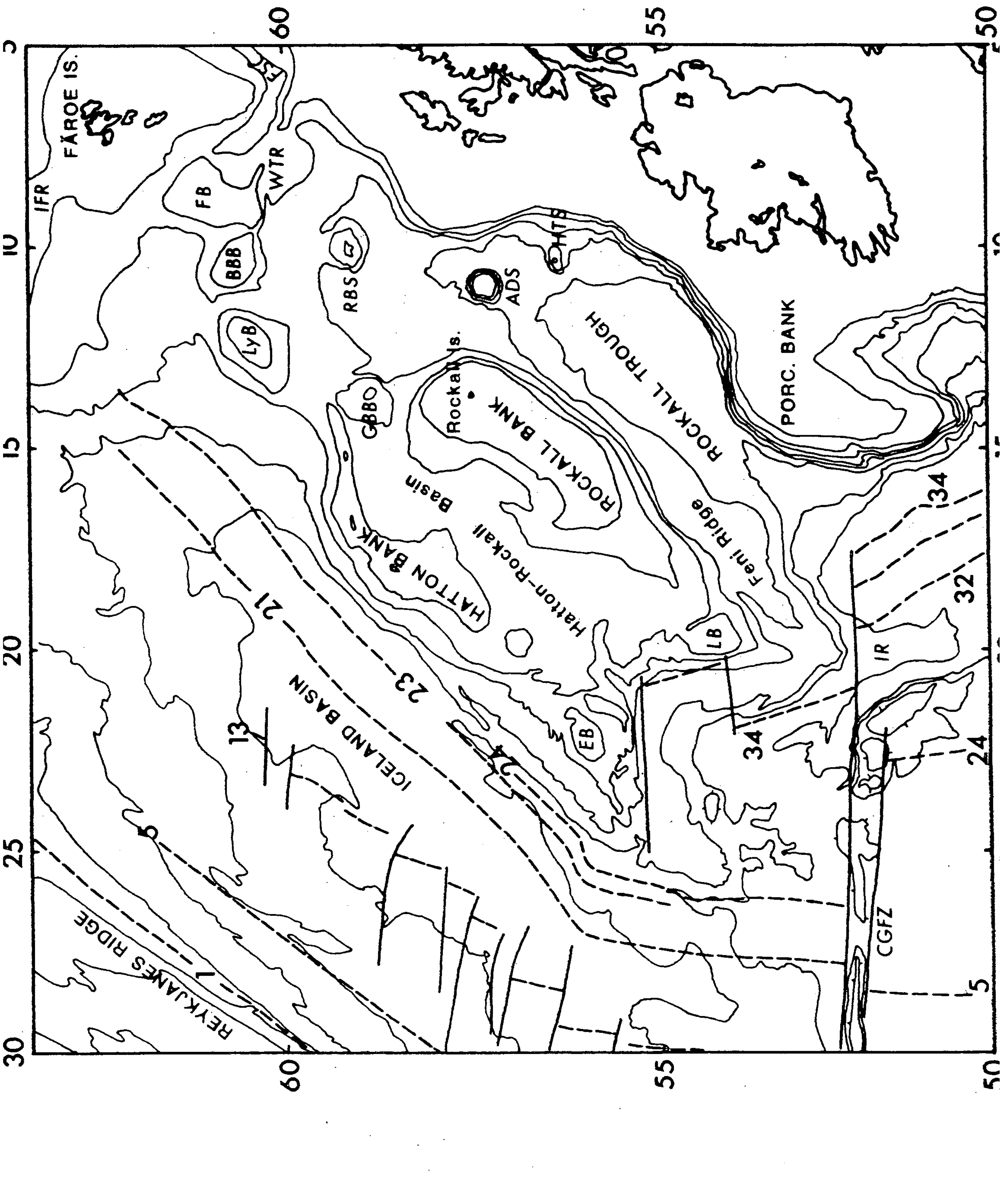
Evidence of a less speculative nature has come from more recent geophysical and geological studies. In the former case seismic refraction investigations (Scrutton 1970, 1972) have established the continental thickness of the crust beneath the plateau - 31 km for Rockall Bank, 22 km for Hatton-Rockall Basin - and also compressional wave velocities incompatible with an oceanic crustal origin. Seismic reflection, gravity and magnetic surveys of the Plateau have been performed on a number of cruises, in particular those of H.M.S. Hecla and R.R.S. Discovery in 1969 (Roberts 1971, Roberts and Jones 1978).

Geological evidence, which is generally less equivocal than geophysical information, is sparse from Rockall Plateau. Moorbath and Welke (1969) performed strontium and lead isotope analyses on samples from Rockall Island and concluded that at least part of the Rockall Bank has continental affinities and may consist of Precambrian basement intruded by at least one Lower Tertiary igneous centre. Dredging and drilling on Rockall Bank have retrieved granulite-facies metamorphic rocks of Laxfordian (in the north) and Grenvillian (in the south) age (Roberts et al. 1972; Roberts et al. 1973; Miller et al. 1973), so fuelling thoughts that the Grenville orogenic front traverses the Plateau.

Roberts et al. (1970) first reported on the structure and sediments of Hatton-Rockall Basin, an area which was the focus of further detailed investigation on Leg 12 of the Deep Sea Drilling Project (Laughton, Berggren et al. 1972) whose primary objective was to derive the subsidence history of this continental fragment in relation to the North Atlantic evolution (like Orphan Knoll on the same cruise).

As suggested in the previous section there appears to be a bathymetric continuity of the Rockall Plateau north-eastwards to the Faeroe Platform. Bott et al. (1974) recognised the fact that continental reconstructions of the Greenland Sea were best satisfied by maintaining the region beneath the Faeroe Islands as continental crust (Fig. 1.3); their explosion seismology experiment stretching from northern Scotland to Iceland identified 27 to 38 km thick crust beneath the islands which was characterised by P-wave velocities with continental affinities.

Figure 1.4 (opposite). Detailed bathymetry and oceanic crustal provinces around the Rockall area. Numbered dashed lines are anomaly identifications. ADS, Anton Dohrn Seamount; BBB, Bill Bailey's Bank; CGFZ, Charlie-Gibbs Fracture Zone; EB, Edoras Bank; FB, Faeroe Bank; FSC, Faeroe-Shetland Channel; GBB, George Bligh Bank; HTS, Hebrides Terrace Seamount; IFR, Iceland-Faeroe Rise; IR, Isengard Ridge; LB, Lorien Bank; LyB, Lousy Bank; RBS, Rosemary Bank Seamount; WTR, Wyville-Thomson Ridge. 500 - 4000 m isobaths everywhere at 500 m interval


FÄROE IS.
IFR
FB
WTR
BBB
LyB
RBS
ADS
HTS
GBBO
Rockall Is.
ROCKALL BANK
ROCKALL TROUGH
PORC. BANK
Feni Ridge
Hatton-Rockall Basin
HATTON BANK
ICELAND BASIN
REYKJANES RIDGE
LB
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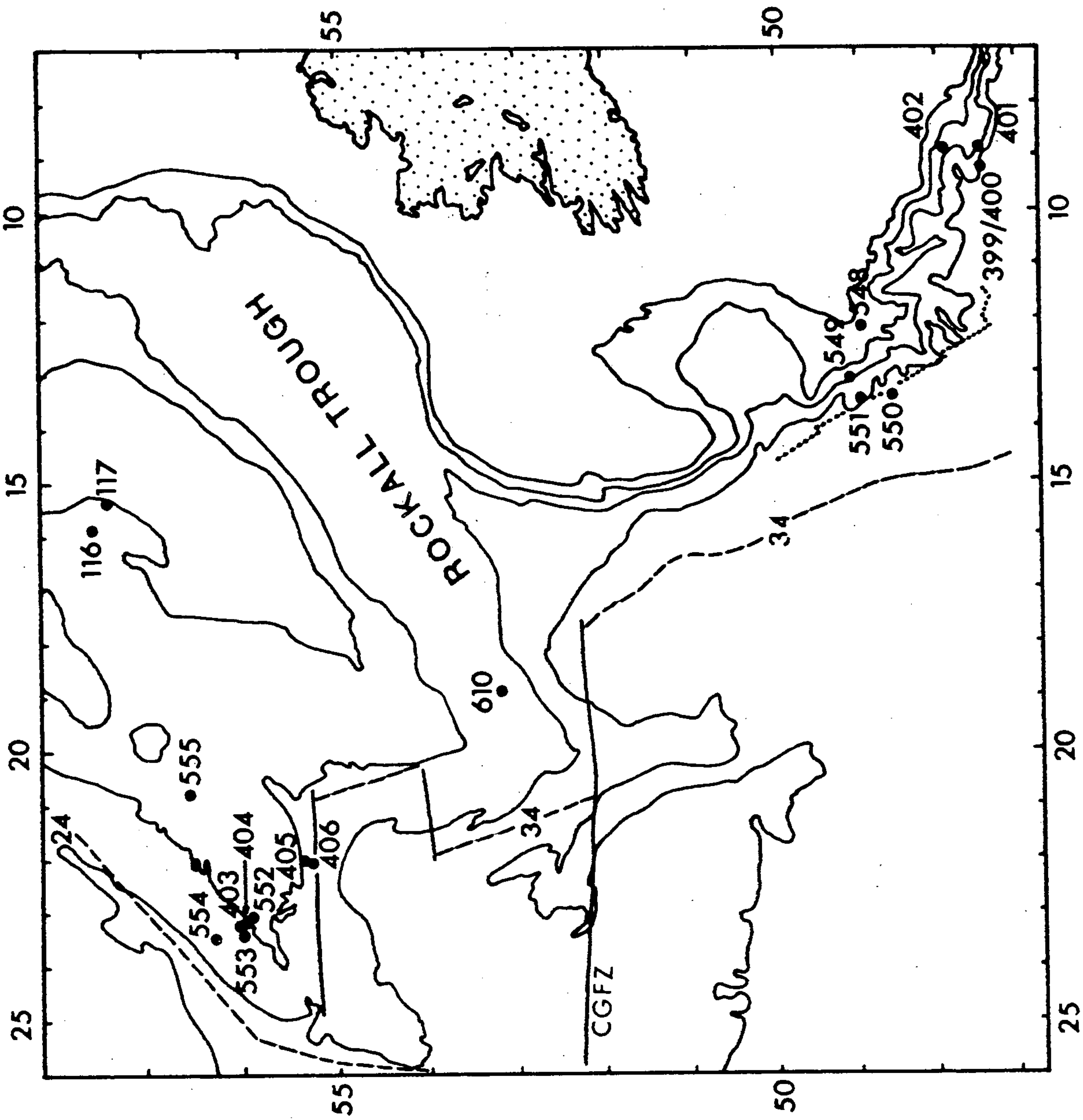
The continental shelf west of Scotland and Ireland has received rather more attention than the more distant (and deeper) regions around the British Isles. The sediments, crust and upper mantle beneath this margin have been investigated in seismic refraction and reflection studies (e.g. Bailey et al. 1974; Jones 1978; Bott et al. 1979; Brewer and Smythe 1984); further south the continental nature of Porcupine Bank has been established by explosion seismology (Whitmarsh et al. 1974) and gravity analysis (Buckley and Bailey 1975). The crustal geology and origin of the Porcupine Seabight remains contentious. Most workers (e.g. Scrutton et al. 1971; Buckley and Bailey 1975; Naylor and Shannon 1982, pp. 68-69) believe this trough, which has a thick (in excess of 5 km) Mesozoic to Recent sedimentary succession, to be a fault-bounded rift floored with continental crust, albeit very thin in places. On the basis of magnetic data Lefort and Max (1984) argued that three small zones of oceanic crust could be recognised in the Seabight, and consequently that the Porcupine Bank constituted an important and discrete segment in the rifting evolution of the Irish margin. Whichever may be true there has been notable commercial hydrocarbon exploration activity centred on this sedimentary basin.

A structural framework and tectonic evolution of the Irish shelf has been suggested by Riddigough and Max (1976) and Max et al. (1982) which emphasises the importance of the offshore continuation of major structural lineations and provinces mapped in Ireland (e.g. Max et al. 1983). Similarly, recent geophysical (Dingle and Scrutton 1979; Scrutton 1979) and geological work (DSDP Leg 80; Graciansky, Poag et al. 1985) has established a tectonic history for the marginal plateau at Goban Spur. Megson (1983) investigated, by geophysical methods, the continental margin from southern Porcupine Seabight to southern Rockall Trough, and its transition into the oceanic realm to the west.

The distribution and age of mature oceanic crust around the Rockall region is illustrated in Figure 1.4. There is an abundance of literature pertaining to the geology and geophysics of these areas of oceanic accretion and the author does not feel it necessary to catalogue all of it; instead the reader's attention is directed to the following more recent contributions: Le Pichon et al. (1977), Srivastava (1978), Kristoffersen (1978), Nunns (1983), and Masson and Miles (1984). Much of the initial magnetic surveying in the vicinity

Figure 1.5 DSDP drilling sites in the Rockall area. 1,2,3 and 4 km isobaths. Dotted line off Goban Spur is continent - ocean transition of Masson et al. (1985). Also shown are oldest ocean magnetic anomalies and fracture zone trends.



was performed by the U.S. Naval Oceanographic Office (Vogt and Avery 1974) and the UK Institute of Oceanographic Sciences (Roberts and Jones 1975; Chart 3).

By virtue of its position the Rockall area affords ample opportunity for addressing the queries and perplexities associated with passive continental margins. The Deep Sea Drilling Project has occupied drill sites over or across the ocean-to-continent transition during Legs 48 (Montadert, Roberts et al. 1979) and 81 (Roberts et al. 1982; Roberts and Ginzburg 1984; Roberts, Schnitker et al., in press) on Rockall Plateau (Fig. 1.5), and during Leg 80 (Graciansky, Poag et al. 1985) on Goban Spur (Fig. 1.5).

The earliest work pertaining to the Rockall Trough proper, other than some limited submarine pendulum gravity measurements made in the 1930s, is that of Hill (1952) who observed three short closely spaced refraction lines over the Feni Ridge at 53°50'N 18°40'W (Fig. 1.6). From the compressional wave velocities he deduced, correctly in part, that a sedimentary succession, ranging between about 2 and 3 km thickness, was present above basement with continental affinities. Ewing and Ewing (1959) occupied two unreversed refraction stations in the Trough (Fig. 1.6) and calculated similar seismic velocities to those of Hill. Although as much as 5.5 km of sediment were encountered at their E10 site they did not propound any clear explanation of the basement refractors.

In 1966 Vine was to place Rockall Plateau and Trough in the much broader context of sea floor spreading in the N. Atlantic as an extension to the speculations of Bullard et al. (1965) and earlier workers; he proposed that the Trough may represent an initial abortive rift, subsequent persistent spreading being taken up across the extant Reykjanes Ridge. Although he did not explicitly propound the existence of oceanic crust in the Trough, the implication as such has been the subject of enquiry in most subsequent research. Seismic refraction and gravity studies (Scrutton 1970, 1972; Scrutton and Roberts 1971; Bott and Watts 1971) drew attention to the very thin and subsided nature of the crust beneath the trough and the rapid attenuation of the continental crust at its margins. The quiet magnetic signature of the Trough was noticed by Roberts (1971) and accounted for by advocating oceanic crustal accretion during the Early-Middle Mesozoic. Integrated studies, incorporating a good deal of seismic reflection profiling, are reported by Jones et al. (1970),

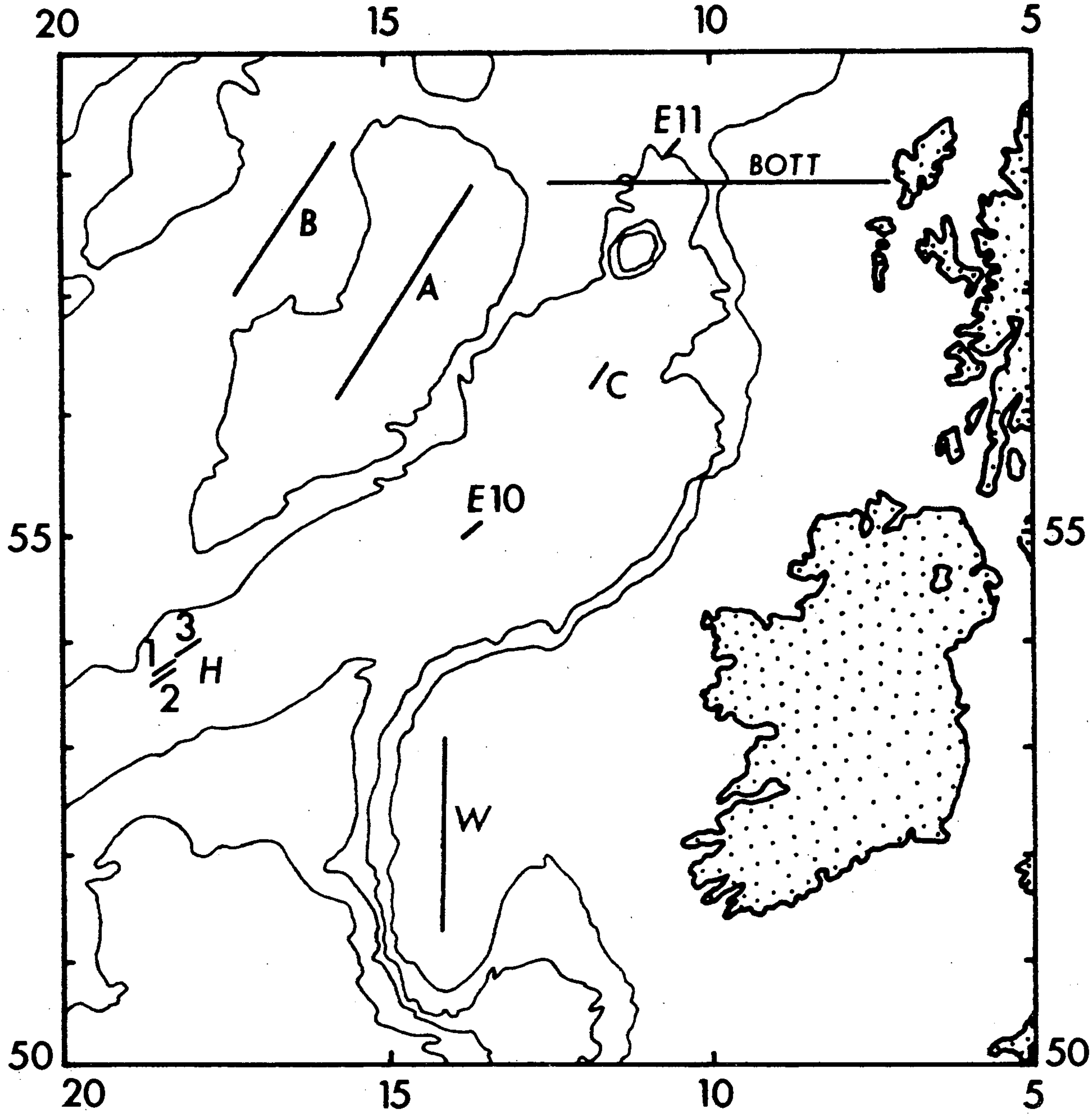


Figure 1.6 Seismic refraction stations in and around Rockall Trough. A,B and C from Scrutton (1972); E10 and E11 from Ewing and Ewing (1959); H1 - 3 from Hill (1952); BOTT from Bott et al. (1979); and W from Whitmarsh et al. (1974).

Roberts (1975) and Roberts et al. (1981) who have attempted to view the Trough in a broader regional perspective. Apart from these there has been comparatively little new data collected within the Trough outwith the domain of commercial exploration. Edinburgh University has collected underway geophysical data (seismic reflection profiling, gravity and magnetics) during scientific cruises in 1979 and 1981 in the vicinity of the mouth of Rockall Trough (Dingle et al. 1982; Megson 1983), as have the Institute of Oceanographic Sciences, UK on R.R.S. Discovery cruises 60 and 84. In addition marine geophysical data were acquired along a small number of lines over Rockall Trough and Plateau during the 1969 voyage of N.O. Jean Charcot (Scientific Group 1971).

In the northern part of the Trough the Rosemary Bank has been the focus of gravity and magnetic investigations (Scrutton 1971; Miles and Roberts 1981) and Anton Dohrn Seamount the subject of a dredging program (Jones et al. 1974). Jones and Ramsay (1982) discussed the sedimentary and volcanic stratigraphy of northern Rockall Trough, and its implications for the circulation in the Norwegian and Iceland Seas, as evidenced from dredging near the Wyville-Thomson Ridge.

It seems likely that the Rockall Trough and Faeroe-Shetland Channel have experienced similar geological and tectonic histories, such that information from one may provide clues to the other. Bott (1978) outlined the general development of the contiguous margins, and an excellent appraisal of the setting of the Channel in the wider context of the Greenland-Scotland Ridge can be found in Bott et al. (1983). Interestingly there seems to be widespread agreement that some form of incipient or nascent oceanic crust is present flooring the Faeroe-Shetland Channel (Smythe et al. 1983; Ridd 1983; Smythe 1983; Roberts et al. 1983; Price and Rattey 1984; and Bott and Smith 1984), although its exact geometry and age remain equivocal.

The finer geometry and age of the sediments in the Rockall Trough has remained problematic since Jones et al. (1970) and Roberts (1975) established their broad division of the succession in the Rockall area. Roberts and Kidd (1979) reported on long-range sidescan sonar studies (GLORIA MK II) of the sediment waves on the Feni Ridge Drift in southern Rockall Trough; site 610 of DSDP Leg 94 (Fig. 1.5; IR DSDP v94, in prep.) was occupied in the same area in 1983 with the objectives of defining the Quaternary and Tertiary stratigraphy here, the process of accumulation of sediment drifts and waves, and their relation to ocean circulation in the NE Atlantic. Miller et al. (1985) used oxygen isotope, biostratigraphic and seismo-stratigraphic data to evaluate the influence of glacio-eustasy and bottom water currents on the Tertiary sediments both in the Rockall area and on the conjugate North American margin. The structure and stratigraphy of the younger Tertiary sediments in northern Rockall Trough have been examined recently using shallow-penetration high resolution seismic profiling (Whittington and Dobson 1985).

Hydrocarbon exploration activity has been patchy in the Rockall area as regards both distribution and density of coverage. North and west of the Shetland Isles a very closely spaced grid of commercial multichannel seismic reflection lines has been obtained and also a considerable number of wells drilled (see, for example, Smythe et al. 1983). To the south, however, in Rockall Trough and on the Plateau similar geophysical and geological data are relatively scarce, and as a consequence the hydrocarbon potential remains largely unknown (Naylor and Shannon 1982). A small number of widely spaced seismic surveys and occasional lines have been commissioned by the British Geological Survey and the Institute of Oceanographic Sciences, UK in the Trough; parts of these surveys have been made available for inspection by the author (see Chart 1). In 1980 the British National Oil Corporation (BNOC) drilled a stratigraphic test well in the far north of Rockall Trough in 1355 m of water from which no results have been released to-date, though encouragingly the recent ninth round of licensing has witnessed the leasing of blocks in Rockall Trough, again in the northern areas.

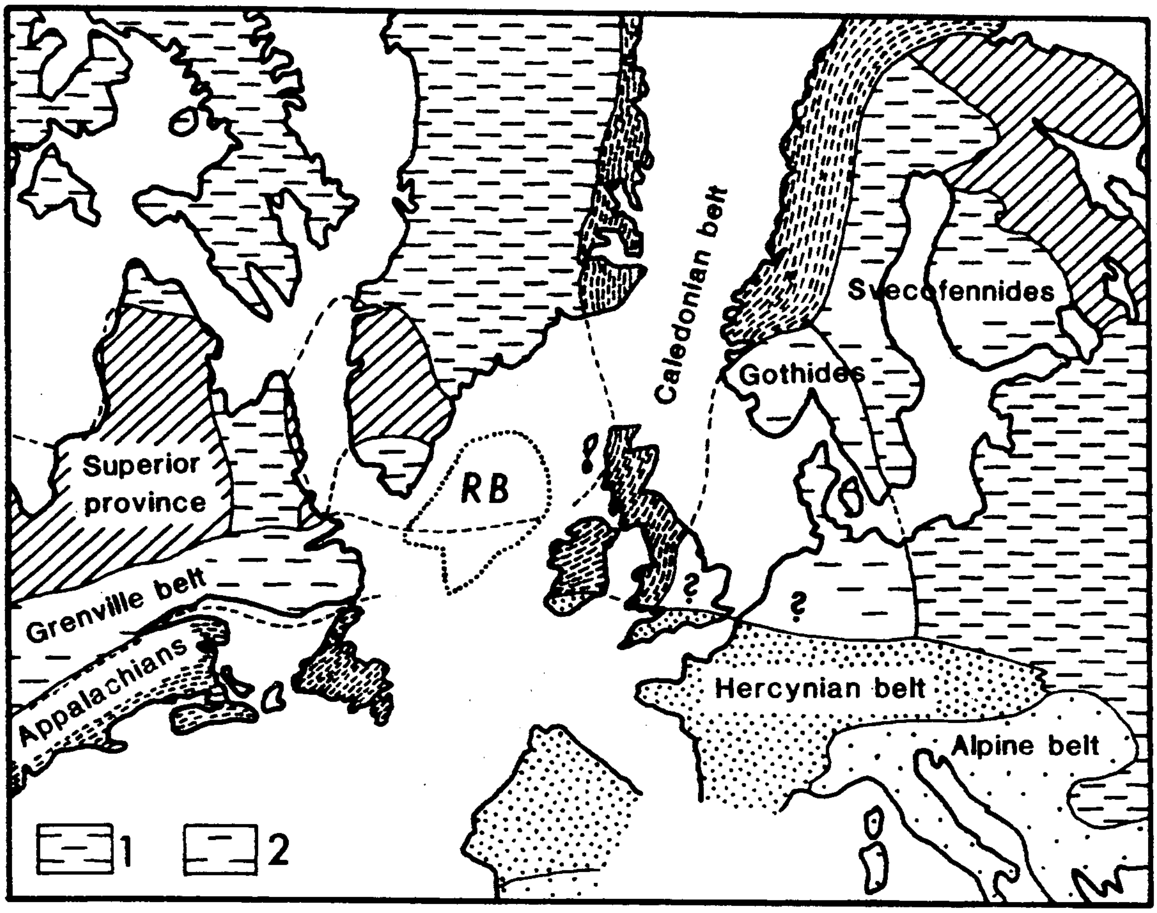


Figure 1.7 Permian pre-drift reconstruction of the N. Atlantic region (after Anderton et al. 1979). 1 = 2500-1700 Ma cratons and orogenic belts; 2 = 1300-900 Ma cratons and orogenic belts. Diagonal ornament = Archaean cratons. Fine dashes = late Proterozoic to early Palaeozoic orogenic belts. Heavy stipple = late Palaeozoic orogenic belts. Light stipple = Mesozoic and Tertiary belts. RB is Rockall Bank.

1.4 Rockall Trough: Its setting in the evolution of the North Atlantic

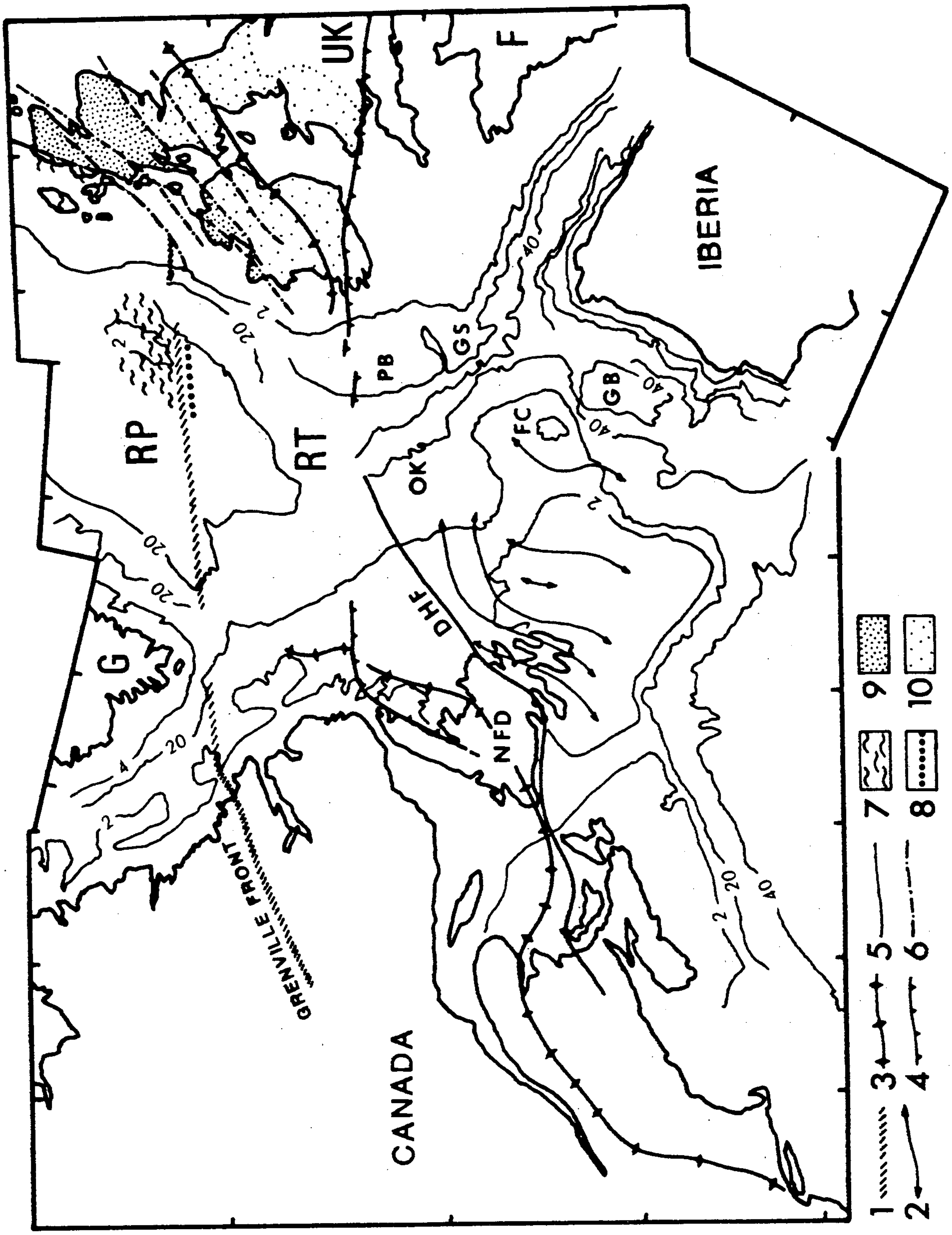
The Crustal Foundations

The mapping of linear magnetic anomalies and transform fault zones in the North Atlantic basins has permitted numerous pre-drift reconstructions of the continental masses surrounding them. Figure 1.7 shows the Permian pre-drift disposition of cratons and orogenic belts within the Laurasian segment of the Pangaean Supercontinent (Anderton et al. 1979). A similar geological situation is depicted by Ziegler (1981, Fig. 3). The British Isles are seen to straddle the Caledonian orogenic belt, while to the west the Rockall Plateau occupies an area where the offshore geology is little known between this belt and the ancient cratonic rocks of Greenland and maritime Canada . It is evident that a better understanding of the Rockall Plateau and Trough would enable a more detailed and reliable assessment of the geological and geophysical continuity across the N. Atlantic.

On a larger scale (Fig. 1.8) the paucity of knowledge regarding the offshore geology west of the British Isles is striking compared, say, with the well documented North Sea basin. The situation has not improved much since Figure 1.8 was published. What the figure does illustrate is the likely continuation of the orthotectonic and paratectonic zones of the Caledonides south-west beneath the Irish continental shelf and north Porcupine Seabight. The rocks of the Hercynian mountain belt and its associated deformation front are also thought to persist to the west where they may constitute the basement beneath the deeper parts of Porcupine Seabight and Goban Spur. There has been speculation as to whether the Hercynian Front can be linked to bathymetric and basement features observed on the western margin of Porcupine Bank (see Chapter 4); in the same vein trans-Atlantic correlations of fundamental structural lineations and provinces have been attempted by Haworth (1977, 1980) using geophysical information.

Geological sampling (Roberts et al. 1973; Miller et al. 1973) and geophysical surveying (Roberts and Jones 1978) have established that most of Rockall Plateau consists of rocks of the north-west foreland to the Caledonian orogen. Grenvillian and Laxfordian ages have been obtained from samples dredged on Rockall Bank (Miller et

Figure 1.8 (opposite). Important basement features on the N. Atlantic pre-drift reconstruction (after Roberts 1974 and Lefort 1985). 1 = Grenville Front; 2 = Precambrian ridges; 3 = suture of the Iapetus Ocean; 4 = Hercynian Front; 5 = late Palaeozoic strike-slip faulting in Canada; 6 = middle to late Palaeozoic strike-slip faulting in the UK; 7 = Laxfordian basement; 8 = Grenvillian basement; 9 = orthotectonic Caledonides; 10 = paratectonic Caledonides. Bathymetry in 100s of metres. DHF = Dover-Hermitage Fault; FC = Flemish Cap; GB = Galicia Bank; GS = Goban Spur; NFD = Newfoundland; OK = Orphan Knoll; PB = Porcupine Bank; RP = Rockall Plateau; RT = Rockall Trough.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al. 1973). Riddihough and Max (1976) proposed that a small segment of Lewisianoid basement is present at shallow depths of burial north of 53.5°N on Porcupine Bank - this based on the Precambrian inliers in nearby County Mayo and their magnetic signature offshore.

The SW to NE trends of the major structures in the north-west foreland and the Caledonian Fold belt are clearly reflected in the orientation of the margins of Rockall Plateau and Trough, Hatton-Rockall Basin and the Faeroe-Shetland Channel. This control on tectonic development by SW-NE trending ancient structures is also witnessed in the maritime provinces of Canada (Haworth 1980, 1982). In a similar fashion Roberts et al. (1979) postulated that the E-W trend of the continental margin on SW Rockall Plateau was determined by the strike of the Grenville Front in this area (Fig. 1.8).

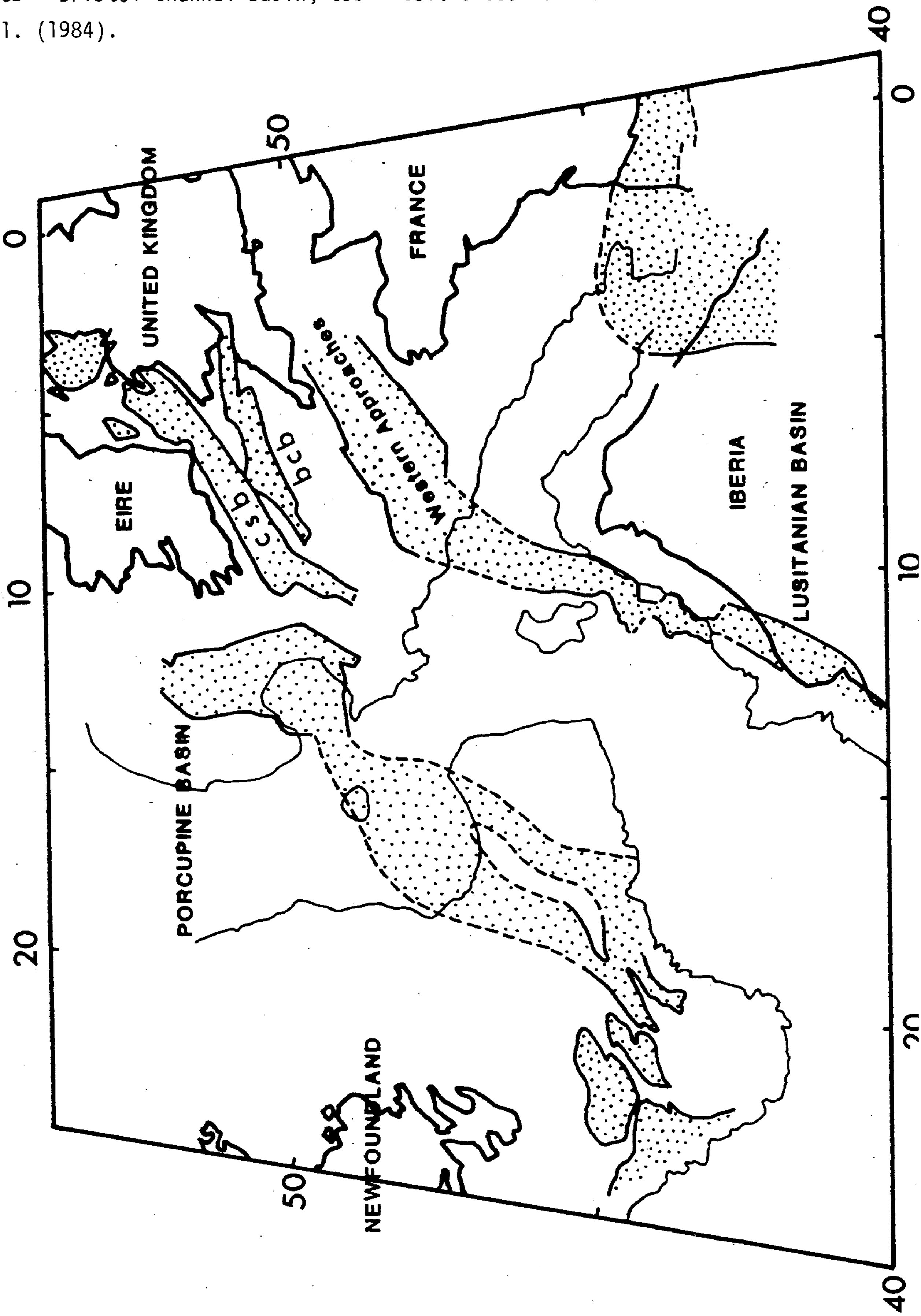
The Consolidation of Pangaea

Following the waning, in the Early Devonian, of the forces that resulted in the Caledonian orogen, structural developments in the British Isles were influenced by the interplay of two important tectonic systems. During the Devonian most of Central and Northwest Europe was governed by crustal extension which gave rise to the continental molasse basins of the UK. In addition, palaeomagnetic data (Kent and Opdyke 1979) indicate that to the NW of the British Isles during the Devonian and Early Carboniferous a sinistral transcurrent system existed between the Laurentian-Greenland and Fennoscandian-Russian shields; the translation along this system was of the order of 2000 km and, while it probably had a negligible effect over most of Europe, it is believed to have been responsible both for syndepositional deformation in the UK Devonian rift basins and for the development of the Great Glen Fault. A left-lateral displacement of up to 300 km from Late Devonian to Early Carboniferous times has been proposed for the Great Glen Fault. This Arctic-North Atlantic wrench system is also likely to have initiated or enhanced the roughly SW-NE lines of weakness which later were to define the shape and extent of Rockall Trough and Plateau (see Ziegler 1981, Fig. 5). The palaeomagnetic studies of Torsvik et al. (1985) may throw some doubt on the existence of this broad transcurrent zone. At the end of the Early Carboniferous the wrench system became inactive and was replaced instead, for the remainder of the Carboniferous, by a north-south trending rift system (e.g. Haszeldine 1984).

In contrast, the Devonian to Carboniferous transition in Europe records the inexorable convergence of the Gondwana and Laurasian plates, the onset of the Variscan orogeny and the welding of the Pangaean supercontinent. In the early stages this collision was directed north-south but by Stephanian times (c.290 My) convergence between the two plates underwent a reorientation to roughly east-west; this was accompanied by the development of a large dextral transcurrent system which connected the Southern Uralides and northern Appalachians (Arthaud and Matte 1975, 1977). Where it traversed Europe this system gave rise to a complex pattern of shear faults and pull-apart structures (Ziegler 1981). To what extent the Variscan orogeny, whose main pulse was during the Late Visean (Sudetic phase), and the Late Carboniferous dextral wrench system influenced the region around Rockall is largely unknown. It appears that for the whole of the Carboniferous much of the Scottish and Irish Atlantic seaboard was elevated and the existence of Rockall Trough and the Western Approaches basins is difficult to verify. Haszeldine (1984) cites drilling evidence for the presence of Visean and thick Namurian and Lower Westphalian sediments in a half-graben on the Rockall Trough margin north-west of Donegal. However, the absence of Palaeozoic sediments younger than Westphalian B from the Rockall-Faeroe Channel does not favour the idea of a long, persistent subsiding rift here at this time.

Haszeldine (1984) incorporated the dating of later Carboniferous dyke swarms and volcanics with the stratigraphic information and proposed initial oceanic accretion in the Rockall-Faeroe Channel as early as Westphalian C/D times, expanding to a maximum width of 220 km by the Early Permian. This is in agreement with the formation of a Rockall ocean (sensu stricto) from lower Permian times as propounded by Russell (1976) and Russell and Smythe (1978). Such a tectonic scenario involves the opening of the Trough deep within a continental interior with transcurrent faults linking the ends of the rift to the Arctic and Tethyan realms, to the north and south

Figure 1.9 Late Triassic - Early Jurassic rift basins across the pre-drift reconstruction of the N. Atlantic. Dashed lines mark inferred boundaries. bcb = Bristol Channel Basin; csb = Celtic Sea Basin. Redrawn from Masson et al. (1984).



respectively (Bott 1978; Haszeldine 1984, Figs 4 and 5). Differential subsidence of the Porcupine Seabight and Western Approaches basins commencing in the Late Carboniferous is documented in the Stephanian and Permian sediments that have been drilled in them both and this subsidence may well be related to movements on the Biscay dextral wrench system.

The Pangaean supercontinent was consolidated by the Early Permian but towards the end of that period it had become unstable, and this instability increased through the Triassic to give rise to the nascent rifts which would later govern its progressive fragmentation and the opening of the North Atlantic. The subsequent Mesozoic and Cenozoic evolution of the North Atlantic is complicated and has been covered in greater detail elsewhere (Roberts 1974, Le Pichon 1977, Svirastava 1978; Kristoffersen 1978; Bott 1978; Ziegler 1981; and Naylor and Shannon 1982).

The Opening of the North Atlantic Ocean

Comparisons with the Permo-Triassic basins in Scotland and East Greenland suggest the presence of extensional intracratonic basins in the Rockall-Faeroe Channel from the Saxonian (260 My B.P.). The Zechstein Seas may have extended into southern parts of the Rockall rift. Intensified rifting occurred in the Late Triassic to Middle Jurassic (the Early and Mid-Cimmerian Phases) and this led to the formation of a coherent north-easterly trending rift system (Fig. 1.9) whose major elements can be traced across the European, Iberian and North American continental margins (Masson, Parson and Miles 1984). Although Northwest Europe was the focus of extensive Palaeozoic suturing and subsequent rifting and fracturing, Figure 1.9 shows that these older trends were clearly crossed in many instances by the expanding N. Atlantic.

This large system of rifts, which probably includes the Rockall-Faeroe Channel, and which is largely absent of volcanism, is believed to have resulted from regional crustal extension that also caused the rifting between Africa and N. America commencing in the Late Triassic. During this period marine conditions may have prevailed in

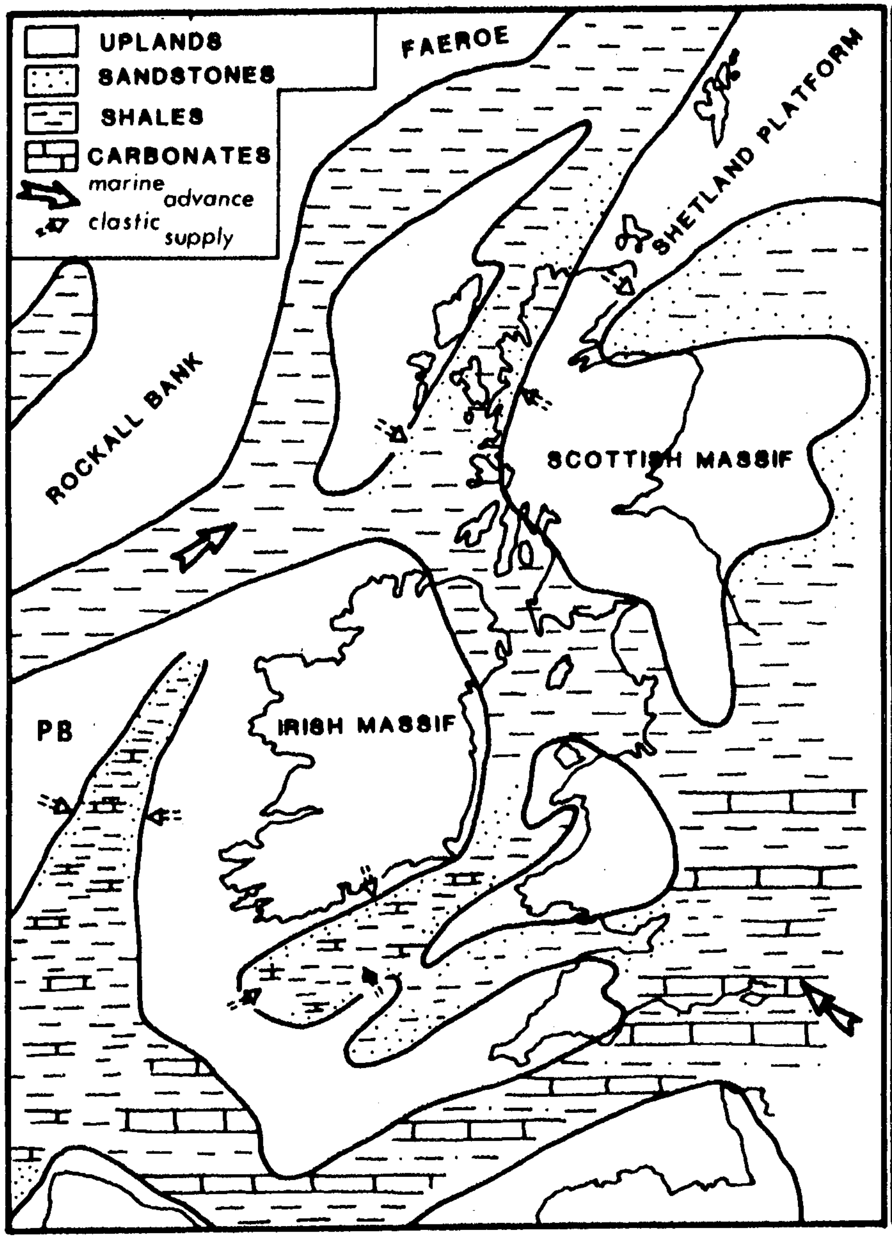


Figure 1.10 Upper Jurassic palaeogeography around the British Isles. After Naylor and Shannon (1982).

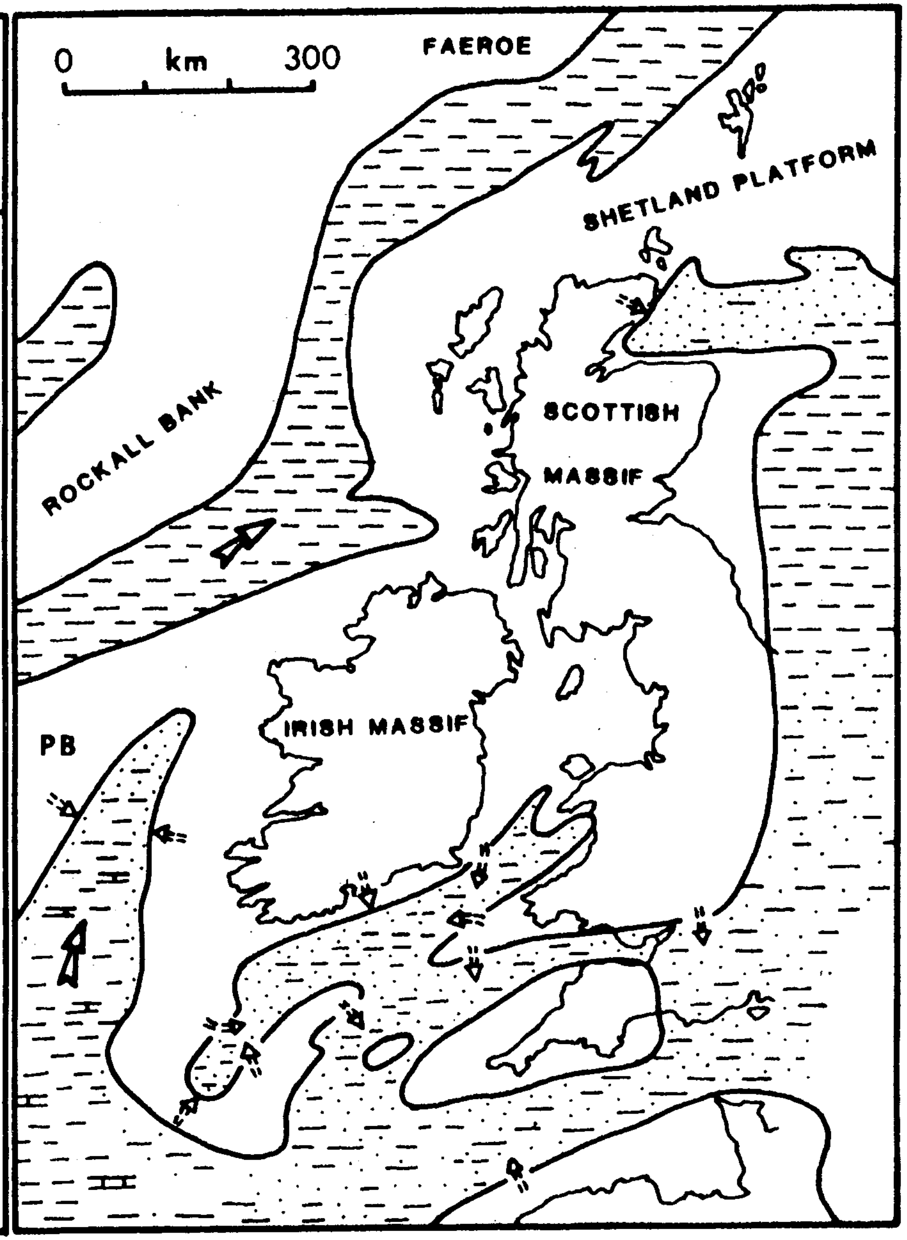


Figure 1.11 Lower Cretaceous palaeogeographyaround the British Isles. After Naylor and Shannon (1982).

the Rockall Trough depositing predominantly shales and sandy-shales. The Trough may have been connected to the basins in the Western Approaches via Porcupine Seabight or, more likely, the West Scottish and Irish Seas (Fig. 1.10; Naylor and Shannon 1982).

The Late Cimmerian major rifting event of the end Jurassic and earliest Cretaceous preceded the Late Neocomian onset of sea floor spreading in the North Atlantic between the Azores-Gibraltar Ridge and the Charlie-Gibbs Fracture Zone. A major regional marine regression associated with this event left most of the British Isles emergent. The Rockall Trough and Porcupine Basin were centres of marine shale and sandstone deposition (Fig. 1.11) and, like most of the Jurassic, are thought to be the setting for little volcanic activity during the Early Cretaceous.

Geophysical information, particularly magnetic data, indicates that sea floor spreading between Iberia-Europe and North America was not synchronous owing to the northerly progress of the tip of the Mid-Atlantic spreading ridge (Masson and Miles 1984). A corollary of this is that the ocean-continent boundary is not an isochron. Thus the opening between S. Iberia and Newfoundland commenced in the Barremian-Hauterivian (c.120 m.y.B.P.) and reached Galicia Bank and into the Bay of Biscay by the Late Aptian (108 m.y.B.P.). Further north still the oldest oceanic crust west of Goban Spur has been proven by drilling to be late Early to Middle Albian (c.100 m.y.B.P.) in age (Masson and Miles 1984; Masson et al. 1985). From a consideration of the relevant magnetic data and the geometry of continental reconstructions obtained using them, together with seismic reflection and refraction work, there is a sizeable body of opinion which argues that this oceanic crust continues north into Rockall Trough and eventually into the Faeroe-Shetland Channel (Jones 1978; Bott et al. 1979; Roberts et al. 1981; Hanisch 1984; and Price and Rattey 1984). The age of the oldest oceanic crust would be approximately Late Albian or Cenomanian by extrapolation from west of Goban Spur. Unfortunately there is no drilling information to substantiate these claims.

The gradual build up of the Mid Atlantic ridges and the acceleration of global sea floor spreading rates were almost certainly responsible for the rise in sea level during the Middle and Late Cretaceous - a global transgression which resulted in the deposition of the Upper Cretaceous Chalk over most of the Northwest

European continental shelf. To the north-west the deeper Rockall, Faeroe and Porcupine Troughs were dominated by shale, rather than chalk, sedimentation. Any sea floor spreading that may have been proceeding in the Rockall Trough ceased when the ridge system extended into the Labrador Sea during the Santonian (c.85 m.y.B.P.), thus separating Greenland-Europe from N. America (Fig. 1.12). The formation of the three seamounts in northern Rockall Trough has been ascribed to the last pulses of sea floor spreading in the Late Cretaceous (Jones et al. 1974; Miles and Roberts 1981).

At the end of the Cretaceous the first major pulse of the Alpine orogeny (Laramide phase) led to large scale regional uplift and associated local downwarping across the British Isles, and the retreat of the Chalk Sea. Large tracts of the continental shelf west of Ireland and around Rockall Plateau and Porcupine Bank were exposed as a result. This Palaeocene period of updoming and extension, the last tensional episode to affect the North Sea and Atlantic shelves, was accompanied by widespread and often intense extrusive and intrusive activity mainly of a basaltic and gabbroic character. This formed the Thulean igneous province of NW Scotland, Ireland, Rockall, Faeroe and E. Greenland.

The onset of oceanic accretion between Rockall-Faeroe-UK and Greenland at the time of magnetic anomaly 24-25 (c.55 m.y.B.P.) marked the release of the stress responsible for extension and therefore the end of rifting along the Scottish-Irish seaboard. Thulean volcanic activity persisted up to the end of the Eocene epoch (Fig. 1.13). Sea floor spreading in the Greenland-Norwegian Sea also led to regional subsidence of Hatton-Rockall Basin, Rockall Trough, Faeroe-Shetland Channel and Porcupine Seabight, especially in Eocene times. This subsidence is still continuing today (Naylor and Shannon 1982).

Large areas of the continental shelf which were emergent during much of the Tertiary and which caused the shelves around Rockall Trough to be largely by-passed by clastic sedimentation became submerged again in the Quaternary.

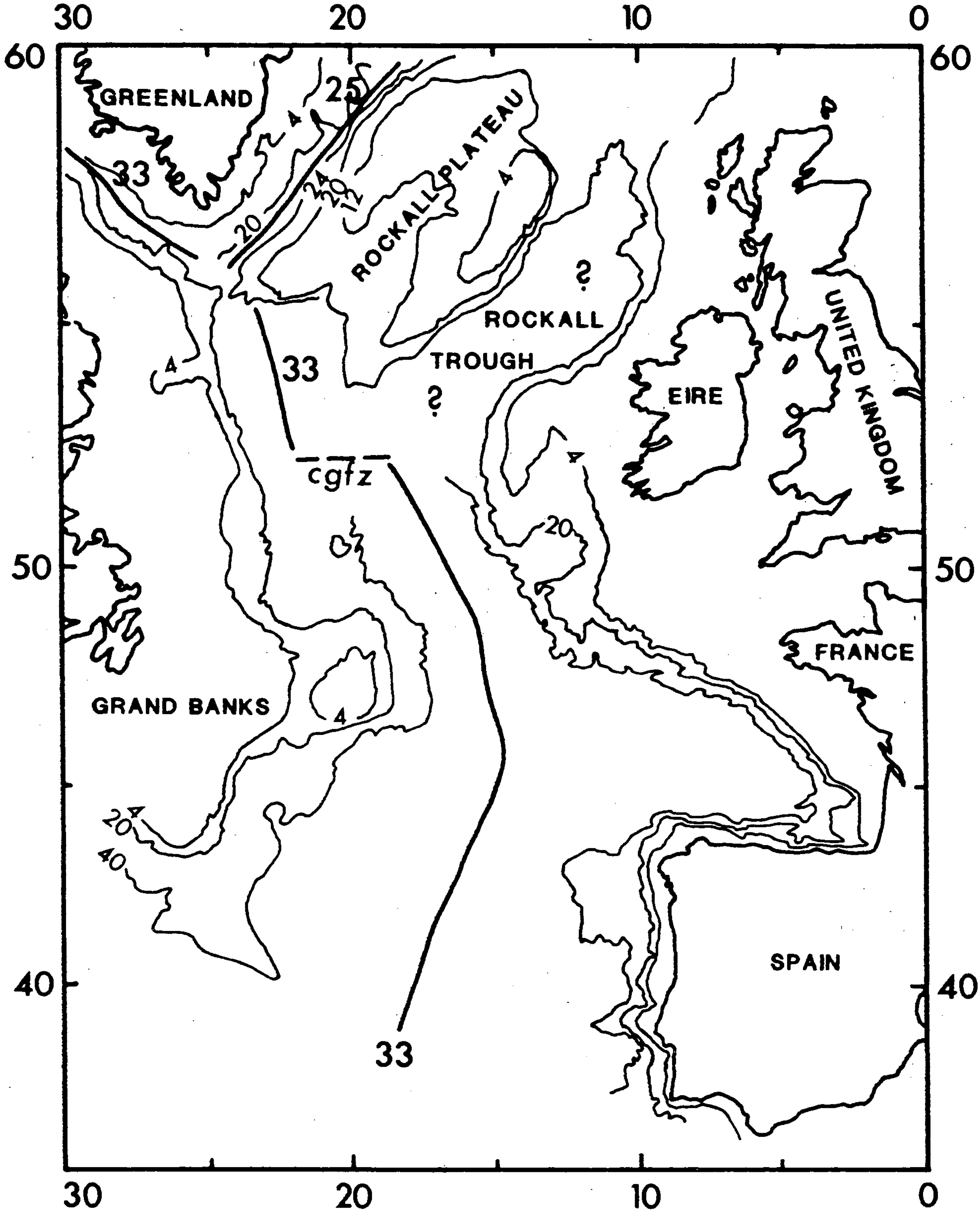


Figure 1.12 Anomaly 33 (ca. 70 m.y.B.P.) reconstruction of the North Atlantic continents, after Kristoffersen (1978). CGFZ = Charlie-Gibbs Fracture Zone. Bathymetry in hundreds of metres.

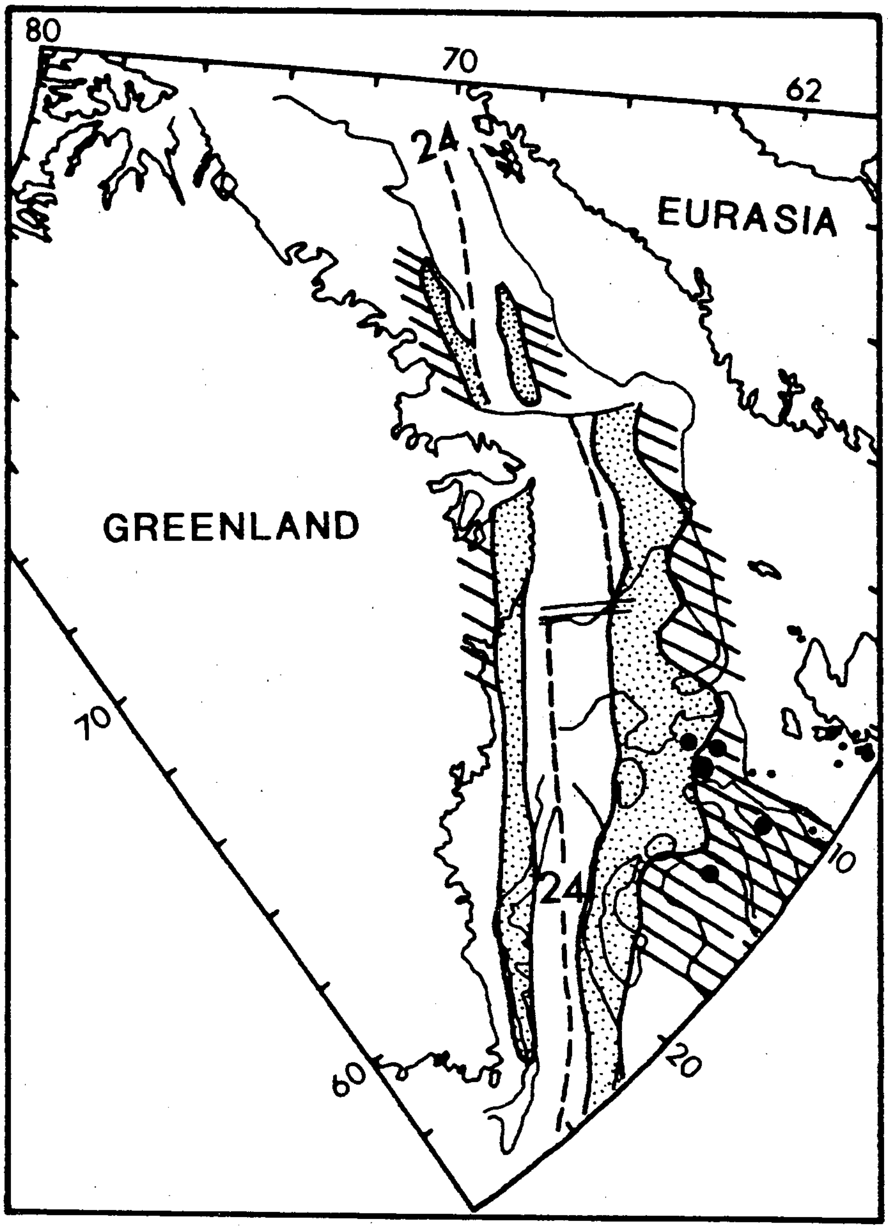


Figure 1.13 Anomaly 24 (ca.52 m.y.B.P., dashed line) reconstruction of the Norwegian-Greenland Sea showing the extent of dipping reflectors (stipple) at the continental margins, of Palaeocene - E.Eocene lavas and sills (diagonal ornament), and of mostly early Tertiary igneous centres in NW Britain and Rockall Trough (black circles). Redrawn from Roberts et al. (Init. Repts. DSDP, V81, pp 913-23; 1985)

1.5 Layout of the Thesis

Earlier sections of this chapter were concerned with the geological evolution of Rockall Trough and its adjacent shelves in the broad setting of Laurasia, its fragmentation, and the consequent growth of the North Atlantic Ocean. Chapter 2 provides an account of the acquisition and processing of the underway geophysical data which form the basis of this research.

Although Chapter 1 documents the development of Rockall Trough in a simple chronological fashion it was not considered practical to extend this approach to the main body of this study. Instead an important geological and tectonic (and in part bathymetric) distinction was made between the deep Rockall Trough and the extensive oceanic domain to the south, the two areas being juxtaposed across the Charlie-Gibbs Fracture Zone. Thus in Chapter 4 attention is given over to the nature and structure of this important discontinuity and part of the oceanic crust to the immediate south, as determined from the geophysical information. Rockall Trough proper is considered in a similar manner, though being the principal concern of this study receives more detailed treatment. Chapter 5 deals with the available seismic reflection profiles across the Trough, Chapter 6 the gravity anomaly data, and Chapter 7 the total intensity magnetic anomaly data. Chapters 4 through to 7 then form the core of this research.

In an attempt to place these two geological provinces into some sort of time framework a fairly comprehensive review of the seismic stratigraphy in the Rockall area is presented in Chapter 3. A modification and extension of the existing stratigraphies enables the components of the study area - the Trough, its sediments, the continental margins, banks and shelves, the Gibbs Fracture Zone, and the oceanic realm - to be placed in a general chronological scheme. In Chapter 3 the Early, Middle and Late terminology adopted by Harland et al. (1982) is used wherever possible for accuracy and consistency, in preference to the looser Lower, Middle and Upper prefixes. A discussion and summary of the main points highlighted by this work are provided in Chapter 8, together with a brief prognosis of further investigation and possible useful target areas.

2. DATA ACQUISITION, REDUCTION AND INTERPRETATION

2.1 Geophysical surveying at sea

This research is based largely on the results obtained through underway geophysical remote sensing involving three standard techniques: measurements of the Earth's gravity and magnetic fields, and continuous seismic reflection profiling.

Nowadays gravity observations are made at sea using a gravity meter of the unstable spring balance type which is housed on a gyroscopic platform that counteracts the variable motions of the ship (Telford et al. 1976). Present marine gravimeters can measure changes in the gravitational acceleration to a precision of 0.1 to 1 mgal and naturally must be tied into absolute determinations of gravity near the start and end points of any cruise in order to calculate the absolute gravity values along the ship's course. Corrections for drift in the meter during a cruise, possibly induced by changes in temperature, may be necessary.

In marine geological applications the free air gravity anomaly is the one most often used, unlike the Bouguer anomaly for land based studies. The free-air gravity anomaly is obtained by subtracting the theoretical from the observed gravity values, where the theoretical value, calculated from the 1967 International Gravity Formula, has been corrected for the latitude of the ship, and its velocity and heading (the Eötvös correction) at each point of measurement. Since the gravity measurements are all made at sea level no elevation correction is required (the International Formula being referenced to sea level datum). The free-air anomaly, then, reflects changes in water depth and density variations in the subjacent geology.

The Earth's magnetic field is recorded at sea using a towed nuclear free precession (or simply proton) magnetometer (Telford et al. 1976). The instrument is towed astern the ship at a distance of approximately 300 m to remove it from the spurious magnetic fields inherent in the vessel. The proton magnetometer measures the magnitude of the total ambient field but not its direction; it can not measure, separately, the horizontal or vertical components of the magnetic field. To its advantage, however, the proton magnetometer requires no levelling (unlike the gravimeter) and hence is well suited to marine applications. The high geographical (and magnetic) latitudes in which this research area is located (51°N to 56°N) means that the measured total field will resemble fairly closely its vertical component (present magnetic inclination = 70° in Rockall area).

The precision with which the proton magnetometer can measure the magnetic field at sea is of the order of a few nanoteslas (nT). But the accuracy to which it is possible to record those disturbances in the field that are due to variations in the magnetic properties of the rocks, namely their magnetic susceptibility, is limited by diurnal oscillations and ship's heading errors. The amplitude of the short period magnetic noise is about 25 nT and it probably largely accounts for the cross-over errors (0-45 nT) within the three Edinburgh cruises around Rockall Trough (Sections 2.2 and 2.3 below). The larger cross-over errors between these three cruises, frequently more than 50 nT and as much as 180 nT on occasions, are due to the longer period (secular) variations in the magnetic field. The implications of these sources of error are discussed briefly in Section 2.4. No corrections for the variations mentioned above were made to the magnetic anomaly data.

Unfortunately the treatment of the principles of magnetism in the literature is confusing both in its algebraic nomenclature and its use of units (for example, Dobrin 1976; Nettleton 1976; Telford et al. 1976; Parasnis 1979; and Dohr 1981). This author has chosen to use the rationalised International System of Units (SI) whenever possible. Hence magnetic field strength or intensity, H, is now measured in amperes per metre (strictly ampere-turns per metre); in the earlier c.g.s. electromagnetic system it was measured in oersted, 1 A/m being equal to 10^{-3} Oe. Magnetic induction or flux density, B, has the unit tesla (T) in the new MKS system, replacing the gauss and gamma units of the c.g.s. system. 1 nT = 1 gamma = 10^{-5} oersted.

The continuous seismic reflection profiling method is now the most important tool in the search for hydrocarbons with the result that the systems available for employing it have become sophisticated and expensive. The major advances in the fields of seismic digital recording and processing have arisen from developments in digital computers. Although these digital systems have been utilised in the commercial sphere for many years, they have only recently been installed on ships for academic operations.

All the seismic reflection data used in this work were collected using one or more compressed air guns as the acoustic source and a hydrophone streamer as the detector. A more thorough treatment of marine seismic reflection techniques can be found in McQuillin and Ardus (1977), McQuillin et al. (1984) and Sheriff and Geldart (1982). The depth of water and thickness of sediments in Rockall Trough place constraints on the combinations of equipment that can be employed. Since the objective of most of the surveys has been to record reflections from geologic basement and the major sedimentary events in the Trough, deep seismic penetration is sought at the expense of good resolution. Shallow penetration, high resolution sources such as sparkers and boomers do not fulfil this requirement. Unfortunately the slower firing rate of the larger air guns (up to 1000 cubic inches), which achieve adequate penetration, causes the seismic profiles widely used in this work to be highly compressed in the horizontal sense - this makes their interpretation difficult at times. The vertical exaggeration on the available multichannel profiles is usually a good deal lower than that above, about 3 times compared with 6 times, and consequently are truer to life and easier to interpret.

All of the seismic reflection profiles shot here by NERC ships (c.6100 km), including the Edinburgh data (sections 2.2 and 2.3), were obtained with airguns on double-channel analogue systems. The remainder of the reflection data, a little over 2500 km, consist of multichannel seismic profiles, some of them reprocessed (see Chart 1).

2.2 The R.R.S. Challenger 1/84 Cruise

Cruise logistics and track policy

A scientific cruise on board R.R.S. Challenger (Fig. 2.1) for the purpose of collecting underway geophysical data in Rockall Tough was funded by NERC from 6 to 21 June 1984. The cruise was originally scheduled 19 January to February 1984 but was postponed to enable replacement of the main engine boiler. One day of the cruise (20

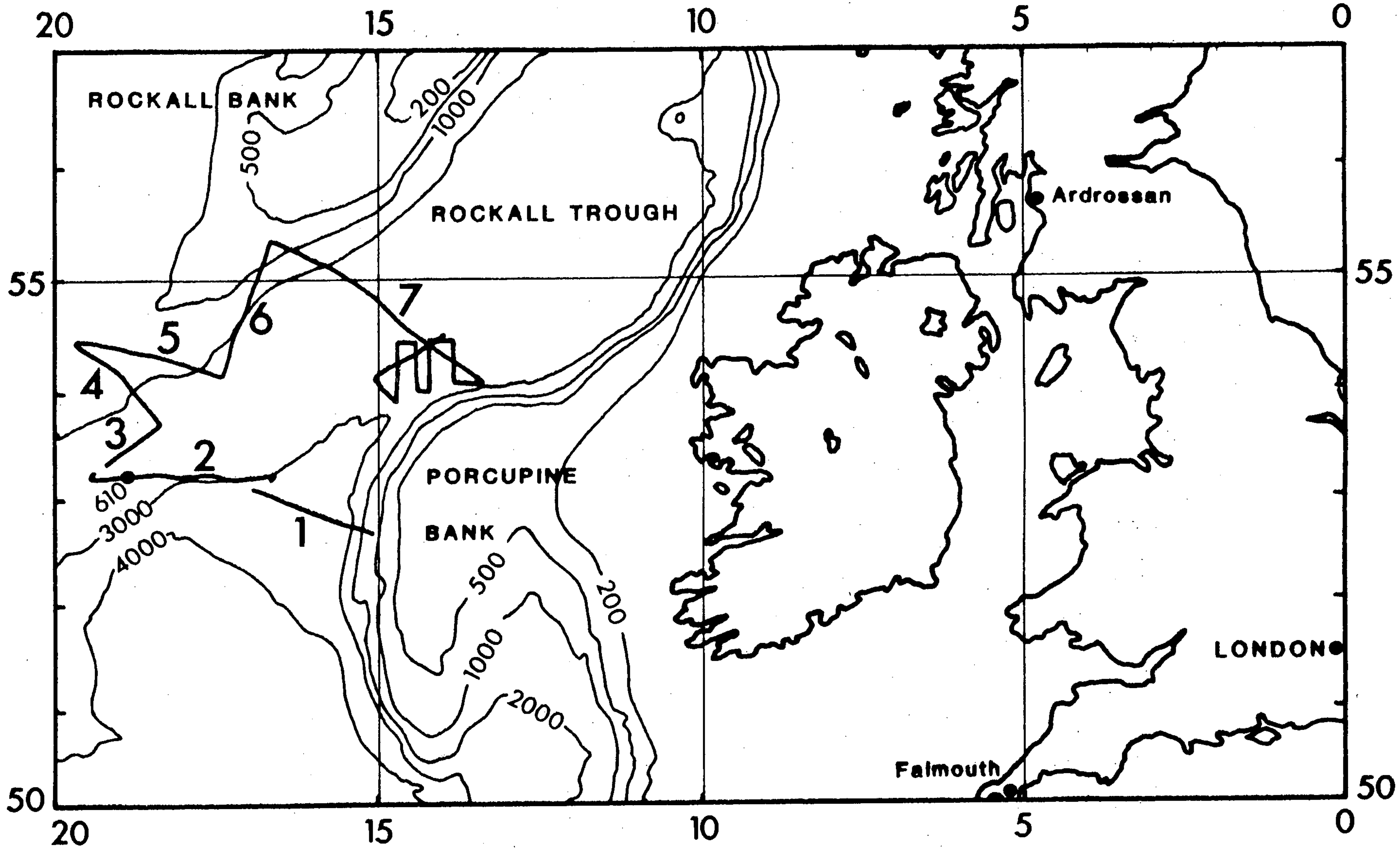


Figure 2.1 Seven main tracks and box survey recorded during the Challenger 1-84 cruise in Rockall Trough. Bathymetry in metres. Solid circle marks DSDP site 610.

June) was set aside for a test deployment of the Pop-Up Multichannel Array (PUMA) developed by a scientific team at Cambridge University. Mechanical problems in the engine room, which originally caused the ship to be laid up in Falmouth, forced a three day delay in sailing. R.R.S. Challenger departed Falmouth at 0000 hrs (BST) on Saturday 9 June and arrived in Ardrossan at 0800 (BST) on Thursday 21 June. During this interval science was accomplished continuously on the eight days 11 to 18 June with the exception of two short periods when bad weather caused engine problems and power failure; these occurred at 0400-0700, 12 June and 1030-1200, 13 June (see Chart 1 and Fig. 2.1). Challenger was on passage to and from Rockall Trough on the 9, 10, 19 and 21 of June.

The course of survey lines finally selected was based on the interpretation of all the available data at the time, i.e. the remaining tracks on Chart 1, and represents an attempt to reduce gaps in the existing data and to extend our knowledge of the peripheral areas.

Lines 1 and 2 were sited to test whether the strong relief observed on the seismic basement on Shackleton 3/79 Line 14 and Challenger 6/80 Line 1 (Chart 1) was continuous between them. Line 2 was made to pass over site 610 of DSDP Leg 94 (IR DSDP V94, report in prep.) in an attempt to establish a seismic stratigraphic link with site 550 of DSDP Leg 80 (Fig. 1.5; Graciansky, Poag et al. 1985). This work was done in collaboration with D.G. Masson at the Institute of Oceanographic Sciences, Wormley (see Chapter 3;also Masson and Kidd, in prep.).

Line 3 runs over a 400 nT positive magnetic anomaly (Fig. 2.1) and combined with Challenger 6/80 line 2 (Chart 1) provides a useful 250 km long magnetic and gravity profile parallel to the north-west margin of the Trough. Lines 4, 5 and 6 were successful in extending gravity coverage from the Trough and up the continental slope; line 4 was particularly useful in completing a margin to margin seismic, gravity and magnetic profile with Challenger 6/80 Line 1.

Line 7 was also designed to provide a margin to margin gravity profile for modelling purposes, but lack of time meant that the south-eastern end of the line was about 30 km too short. Inspection of seismic profile CM-03 (Chart 1) revealed that a major discontinuity was present in the basement between 14°W and 15°W below north Porcupine Bank. The grid survey (Figs 2.1 and 2.2) consisting of lines 8 to 17 was followed in an effort to map out in more detail the extent of the discontinuity and the distribution of the basement types. Altogether approximately 1420 km of underway geophysical data were collected during the cruise.

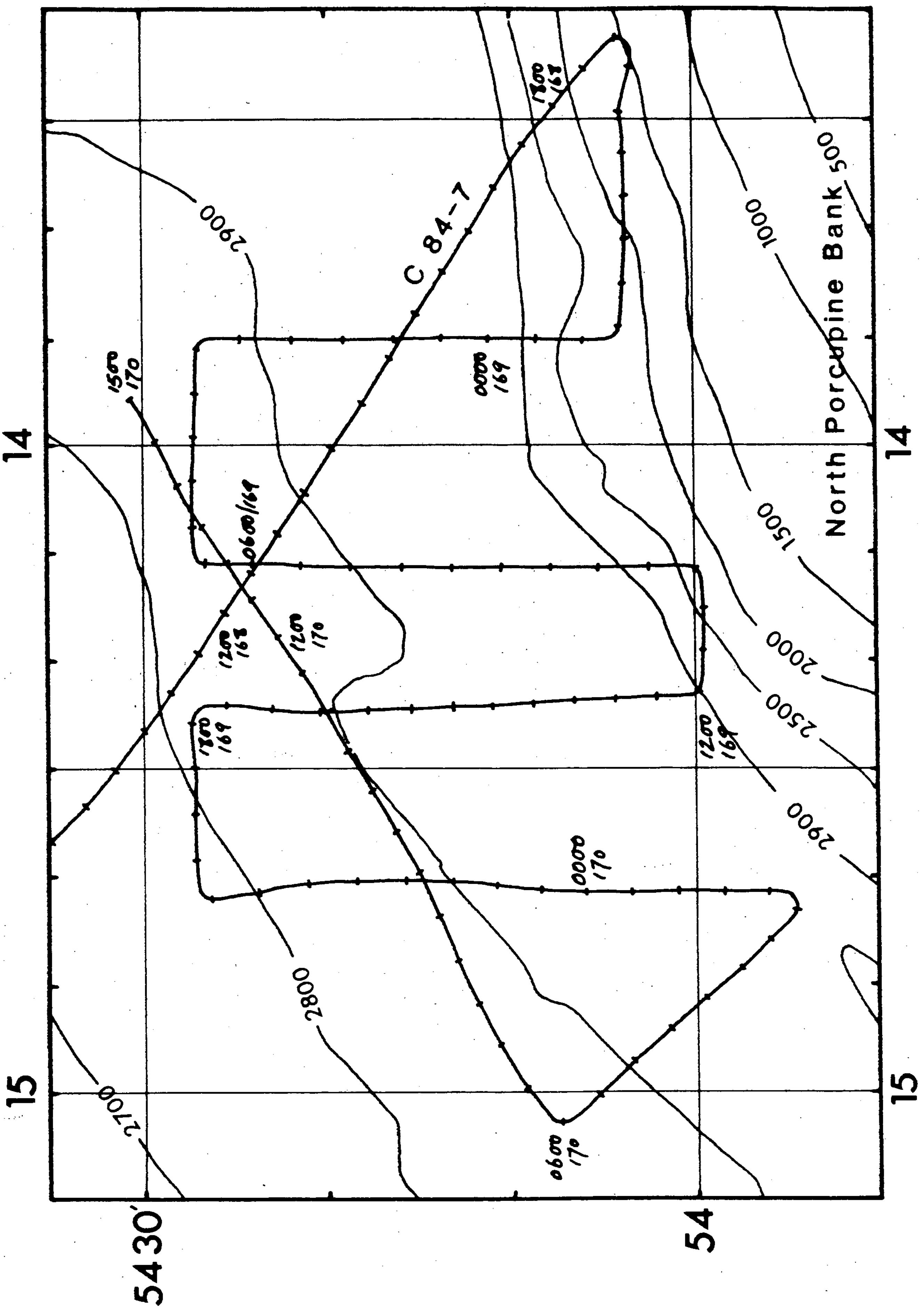
Data collection and reduction

Navigation on the cruise was provided by two systems. A Magnavox 1107 satellite navigator received signals from the six U.S. Government Transit Satellites which complete polar orbits every 107 minutes. Good position fixes were obtained every 1 to 3 hours and the system has an accuracy of about 0.5-1 km. This was supported by position fixes from the Loran C system (Long Range Aid to Navigation). It provides an accuracy of 100 to 200 m depending on the distance of the ship from the transmitting station.

Bathymetry was measured continuously using a shallow-towed Precision Echo Sounder (PES). Total field magnetic values were recorded using a Barringer proton precession magnetometer towed roughly 300-500 m astern the ship. Relative gravity measurements were made with a LaCoste and Romberg unstable-type gravimeter, housed near the centre of gravity of the ship to minimise unwanted accelerations at sea. Gravity tie-ins were performed to base stations on Falmouth docks (County Wharf, station number 009.28) and on Ardrossan docks. The drift in the gravimeter during the cruise was only 2 to 3 mgals, so no corrections to the processing of the gravity data were thought necessary.

The acoustic signal for the seismic reflection profiles was provided by Bolt PAR airguns operating at 1700-1800 p.s.i.. Almost all of the c.1420 km of seismic lines were shot using a 160 cubic inch airgun firing at a rate of one shock every 24 seconds. To achieve good resolution of the sedimentary reflectors along Line 2 (Chart 1) and over DSDP Leg 94 site 610 (Fig. 1.5), for the purpose of seismic correlation, it was hoped to be able to deploy a tuned array of two 40 cubic inch airguns firing every 16 seconds. Although an excellent acoustic signal was obtained for a short period after deployment, leakage in one of the airguns forced a return to 160 cu. in. capacity at the expense of the required resolution.

Figure 2.2 (opposite). Navigation chart for the Challenger 1-84 grid survey over north Porcupine Bank - lines C 84-8 to C 84-17. Bathymetry in metres. Time marks every 30 minutes. See Figure 2.1 for regional setting.



54 30'
54
14
15
C 84-7
North Porcupine Bank
500
1000
1500
2000
2500
2900
2800
2700
1500 170
0000 169
1800 168
0600/169
1200 168
1200 170
1800 169
1200 169
0000 170
0600 170

Seismic reflections were detected by a two-channel GEOMECANIQUE hydrophone streamer containing a 30 m active section consisting of one group of 32 hydrophones, and another group of 16 hydrophones. Signals from the hydrophone streamer were recorded on a RACAL store 4D tape recorder and played back through a variety of filters and gain devices onto three EPC continuous roll graphic recorders. EPC recorders 1 and 3 both displayed an 8 second sweep with filter passbands of 20-200 Hz and 10-50 Hz, respectively. EPC recorder 2 was given a 4 second sweep and a filter setting of 40 to 150 Hz. In each case delays were applied to the time interval that was displayed, and time-varying gain (TVG) to the tape output signal, as necessitated by changes in the water depth.

The raw along-track bathymetric, magnetic and gravity data were recorded in analogue form by electrical stylus (PES) or pen and chart. Magnavox satellite fixes were printed onto rolls. Every second the day and time, ship's speed and heading, and bathymetric, magnetic and gravity values were recorded by DIGITAL data logger. A manual record of this raw data was maintained at 10 minute intervals.

A shipborne DIGITAL PDP 11/34 mini-computer was used to process sizeable portions of the geophysical data as they became available. Navigation information from dead reckoning, Loran C and satellite fixes was combined to produce a final track chart at 1:500,000 which provides a base map for plotting along-track observations. The bathymetric values were corrected according to Matthews' areas (variation of sound velocity in sea water) and total field magnetic anomalies were calculated by subtracting the 1980 International Geomagnetic Reference Field (IGRF) from the observed magnetic values. Absolute gravity values were obtained by adding the relative measurements to the absolute gravity determined at Falmouth docks, and the free-air anomaly was found by applying latitude and Eötvös corrections and subtracting the theoretical gravity field, according to the 1967 International Formula.

Bathymetry, magnetic anomaly and free-air anomaly values were plotted along-track every 12 minutes at 1:500,000 using a CALCOMP roll graph plotter. The grid survey comprising lines 8 to 17 (Chart 1) was plotted at a scale of 1:250,000 with the relevant values every 6 minutes (see section 2.4 below).

2.3 Previous geophysical surveys

Geophysical information from a number of other sources has been made available to the author, without which it would not have been possible to undertake this research. Each of the surveys and their constituent lines are illustrated on Chart 1.

Bathymetry, gravity, magnetic and two-fold seismic reflection data have been used frequently from two scientific cruises carried out by the Grant Institute of Geology, Edinburgh University and funded by NERC. R.R.S. Shackleton cruise 3/1979 surveyed over the southern Porcupine Bank continental margin and the abyssal zone to the west around the Charlie-Gibbs Fracture Zone. R.R.S. Challenger cruise 6/1980 observed 21 lines totalling approximately 2320 km in southern Rockall Trough and across the continental margin on north and west Porcupine Bank; the data from this cruise were processed by the author at the NERC Research Vessel Services base, Barry, S. Wales in June 1983.

The Institute of Oceanographic Sciences provided copies of two-channel seismic reflection and magnetic anomaly profiles obtained on R.R.S. Discovery cruises 60/1974 and 84/1977 (see IOS annual reports for those years) in parts of the Rockall Trough. In addition the author was permitted to inspect and make tracings of several useful multichannel seismic reflection profiles: lines CM-03, CM-04, GSI-1, NA-1 and NA-1 Ext, RH111, RH112, RH113, and WI-32. The CM survey also provided the total intensity magnetic anomaly every 2.5 km and free-air gravity anomaly every 5 km.

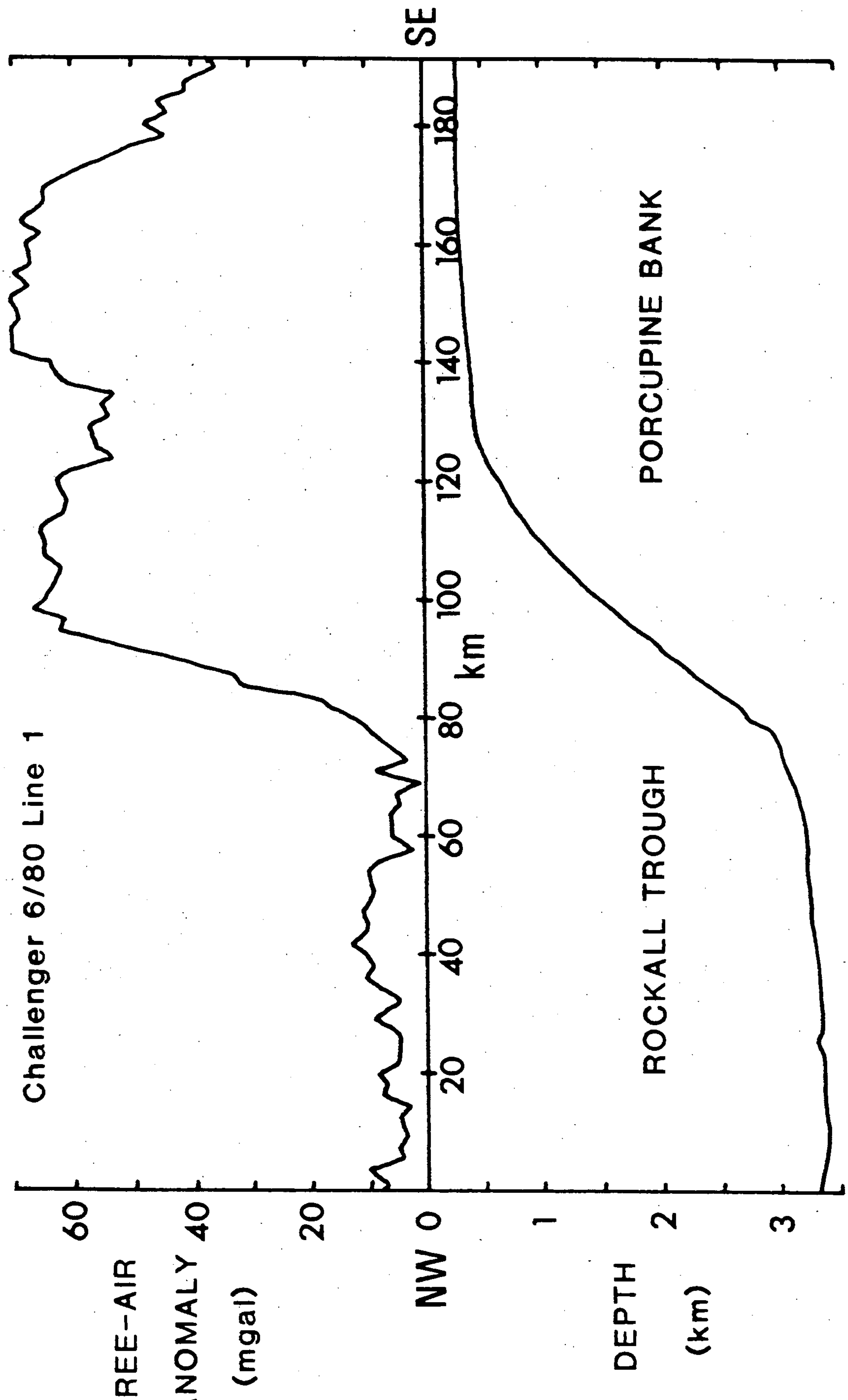
The 48-fold seismic reflection profiles WI-1A, WI-3, WI-8, WI-10 and WI-26 (shot in 1975) over north Porcupine Bank, and extending down the continental slope, were supplied by the Irish Department of Industry and Energy, Dublin. Single-channel reflection profiles, and free-air and total field magnetic anomaly values collected by Jean Charcot in Rockall Trough in 1969 have also been used in this work (Scientific Group 1971).

2.4 Interpretation and modelling of gravity and magnetic anomalies

A free-air gravity anomaly chart (Chart 2) has been produced by hand contouring all the available along-track gravity data which were combined on a 1:500,000 base map. The values from the Shackleton and two Challenger cruises were posted mainly every 12 minutes (c.2-3 km). For the grid survey of the Challenger 1/84 cruise the values were posted every 6 minutes, thus permitting a more accurate contouring in the area off north Porcupine Bank. Free-air anomaly values were posted every 5 km along the CM survey lines and a constant value of 10 mgal was added to each of them to improve tie-ins with the other lines. A similar value was used independently by D.G. Masson (pers. comm.) who supplied the CM gravity data. Cross-over errors both within cruises and between cruises are mostly 5 mgal or less. Indeed within cruises the error is generally only 2 or 3 mgal. The contoured chart was photographically reduced to a scale of 1:1,000,000 to enable direct comparisons with the other charts at the same scale. The free-air anomaly chart is useful in providing a qualitative assessment of the regional trends and gradients in the gravity field, qualities that reflect upon the thickness and density variations within the subjacent crust and upper mantle. Many of the features visible on Chart 2 however are due to changes in bathymetry, most notably the free-air edge effect associated with the sudden increase in water depth at the continental slope (Fig. 2.3).

A more rigorous quantitative approach is possible through forward modelling in two or three dimensions, that is, establishing a density model and calculating the resultant gravity anomaly. Two-dimensional modelling was performed for a number of profiles in the study area where the free-air anomaly chart suggested that such a simplification was valid. This modelling has only been done along actual tracks where the gravity field was observed, and not along profiles interpolated from the free-air contour chart. Because the largest density contrast is between sea water (1.03 g/cm^3) and the rocks constituting the sea floor (upwards of 2.0 g/cm^3), a large contribution to the gravity anomaly is going to derive from variations in the bathymetry. For this reason the depth to sea floor in the gravity models was obtained from the along-track Precision Echo Sounder readings rather than published bathymetric contour charts.

Figure 2.3 Part of Challenger 80-1 bathymetry and free-air anomaly profiles across W Porcupine Bank showing relationship between steep continental slope and 60 mgal free-air edge effect. See Chart 1 for track location.



Owing to the scarcity of good quality seismic refraction results in Rockall Trough it is not possible to constrain the depths of the main crustal and density divisions with precision. The density values assigned to the various sedimentary and crustal units identified from the seismic reflection and refraction profiles are obtained either by comparison with geologically analogous regions where densities have been directly determined, or by using the Nafe-Drake relationship between compressional wave velocity and bulk density (Nafe and Drake 1963; Fig. 2.4).

Although forward modelling in two dimensions is easy to perform, especially with today's digital mainframe computers and their graphic viewing facilities, the final result or model should be interpreted with caution. No particular combination of prism geometries and densities can be considered unique for a specified gravity profile; adjustments can be made to depths and density contrasts which may equally well satisfy the same profile. This difficulty of non-uniqueness can largely be avoided by adhering to sensible geological situations. Where it is clear that the simplified two-dimensional case does not exist, end corrections can be applied (Nettleton 1976) to take account of the finite extent of the prisms in the two-dimensional model. Alternatively it is possible to resort to modelling in three dimensions. This technique is less precise and more time-consuming (particularly on the computer) than the 2D method, but it tends to be more meaningful in geological terms. A computer program for modelling gravity anomalies in three dimensions based on horizontal slabs, rather than prisms, has been used to test ideas about an important gravity anomaly in southern Rockall Trough (see Chapter 6).

A total field magnetic anomaly chart has not been contoured from the available data because they are too widely spaced to adequately resolve the short wavelength anomalies (compared to the gravity field), because the several different surveys have been reduced according to varying geomagnetic reference fields (IGRFs, see section 2.1), and because the track cross-over errors within, and particularly between, cruises are too high. The use of different IGRFs is almost certainly responsible for a large part of each cross-over error between cruises. Fortunately good magnetic anomaly compilation maps for the Rockall area have been published by Roberts and Jones

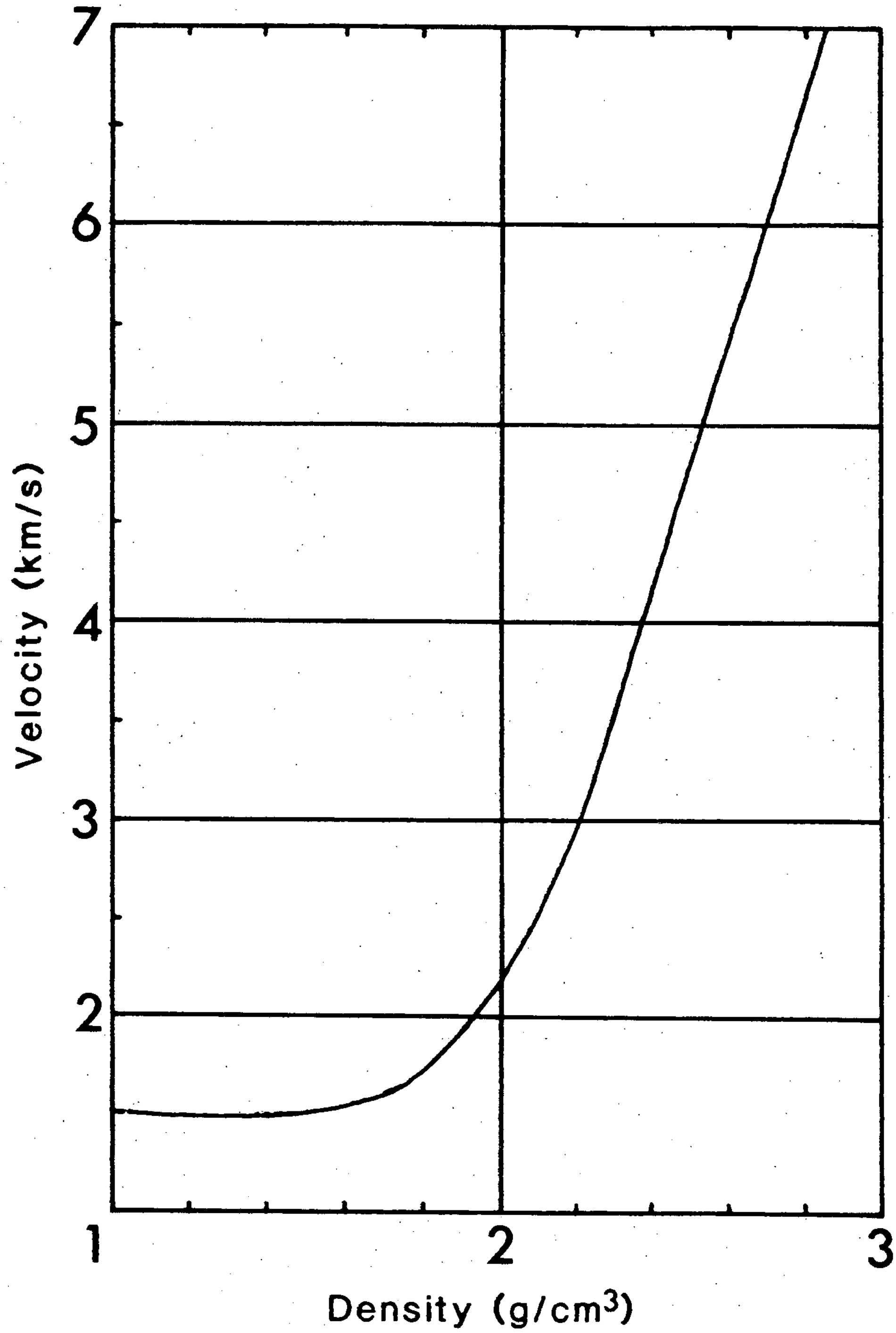


Figure 2.4 The Nafe-Drake curve illustrating the relationship between density and compressional wave velocity of common rock types. Redrawn from Nafe and Drake (1963) and Bott (1982).

(1975) and Max et al. (1982), and these have been useful for qualitative studies in the same way that the free-air anomaly chart was with respect to the gravity field.

Forward magnetic modelling using the ICL 2900 mainframe computer has been performed along selected tracks where an interesting and potentially informative anomaly profile is observed. The anomaly contour charts are used both to locate such profiles and to assess the validity of the assumption of two dimensionality. This modelling faces the same problems of ambiguity as that of gravity modelling and is made harder to use and interpret owing to the greater number of variables involved. It is required to know the magnitudes of the remnant and induced magnetisation, the orientation of the magnetic bodies, the inclination and declination of the present and remnant magnetic vectors, and so on. Convenient values for some of these variables have been assumed for the purposes of modelling, but an approximate knowledge of the inclination and magnitude of the remnant magnetic vector is essential. Suitable ranges of values for these are widely reported in the literature.

There have been many studies concerned with the source of the linear magnetic anomalies detected over oceanic crust (Heirtzler et al. 1968; Parker and Huestis 1974; Lowrie 1974, 1977; Cande and Kent 1976; Kent et al. 1978; Harrison 1981; and Swift and Johnson 1984); these studies are relevant to the modelling and interpretation of the magnetic signature in Rockall Trough. It is unlikely that there are any magnetic sources deeper than 15 to 20 km in the lithosphere since at these depths the Curie point of the magnetic minerals is exceeded (Harrison 1981). Hence the magnetic field tends to reflect the shallower crustal geology, unlike the gravity field which is influenced predominantly by the deeper crust and upper mantle.

Although no magnetic modelling in three dimensions has been attempted, a computer program supplied by Durham University (Nunns 1980) has enabled 2D inverse modelling to be carried out along two lines (Shackleton 3/79 line 14 and Challenger 6/80 line 2; section 7.2) that traverse a number of high amplitude, curvilinear anomalies. The magnetic anomaly profile observed along these two tracks is used to predict the thickness of the magnetic source layer and the variation of magnetic intensity within it, given certain initial assumptions (two dimensional case, fixed direction for remnant magnetic vector, and constant layer thickness).

2.5 Interpretation of seismic reflection profiles

The seismic reflection method produces a travel time cross-section of the crustal layers beneath the track followed by the survey ship. The double-channel and some of the multichannel profiles used in this work are unprocessed and represent the technique in its most basic form. Because of the strong visual element involved in seismic reflection interpretation it is tempting to view the results (the profiles) as geological cross-sections, albeit with distorted horizontal and vertical scales. Sadly this is not the case and the profiles need to be interpreted with caution. The slow and ever-changing speed of the ship, together with the slow shot rate of the air guns, means that the horizontal scale is not constant and that the profiles exhibit a high vertical exaggeration, typically about 4:1 or worse. These shortcomings are not so acute on the multichannel profiles.

An important objective here has been to distinguish seismic (or acoustic) basement from the overlying sediments and bodies within them (Fig. 2.5). The sedimentary column is usually characterised by good sub-horizontal seismic layering resulting from velocity and density changes between different lithologies or across unconformities (Fig. 2.5). A measure of these changes is the contrast in acoustic impedance (the reflection coefficient) across an interface; this may be small, giving rise to seismically transparent layers, or large in which case strong reflectors are observed. Deep drilling in the vicinity of seismic profiles indicates that many of the strong reflectors occur at the level of unconformities.

Seismic basement refers to the deepest recognisable coherent reflection event in the area under consideration. It does not have to be coincident with geologic or economic basement, indeed often it is not. In Rockall Trough and beneath its continental margins seismic basement usually has an irregular surface and a strong reflection with plenty of diffraction events (Fig. 2.5). The variation in these qualities, where they are well seen, can enable the distinction between continental and oceanic basement (see, for example, Chapters 4 and 5). However, in some instances the nature of the seismic basement cannot be identified with certainty, as Chapter 5 testifies.

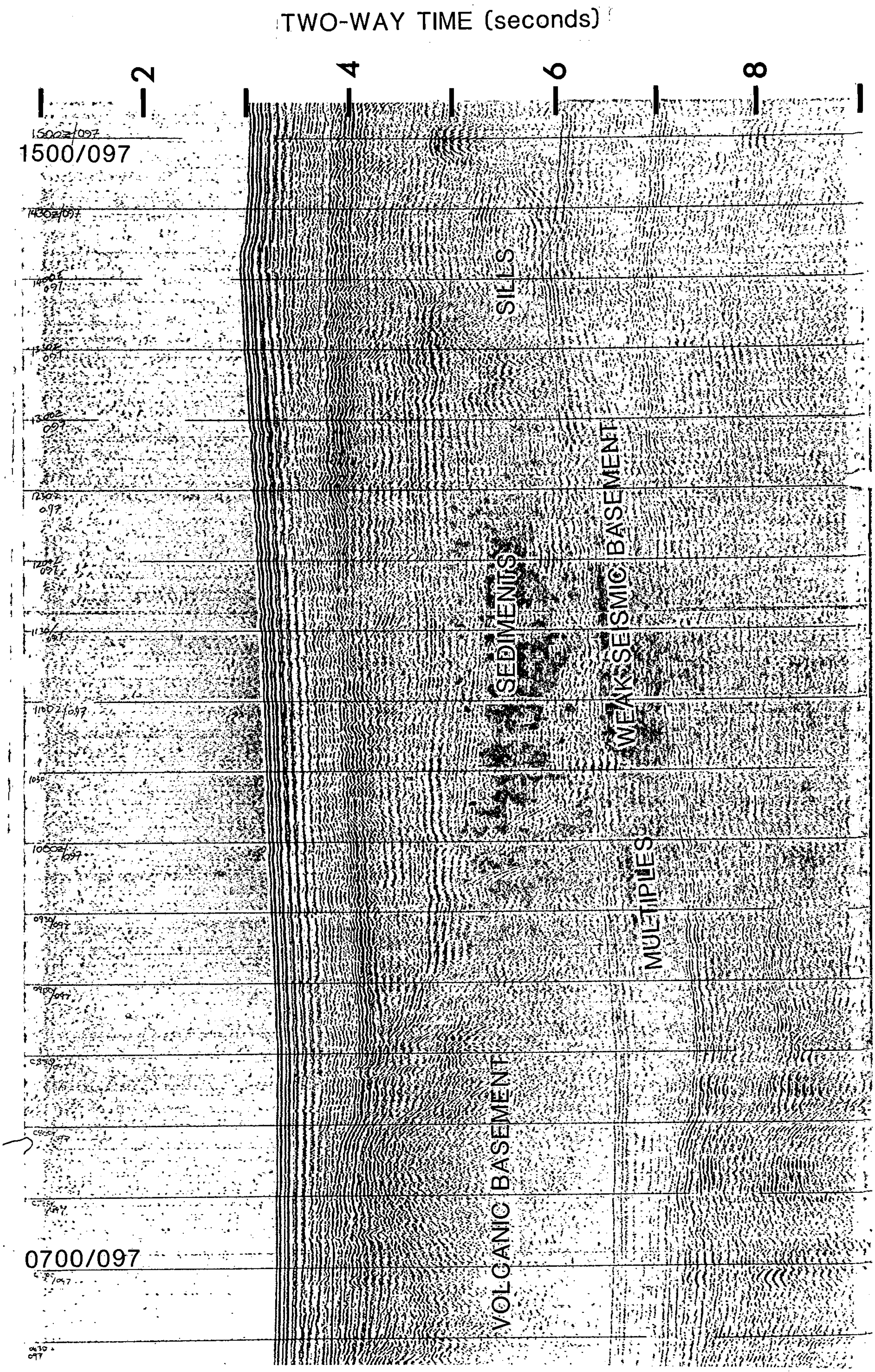


Figure 2.5. Representative section of Challenger 80-2 seismic profile (see Chart 1 for location)

Despite the words of caution expressed at the start of this section, seismic profiles are very useful in showing the general structure and disposition of the sediments and basement. Faults can often be detected by offsets to reflectors or the presence of diffraction events. The variations in the time-thickness of the sedimentary layers are readily visible and these have been quantified to an extent by constructing charts of two-way travel time down to and between conspicuous and persistent reflectors. The compilation of these isochron and isopach maps (e.g. Charts 4 and 5) depends on the ability to trace distinct reflectors from one seismic profile to another with reliability. It is necessary then to ensure that the reflectors tie-in properly at the intersections of seismic lines. In this way the regional geometry of the main unconformities can be established, as well as a crude seismic stratigraphy if geological information is available from drilling (Payton 1977; Masson and Kidd, in press).

The identification of finer scale sedimentary layering and small internal unconformities within the main sediment intervals depends on the resolution of each seismic profile (Fig. 2.5). In Rockall Trough the reflection profiles have been used to show the pattern of build up of the sediments and the structures, such as sediment waves, within them. Extrusive and intrusive igneous rocks within the sediments can also be recognised by their typically strong reflections and diffractions and, in the case of sills, their abrupt terminations. More detailed treatment of seismic reflection interpretation and applications can be found in Sheriff (1981), Sheriff and Geldart (1983) and McQuillin et al. (1984).

3. SEISMIC STRATIGRAPHY IN THE ROCKALL AREA

3.1 Introduction

It will be realised from reading Chapters 4 through to 7 that three distinct geological and tectonic provinces can be recognised from the geophysical data: the continental shelf and slope; the Gibbs Fracture Zone and the oceanic realm to the south; and the deep Rockall Trough. Although this is a convenient broad division of the area difficulties are encountered when one tries to discuss each province separately. It seems unnatural and disjointed to discuss the continental shelf and slope without attention also being given to the geology in the Trough immediately beyond the base of the slope. Similarly, developments in the southern-most Trough need to be related to those of the Charlie-Gibbs F.Z. and the accompanying formation of oceanic crust.

For these reasons this chapter aims at providing a tool by which the three different environments can be set in context, both individually and together, in terms of the relative ages of the important tectonic and stratigraphic events. A review is given of the seismo-stratigraphic work that has been undertaken offshore west of Britain and Ireland, and specifically in the Rockall area; the stratigraphy of the continental margin off the North American seaboard is also of importance. From this is established a nomenclature for the seismic reflectors in the present research area which, in addition to being up to date and as accurate as possible, serves to connect the histories of the widely separated areas, thus enabling them to be considered as a whole - the final aim being to form a coherent regional evolutionary story.

The stratigraphies that have been proposed so far are all necessarily rather general because of the poor quality and low resolution of most of the seismic profiles used, and because of the paucity of geological control from drill sites. Most of the dating of reflectors has been made possible from Deep Sea Drilling Project sites; no commercial core data have been released. There has been a considerable degree of ambiguity concerning the extrapolation of stratigraphic information from one area to another owing to changes in bathymetry, sediment thickness and lithology, and variations in the persistence and amplitude of the relevant reflectors and the intervening seismic sequences. These factors are all closely related and the difficulties of extrapolation that they induce have been the stumbling-block of this and previous seismic stratigraphies since no deep boreholes have been drilled in Rockall Trough.

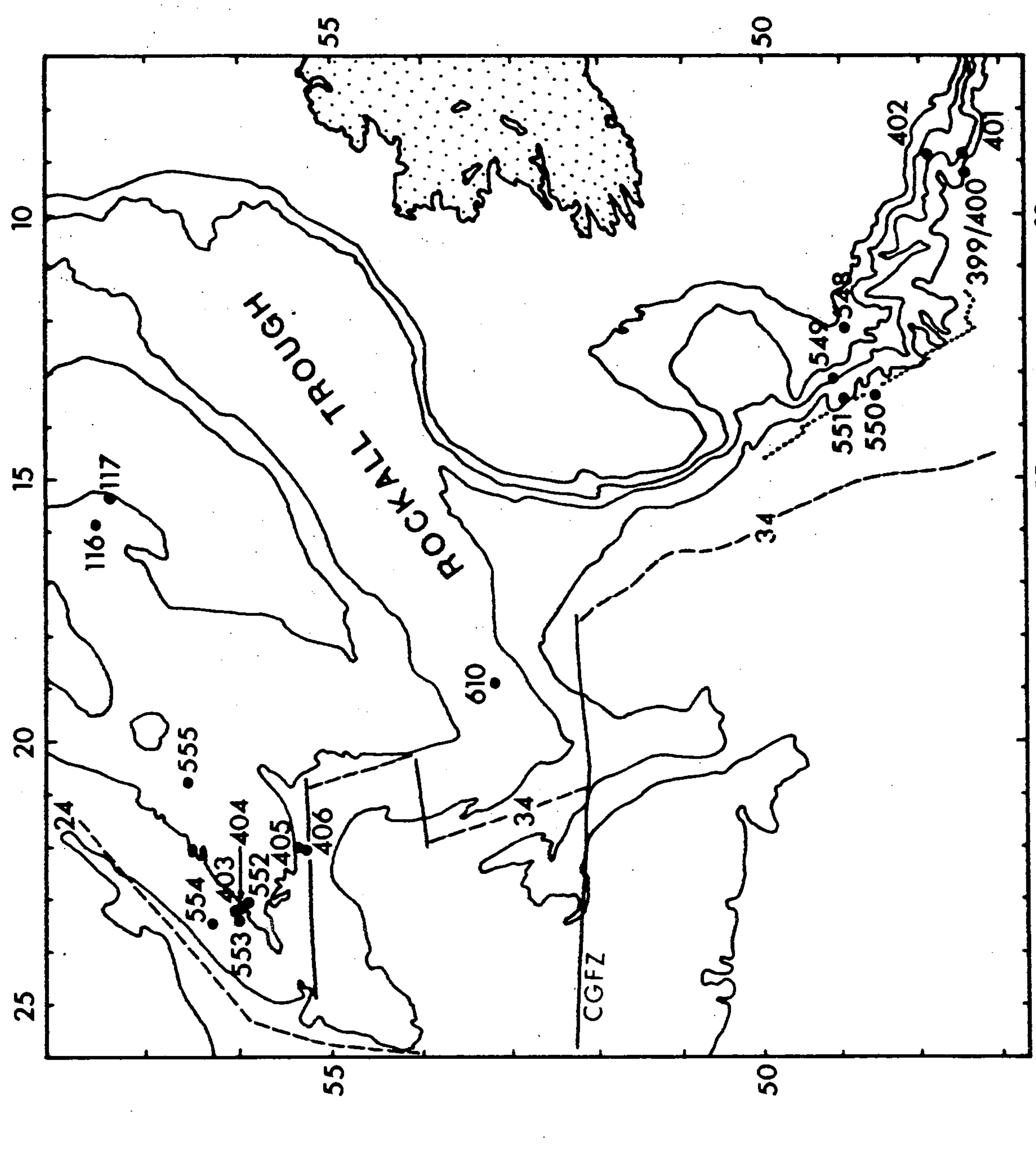
Unfortunately a wide variety of names and labels have now been given to certain reflectors and this has contributed to much of the present confusion. This author has strived to be consistent with the generally accepted nomenclature; the names of the main reflectors have been changed as little as possible, though where they are the new names have been chosen in keeping with the older labels. No attempt is made to invoke or enforce a fresh all-embracing seismic stratigraphy for this would merely add to the existing complexities.

3.2 Previous stratigraphic work

The first attempt at assigning ages to seismic reflectors in the Rockall area was that of Jones et al. (1970). Penetration of their seismic profiles over much of the Rockall Trough was generally limited to a strong sedimentary reflector, or package of reflectors, that usually occurred at between 0.75 and 1.0 s two-way travel time (TWT), corresponding to depths roughly between 0.75 and 1.0 km. Comparison of this reflector, which they named R, with the seismic refraction studies of Hill (1952) and Ewing and Ewing (1959) indicated that it was also a good refractor and as such it may mark an important lithological boundary separating sediments above, with seismic velocities less than 2.0 km/s, from those below with higher velocities.

Jones et al. (1970) traced reflector R around Lorien Bank and over oceanic crust south-west of Rockall Plateau using intersecting seismic profiles. In this area their profile P showed that R probably occurred at fairly shallow levels beneath a prominent scarp; a piston core at this site (55°18'N, 23°00'W) penetrated Upper Oligocene soft chalk at total depth (11 m), approximately 90 m above R. From this they suggested that the R reflector may be as old as the Eocene, and noted the similarity in age of the Lower to Middle Eocene chert beds sampled during Joides Leg 2 (sites 8 to 12) in the western and eastern basins of the North Atlantic. Jones et al. also identified a drift feature, similar to the Feni Ridge, in the

Figure 3.2 Map illustrating DSDP sites in the Rockall area. Refer to text for references and discussion of sites. Bathymetry and symbols as for Figure 1.5.



Labrador Basin south of Cape Farewell, Greenland in which they recognised a persistent reflector ?R. They dated ?R loosely as Middle to Upper Eocene and equated it with R in the Rockall region. R and ?R were considered to relate to changes in oceanographic circulation which developed in early Tertiary times in the opening N. Atlantic. Current controlled differential deposition, giving rise to the sediment drifts, was thought to be in operation at or soon after the formation of R and ?R.

The sediments in the Hatton-Rockall basin were first described by Roberts et al. (1970) who identified a basin at least 1.7 km thick in places in which they recognised four prominent sedimentary reflectorsabove a possible seismic basement event (Fig. 3.3). The upper four reflectors were labelled 1,2,3 and 4 in order of increasing age with reflector 4 appearing at about 1.0 s TWT below the sea bed in the central parts of the basin. No direct dating information was available to them to constrain the ages of each reflector. But they did note that the thin sedimentary cover on the adjacent banks is continuous with the younger strata in the deeper basin. Roberts et al. believed the sediments older than reflector 4 were deposited before the early phase of spreading in the N. Atlantic between N. America-Greenland-Rockall and Europe. The sediments in the interval between reflectors 1 and 4 were considered to have been laid down during the subsequent spreading episode, while those above reflector 1 were related to the younger accretionary phase between Greenland and Rockall Plateau. These relationships would place reflector 4 in the middle Cretaceous (c. 95 m.y.B.P.) and reflector 1 near the top of the Palaeocene (55-60 m.y.B.P.) - though at the time of writing Roberts et al. (1970) did not know the precise timing of these events.

Scrutton and Roberts (1971) extended this reflector scheme into the Rockall Trough but did not attempt any rigorous dating of the reflectors. They did note, however, that reflector 3 seemed to form the base of the Feni Ridge and accordingly assigned it an age near the base of the Tertiary, following Jones et al. (1970). Furthermore they made the significant observation that the sediments below reflectors 3 and 4 were probably deposited after the formation of oceanic crust in the floor of the Trough during the late Mesozoic -this correlation is at variance with the Roberts et al. (1970) age assignments.

Figure 3.3 Seismic stratigraphy across Hatton-Rockall Basin (after Roberts et al. 1970).

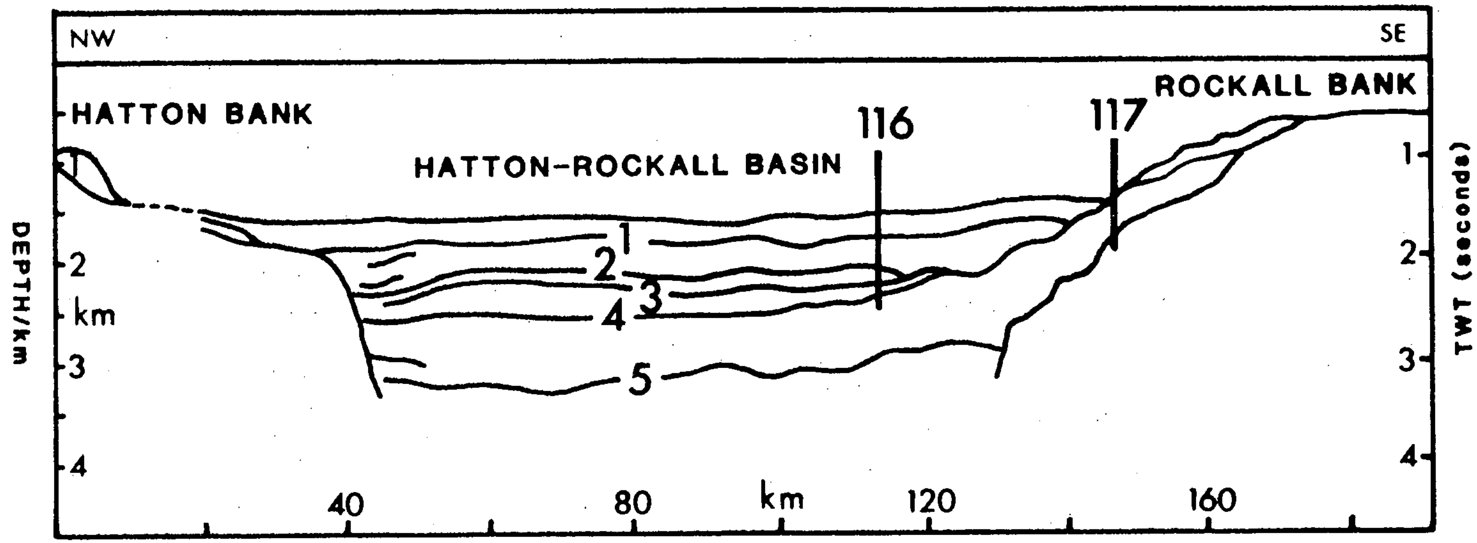
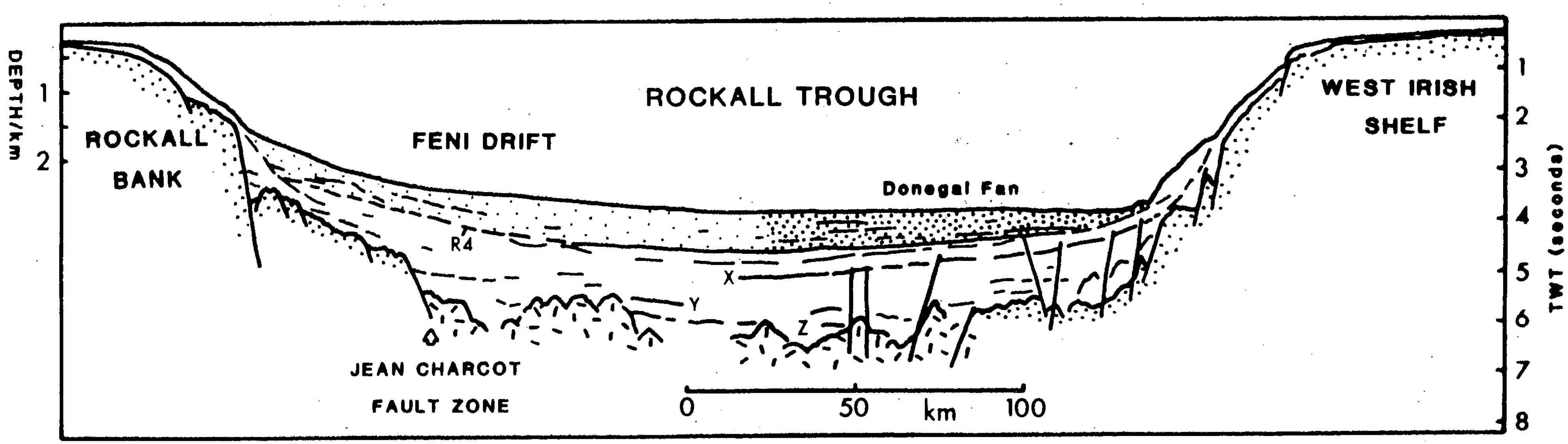


Figure 3.5 Seismic interpretation and stratigraphy across Rockall Trough (after Roberts 1975).



During June-August 1970 Glomar Challenger drilled nine sites (111-119) in the N. Atlantic on DSDP Leg 12 (Laughton, Berggren et al. 1972). Two sites, 116 and 117 (Fig. 3.3), were occupied in the Hatton-Rockall Basin with a view to establishing the subsidence history of the continental Rockall Plateau. The seismic stratigraphy at these two sites conforms with the labellingthough not the ages of Roberts et al. (1970) given above. At site 116 reflectors 1,2 and 3 do not correlate with unconformities or lithological boundaries, though reflectors 2 and 3 do relate in general to an increase in drilling hardness. Reflector 4 was seen to have a fairly diffuse seismic character, compared with the stronger and clearer 1 and 3 events above, and it was thought to arise from increased lithification of chalks and associated cherts corresponding to an Early-Late Oligocene unconformity between 730 and 750 m sub-bottom (Fig. 3.4).

Between sites 116 and 117 reflectors 1,2 and 3 pinch out against the underlying reflector 4 which is then observed at fairly shallow depths below site 117 (Fig. 3.3). Here reflector 4 is again rather diffuse on the seismic records, a result of stringers of hard Oligocene cherty limestones which are spread over a thickness of a few tens of metres. Reflector 4 is associated with an angular unconformity: the overlying sub-horizontal Oligocene limestones pinch out against the Early Eocene clays beneath, which rise steeply up the basin margin (Figs 3.3 and 3.4). This unconformity seems to lengthen in age from the centre of the basin, where the sediments may be conformable or marked by a break of up to 7 m.y., to its margins where a break of about 20 m.y. is recorded (site 117; Laughton, Berggren et al. 1972).

The widespread continuity of the zone of diffuse reflectors usually between 0.75 and 1.0 s TWT as evidenced from seismic profiles, and which is thought to be equivalent to a variable sequence of cherts and chalks, was first highlighted by Roberts (1975). Earlier Ruddiman (1972), in his study of the Iceland Basin, identified a reflector which pinches out to the west over the Reykjanes Ridge flank on 37 m.y. old oceanic basement (Eocene-Oligocene boundary) and which he termed R to correlate with the same event observed by Jones et al. (1970). In noting the close similarity between reflector 4 in the Hatton-Rockall Basin and R in Rockall Trough and south and west of Rockall Plateau, Roberts (1975) renamed the event R4 and suggested that it marks the top of Late Eocene lithified oozes beneath thick

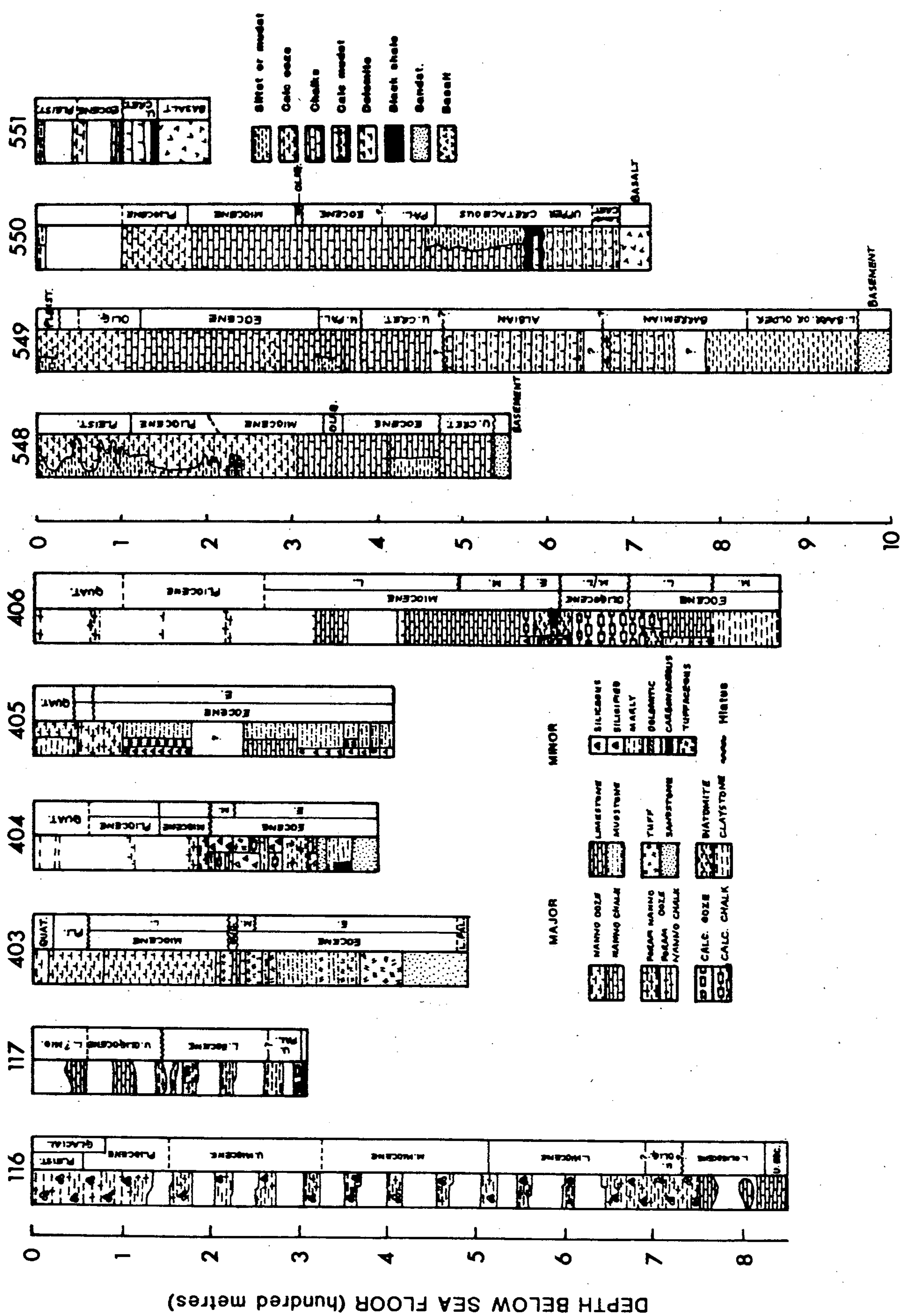


Figure 3.4 For explanation see overleaf.

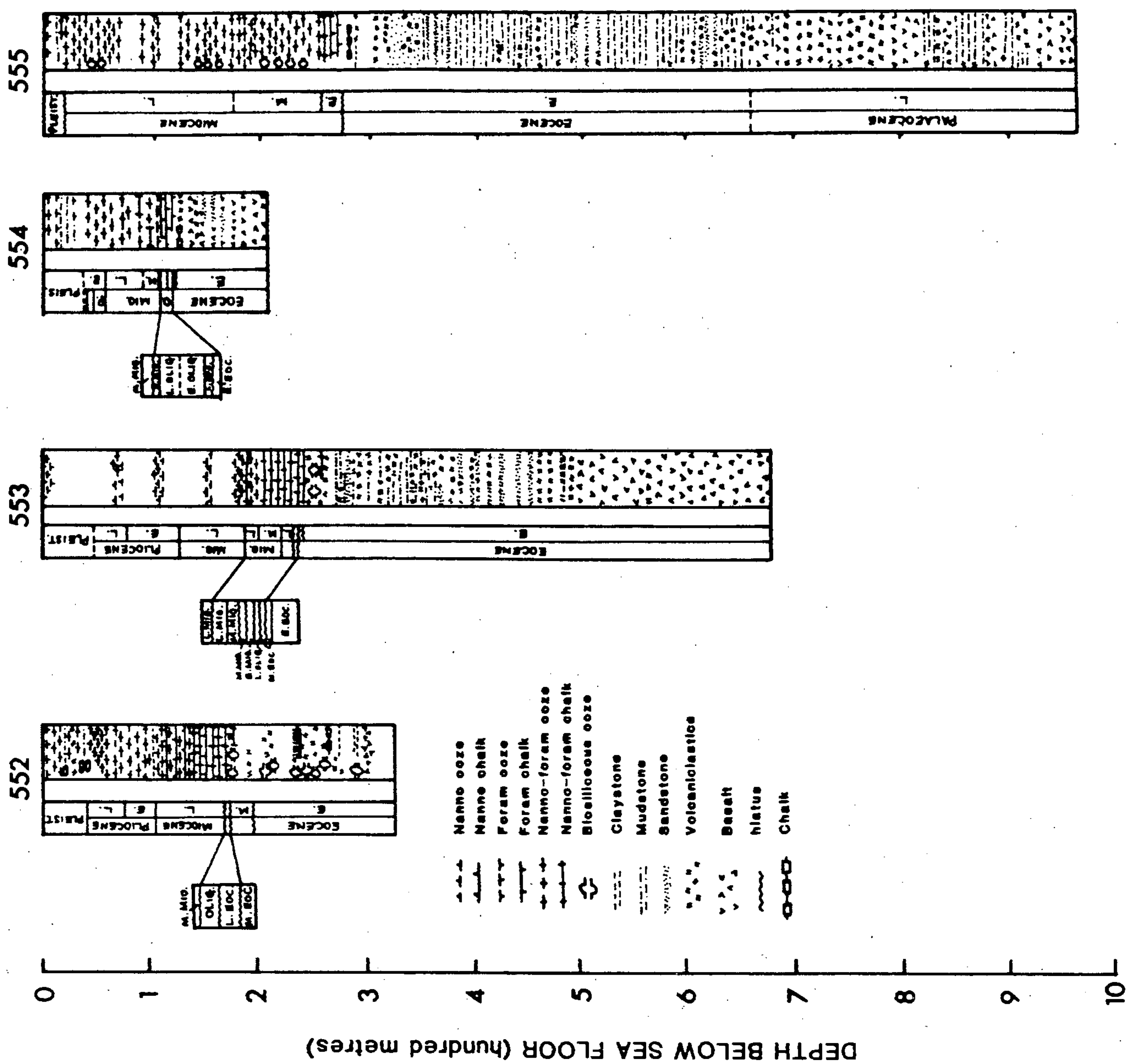


Figure 3.4 (contd). Simplified lithological logs of 14 DSDP drill sites in the vicinity of Rockall Trough. See text for site references and discussion. See Figure 3.2 for site locations. For a more detailed description of the lithological symbols refer to the explanatory notes section in the apprpr-iate DSDP Initial Reports volumes.

cherts. The lateral continuity of reflectors 1,2 and 3 away from Hatton-Rockall Basin, in a like manner, was not recognised.

Although Laughton, Berggren et al. (1972) identified reflector 4 as an intra-Oligocene unconformity at site 116 it was not cored but estimated from sedimentation rates to be a hiatus of 10-15 m.y. Roberts (1975) proposed instead that there was merely reduced, yet continuous sedimentation during the Oligocene at site 116 and so favoured an older Late Eocene age for R4. Miller and Tucholke (1983) agreed with this reinterpretation from their own analysis of the foraminifera and coccolith faunas, and further corroboration comes from DSDP Legs 48 and 81 over SW Rockall Plateau where highly condensed Oligocene successions have been recovered indicative of slow sedimentation rates (Montadert, Roberts et al. 1979; Roberts et al. 1982).

Roberts (1975) described three important reflectors X, Y and Z in the predominantly transparent pre-R4 sedimentary sequence in the Rockall region (Figs 3.5 and 3.6). Reflector X, the shallowest, was originally identified and dated south-west of Rockall Plateau where it drapes oceanic basement and is conformable with R4. It was dated at about 60 m.y. from its pinch out on oceanic crust of that age and Roberts speculated that it may be related to the onsetsea floor spreading between Greenland and Rockall. He traced X into Rockall Trough and remarked that in the south the X and Y reflectors were unconformable because of warping of the latter. In the northern axial regions of the Trough R4, X and Y were seen to be conformable. The present author has observed the same pattern and recognised a causal relationship between the occurrence of the unconformity and the distribution of supposed volcanic ridges in the southern parts of the Trough which result in onlap of the Y event.

Reflector Y was dated by Roberts (1975) as c. 76 m.y.B.P. owing to its pinch out over anomaly 32 oceanic basement south of the Charlie-Gibbs Fracture Zone (Fig. 3.6). He followed it north into Rockall Trough where it defines a pre-Y interval which he further subdivided into two sequences by a reflector Z which occurs in the deeper parts of the basin and appears to be draped over presumed oceanic basement. Roberts tentatively estimated Z to be about 100 m.y. old from a consideration of overburden thicknesses and sedimentation rates (Fig. 3.6). If, as is now widely thought, Late Albian to

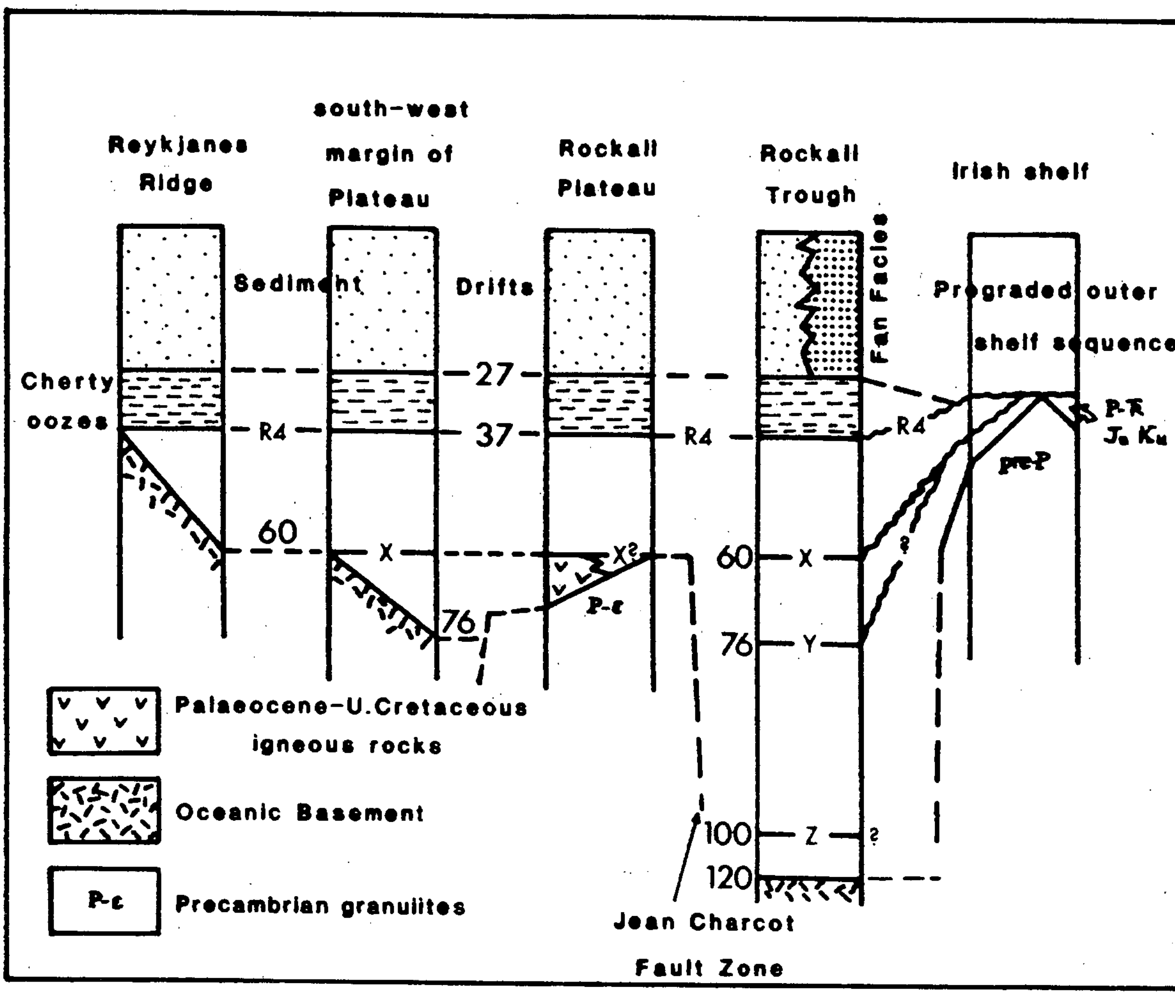


Figure 3.6 Regional seismic stratigraphy of the Rockall area according to Roberts (1975). Refer to text for full discussion. Numbers beside lithological columns denote approximate age (in million years) of major reflectors. Solid wavy lines are unconformities.

Santonian (c. 100-83 m.y.B.P.) oceanic crust exists beneath Rockall Trough then his estimate for Z is about 20 m.y. too large, at the very least.

The important stratigraphic information from DSDP Legs 48 and 81 around SW Rockall Plateau (Montadert, Roberts et al. 1979; Roberts et al. 1982) is summarised in Figure 3.4. The most obvious and perhaps surprising feature evident from this figure is the condensed or absent Palaeogene and early Neogene sections in each of the eight drill holes. At sites 404 and 405 reflector R4 was correlated with a hiatus encompassing the Middle Eocene to Late Miocene (Roberts, Montadert and Searle 1979) - a hiatus caused by erosion, non-deposition or both. At site 406 R4 is marked by a shorter Late Eocene to Middle Oligocene hiatus. And at site 403 it was considered to arise from a late Early Eocene tuff bed which marks an unconformity and which lies just beneath two other closely spaced unconformities of Middle Eocene-Oligocene and Oligocene-Late Miocene age, respectively. Jones and Ramsay (1982) also found an Ypresian age (50-55 m.y. B.P.) for tuffaceous material dredged in the vicinity of the Wyville-Thomson Ridge. They equated it with the widespread lower Eocene ash falls of the North Sea and Rockall Plateau.

Reflector R4 can be correlated broadly with the striking condensed sequences or hiatuses ranging from Early Eocene to Middle Miocene at sites 552 to 555, although Roberts et al. (1982), who discussed the stratigraphic logs,made no attempt at seismic stratigraphic correlations themselves. The Middle Eocene through Oligocene hiatus at site 555 - a stratigraphic break of about 30 m.y. - is particularly noteworthy. What is apparent from the borehole data is that the R4 reflector cannot be ascribed to any definite lithological boundary or length of hiatus: it is different things at different times at different localities. This is entirely to be expected if, as many workers now believe, the diffuse R4 event is a manifestation of profound changes in oceanic circulation at the Eocene-Oligocene boundary; one would not envisage the erosional and non-depositional events, which are thought to have accompanied these changes, being synchronous over such a large area with its marked bathymetric relief. Evidently then caution is required in pursuing long distance correlations of the seismic reflectors in this area.

Roberts et al. (1979b) recorded the presence of a strong, generally flat-lying reflector dividing the pre-Oligocene interval in the vicinity of Leg 48. They ascribed this reflector to a velocity inversion at a Late Palaeocene extrusive horizon which they suggested may correlate with the subaerial lava flow drilled at the base of site 117 (Fig. 3.3). They also speculated that this Late Palaeocene reflector could mark the onset of spreading between Greenland and Rockall Plateau-Faeroe Shelf. If this is so then the presence of similar reflecting horizons in the nearby Rockall Trough would seem most likely bearing in mind its proximity; indeed there are a number of possible candidates for this charge (see next section).

There is some evidence from Leg 48 for a localised parallel unconformity (disconformity) in the Middle Miocene (Roberts et al. 1979b) that divides the post-R4 sequence. It is marked by a sharp change in seismic character corresponding to a sharp upwards decrease in silica content. Below the unconformity strong reflectors are caused by diatomite beds, while above it calcareous lithologies predominate and strong reflectors are absent. The unconformity and attendant reflector were dated at about 16 m.y.B.P. and so are probably the lateral equivalent of the persistent IR (intermediate) reflector of Ruddiman (1972) which he dated as 17 m.y or younger, since it pinched out on oceanic crust of that age in the Iceland Basin. Similar unconformities are seen at all the Leg 81 drill sites (Fig. 3.4; Roberts et al. 1982). In particular, drilling at site 553 penetrated five unconformities in the 1.75 m condensed interval representing the Middle Eocene through to the Early Miocene. Non-deposition or erosion, or both, during that time is indicated by the occurrence of manganese nodules (Roberts et al. 1982).

It is clear from Figure 3.4 that there have occurred a number of important erosional and non-depositional events between the end of the Palaeocene and the beginning of the Late Miocene, a period of almost 45 million years. Within this interval the R4 reflector has been attributed equally to individual horizons of restricted age (e.g. the Early Eocene tuff bed at site 403), as to long hiatuses (e.g. the Middle Eocene to Late Miocene at Site 405). This author considers that these assignments are incorrect in their finer details, and that there are some important inferences to be made from the drilling and seismic information.

An obvious point to stress is that, up until now, any hiatus or lithological change near or straddling the Eocene-Oligocene boundary, with or without an associated seismic reflector, has been termed R4 - a tradition started by Roberts (1975). The Rockall Plateau deep drilling data show that there can be up to three significant and discrete events above this boundary. At certain sites it appears that a younger erosional event has scoured away the evidence of any earlier unconformities. Presumably a seismic reflector corresponding to such an interface should be identified by the younger unconformity. Hence at sites such as 404, 405 and 555 a Middle or Late Miocene period of erosion may have removed all traces of previous unconformities back into the early Tertiary, excavating R4 in the process. Miller and Tucholke (1983) proposed that their R2 was probably responsible for re-excavating the R4 unconformity in places. Should they have named the interface produced R2 (as they did) or R4? These difficulties are made worse by the possibility of long periods of non-deposition, rather than erosion, producing the observed breaks.

The problems of nomenclature are also heightened by the frequently poor resolution of reflectors on seismic profiles, especially where unconformities are closely spaced, as is the case on Rockall Plateau. It is often difficult to decide which reflectors correspond to which unconformities or lithologies, or whether the latter have any seismic expression at all if spurious multiple reflections, bubble pulse events and the like are present. The stratigraphic information from the DSDP sites on the Plateau indicates that the important events which should produce regional reflectors are either very close together or sometimes absent. It is in attempting to extrapolate these reflectors from the shallow banks to the deeper Rockall Trough - where the seismic profiles suggest that sedimentation has been more continuous, depositional rates higher, and therefore potential reflectors more widely spaced - that the present confusion has arisen.

For example, the Middle Eocene to latest Miocene interval, which at site 405 is represented by the single reflector R4 of Roberts et al. (1979), seems to contain four reflectors in the heavily sedimented parts of the Trough (see next section). The quality of the seismic data ultimately determine whether these reflectors can everywhere be resolved and what form they assume in between areas - erosional or non-depositional.

In their reappraisal of the evolution and seismic stratigraphy of Rockall Trough Roberts et al. (1981) reviewed the ages of the R4, X, Y and Z reflectors in the light of new data. R4 was maintained as Late Eocene to Early Oligocene (c. 37 m.y.) in the Trough and equated with Middle Eocene to Oligocene unconformities penetrated during Leg 48 drilling on the northern margin of the Bay of Biscay. Again in the Bay of Biscay they correlated reflector Y with a hiatus of Maastrichtian to Late Palaeocene age at site 400A and with a shorter Mid-Palaeocene hiatus at site 401. In Rockall Trough a reflector occurring at the level of abundant lava flows seen on their seismic profiles was designated Y. The lavas were postulated as belonging to that period of igneous activity peaking at about 52-55 m.y. B.P. and Y was dated approximately that age accordingly.

It now seems from the present work that most, if not all, of these igneous reflectors are caused by sills not lavas; by their intrusive nature sills cannot provide accurate ages of the surrounding sediments. The work of Einsele (1982, 1985) on recent igneous activity in the Gulf of California indicates that sills are usually emplaced at depths of between about 400 m and 50 m, depending on the physical properties of the sediments and magma and the thickness of the water column. Thus in the Rockall Trough reflector Y may be only slightly older than the underlying sills, or it may be some considerably older age; the method is not precise so Y is presumed to be at least 55 m.y. old.

In reducing the age of reflector Y from 76 to c.55 m.y.B.P. it was necessary for Roberts et al. (1981) to revise the age for X from 60 m.y. to between 55 and 37 m.y.B.P. They did not ascribe X to any particular stratigraphic event but noted that towards Biscay it merged with R4 above. Within the pre-Y interval the Z reflector divides an upper and lower seismic unit. The upper unit is seismically transparent for the most part but a number of persistent flat-lying reflectors were noted. Roberts et al. (1981) made a tentative correlation of this upper unit with the pre-Late Palaeocene interval observed at sites 403 and 404. The lower, pre-Z unit, which they suggested was deposited through the rifting and spreading episodes that structured the Trough, is characterised by numerous strong reflectors that locally prograde towards the axis of the

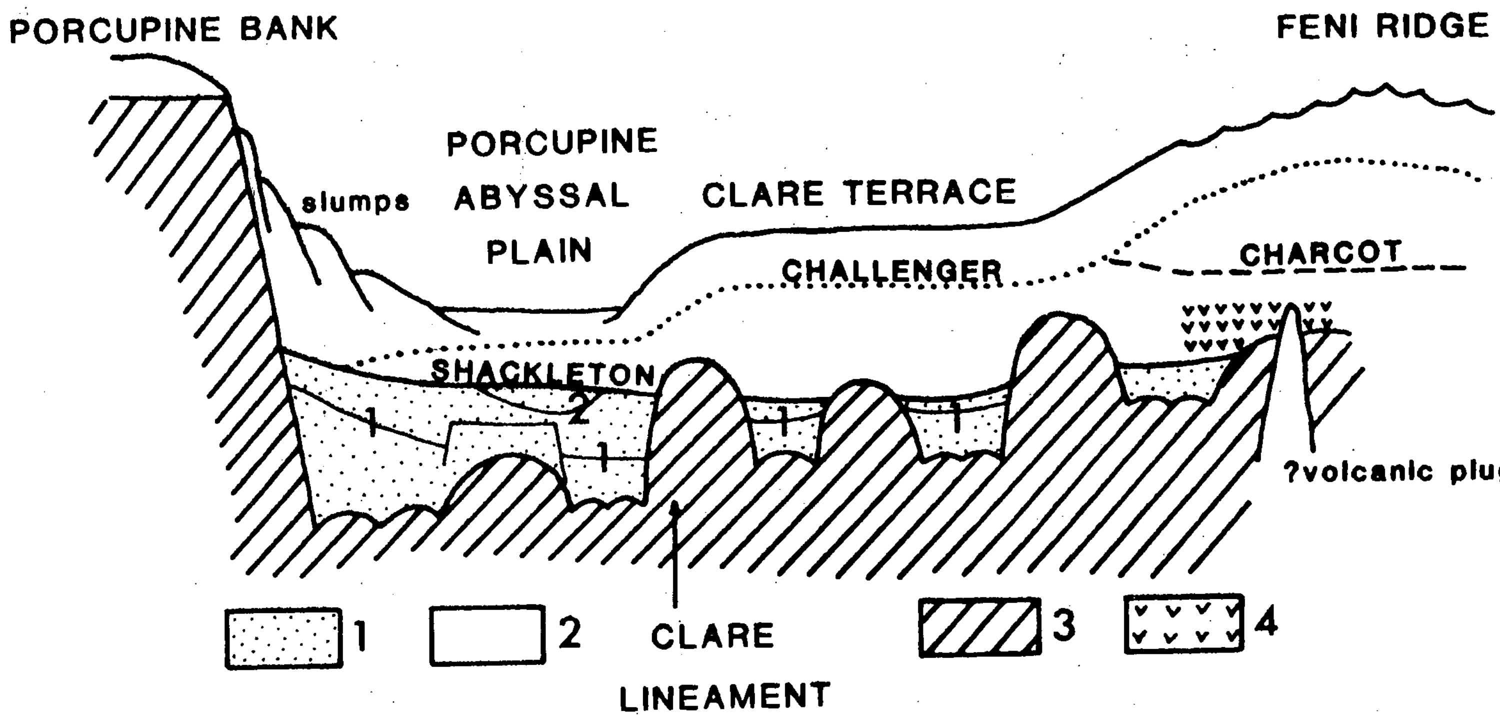


Figure 3.7 Regional seismic stratigraphy adopted by Dingle et al. (1982). Refer to text for discussion. 1 = Pre-Shackleton sequence. 2 = Post-Shackleton sequence. 3 = Acoustic basement. 4 = Lavas (?) or sills. Diagram not to scale.

Trough. But they did not illustrate the thinning of the lower unit from the margins to the axis of the basin that would be expected from this diachronous behaviour.

A qualitative assessment of the regional seismic stratigraphy of the Rockall Trough and immediate surrounds was provided by Dingle et al. (1982) who recognised three widely developed reflectors which they termed Shackleton, Challenger and Charcot (Fig. 3.7). The double-channel reflection profiles they used in the Trough (Challenger 6/1980 cruise; see section 2.2) also form part of the data base for this study.

The Shackleton reflector, which correlates with Y of Roberts et al. (1981) in many places, defines the upper limit of a sedimentary sequence, often seismically transparent, that fills in a very irregular basement relief. In the pre-Shackleton section their reflector 1 is the equivalent of Z of Roberts et al. (1981) since it divides a lower highly reflective unit from an upper paler unit. Their reflector 2 (Fig. 3.7) is not of regional importance. Within the post-Shackleton sequence their prominent Challenger horizon correlates in a general sense with the ubiquitous R4 of Roberts (1975), though Dingle et al. (1982) neither stated this equivalence nor attempted any dating of the reflectors for lack of geological constraints.

Beneath the Feni Ridge drift the Charcot horizon probably correlates with the intermittent X reflector of Roberts (1975) and Roberts et al. (1981). However, inspection of the Edinburgh profiles that Dingle et al. used and also multichannel profiles in the area (Chart 1) shows that the Charcot reflector is continuous over the whole of southern and central Rockall Trough except, perhaps, over some of the shallower seismic basement highs - here and in other areas of thin post-Shackleton sediments the bubble pulse reverberations from Challenger obscure the underlying Charcot event on the double-channel records.

Dingle et al. (1982) state that the main part of the thickness increase below the Feni Ridge is taken up in the Charcot-Challenger package and that post-Challenger sediments do not vary significantly in thickness over the whole area (Fig. 3.7). This is contrary to Ruddiman (1972) and Roberts (1975) who both proposed that such changes are confined to the post-R4 (=Challenger) interval. It has now become clear, from a study of all the relevant data (see next section), that differential deposition has been operating since approximately the formation of the X reflector (partly = Charcot).

Seismic stratigraphic and borehole information from the whole of the North Atlantic were reviewed by Miller and Tucholke (1983). They noticed that an unconformity at the Eocene-Oligocene boundary occurs in every DSDP borehole (16 in all) except sites 116 in the Hatton-Rockall Basin and 112 in the Labrador Basin (Fig. 3.3). They believed that the absence of one at these two sites may be due to their position away from basin margins; all other sites are on or near areas where topographic boundary effects should have intensified abyssal circulation. The present author does not think this is necessarily so at site 116 which lies within a region of quite strong relief (Fig. 3.2). The fact that the R4 reflector is prominent here indicates that an unconformity at the Eocene-Oligocene boundary is not a prerequisite for its formation: some manner of lithological or physical change, or both, must be present. The confusing biostratigraphy at site 116 is illustrated by the slightly different interpretations of Laughton, Berggren et al. (1972), Roberts (1975) and Miller and Tucholke (1983), though the last two are in general agreement.

Miller and Tucholke (1983) identified three horizons in the Iceland Basin-Rockall region above reflector R4 which correspond to reflectors 1,2 and 3 of Roberts et al. (1970) and which they termed sensibly, R1, R2 and R3. From the combined drill site information they were able to characterise each of these major reflectors by lithogic and age limits as follows:

Reflector R4. There is a strong correlation of R4 to the widely distributed unconformity straddling the Eocene-Oligocene boundary. Away from basin margins the unconformity is not seen and there R4 correlates with the uppermost Eocene to lower Oligocene section in the N. Atlantic.

Reflector R3. This is observed only in the Rockall region where it dates to the latest Oligocene-earliest Miocene sedimentary section. There is no consistent lithogic correlation but the seismic signature suggests R3 may cap an Oligocene sequence of cherty chalks.

Reflector R2. R2 is dated as latest Early Miocene at sites 116 and 406 on Rockall Plateau. At most other sites in the N. Atlantic a hiatus is present at this time and it is likely that R2 is another widespread unconformity. In many places it may have re-excavated the R4 horizon. R2 correlates approximately to the level below which siliceous debris is a significant sedimentary component: it also correlates with the IR reflector of Ruddiman (1972).

Reflector R1. This reflector was thought to be important only in the Rockall Plateau region where it dates to the Late Miocene and may represent a lithification boundary (e.g. ooze to chalk) in calcareous sediments.

While this scheme of correlations appears to be accurate and consistent the problem still exists of extrapolation of events into the Rockall Trough. Apart from the recent site 610 drilled on Feni Ridge during DSDP Leg 94 there have been no boreholes drilled in the Trough, and furthermore no precise seismic correlation from the widely drilled Rockall Plateau and Goban Spur/Biscay areas. For instance, the next section indicates that R1 may be a persistent reflecting horizon beneath the Feni Ridge. But this is merely an age correlation with the upper Miocene R1 event put forward by Miller and Tucholke (1983), and not a true tracing of reflector continuity to a drill site. The NW-SE trending graben between Rockall and Lorien Banks would appear to offer the best hope of tie-ins to the 10 DSDP sites on the Plateau.

In their analysis of seismic, biostratigraphic and oxygen isotope data from both sides of the N. Atlantic Miller et al. (1985) interpreted the widespread Horizon A^u off N. America (Tucholke and Mountain 1979) as representing an angular unconformity at the Eocene-Oligocene boundary probably eroded by bottom currents, and not caused by fluctuations in sea level as had been suggested previously (Dillon et al. 1984). Miller et al. (1985) equated Horizon A^u with R4 in the eastern N. Atlantic, hence emphasising the truly regional, oceanic nature of the erosional event and its accompanying reflector.

In the same work Miller et al. (1985) also presented evidence for two other erosional events in the N. Atlantic: one in the Middle Oligocene (c. 30-31 m.y.) and another in the Late Oligocene-Early Miocene (c. 25 m.y.). The latter probably corresponds to R3 and both events were thought to relate to rapid rates of fall of eustatic sea level, rather than to deep water erosional episodes. They could find

Jones et al.(1970)	Roberts (1975) Roberts et al.(1981)	Dingle et al.(1982)	Miller and Tucholke (1983)	Masson and Kidd (in prep.)	This study
			R1	Yellow	R1
				Purple	
R	R4	Challenger	R2	Green	R2
					R4
	X (?Y)	Charcot	R4	Brown	R5
	X or Y			Blue	R6
				Pink	Observed but not correlated
	Y (?Z)	Shackleton		Orange	R7
	Z	reflector 1		Grey	pre-R7 unconformity

Table 3.1. Seismic reflector correlations in the Northeast Atlantic.

no causal relationship between this Oligocene sea level phase and the older deep water phase which formed R4. On the same theme Miller and Tucholke (1983) stated that a coherent erosional pulse may have occurred at the level of R2 in the late Early Miocene, and another pulse (surprisingly not marked by a reflector) almost certainly occurred in the late Middle Miocene above R2. It was not stated whether these two horizons resulted from the sea level fall or deep water erosional processes, but the first option seems the more likely in view of the spatial and temporal proximity to the R3 reflector. What is clear, however, is that the late Palaeogene and Neogene in the N. Atlantic were the setting for frequent oceanic disturbances which resulted in periods of erosion and non-deposition and which are manifested in a complex pattern of unconformities and seismic reflectors.

From the foregoing accounts it is obvious that there is an excess of schemes and terminology involved in the seismic stratigraphy of the eastern N. Atlantic. Table 3.1 is an attempt at clarifying the correlations and ages of the main reflectors discussed in this and the next section.

3.3 A seismic stratigraphy for the Rockall Trough: correlation with DSDP results

An attempt is made at establishing a broad correlation and dating of reflectors between DSDP site 550 on oceanic crust west of Goban Spur (Graciansky, Poag et al. 1985) and site 610 over Feni Ridge in southern Rockall Trough (Initial Reports of DSDP v94, in prep.). The positions of these two sites and the seismic profiles joining them are shown in Figure 3.8. Preliminary details of the site 610 results and the original CM and W1-32 multichannel seismic lines were made available at the Institute of Oceanographic Sciences, U.K. in collaboration with D.G. Masson. The findings of the Tertiary seismic stratigraphy at site 610 are reported in Masson and Kidd (in press). The location of sites 548, 549 and 551 either above basement highs where sediments are thin or on the opposite (landward) side of highs from 550 means that proper seismic correlations with this site are not possible.

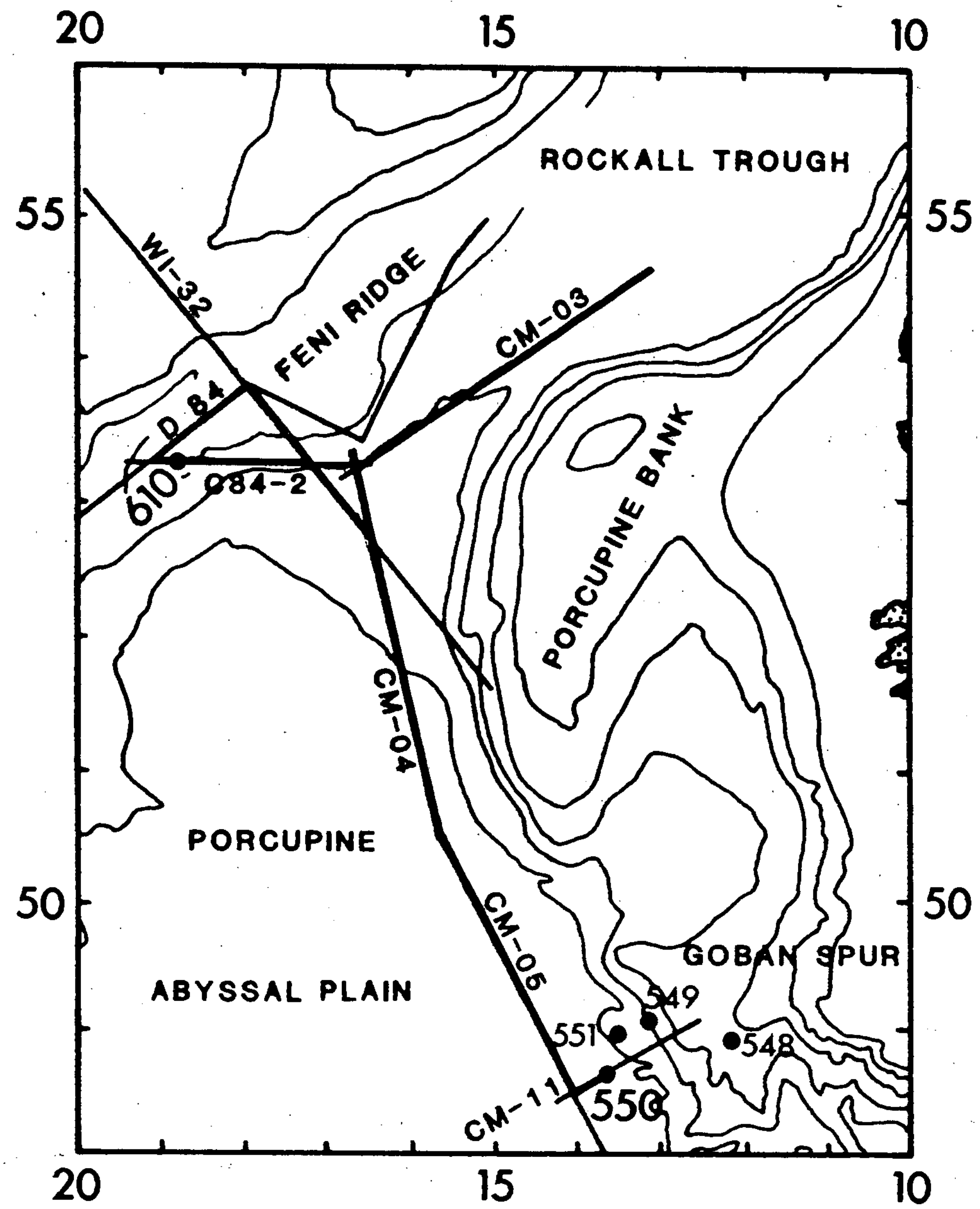


Figure 3.8 General bathymetric map illustrating the position of profiles used for seismic stratigraphic correlation between Goban Spur and Rockall Trough. 200 m isobath and 500 to 4000 m isobaths at 500 m intervals are shown. Solid circles are selected DSDP sites. Emboldened lines indicate sections of profiles used in the correlations and depicted in Figures 3.16 and 3.17 (back pocket).

Seismic stratigraphy at Site 550

The stratigraphic section penetrated at site 550 is depicted in Figure 3.9. The middle Cretaceous (Albian) to Pleistocene sedimentary succession is dominated by nannofossil and nanno-foraminifera chalks and oozes; the oozes are lithified to chalks at approximately 170 m depth below sea floor. Terrigenous sediments are mostly absent except in the Late Cretaceous sequence where a good proportion of claystones and mudstones were recovered. Basalt cored from 685 m to total depth (720.5 m) was thought to represent the top of Middle Albian oceanic layer 2 (DSDP v80, 550 site chapter). Major unconformities and condensed sequences are seen from the Middle Eocene to Late Oligocene and from the Late Cenomanian to Campanian; also, there is a conspicuous unconformity between the deepest sediments and basaltic basement.

Montadert and Poag (1985) presented a general, low resolution correlation of the lithostratigraphic and chronostratigraphic events penetrated at site 550 with the seismic reflectors observed on the CM-11 multichannel reflection profile (Figs 3.8 to 3.11). They identified seismic sequences A, B and C above oceanic basement (Figs 3.10 and 3.11). In the vicinity of site 550 sequence A displays only slight thickness variations over basement highs and is characterised by faint reflectors (Fig. 3.11). A distinct reflector, A2 (possibly R2 of Miller and Tucholke 1983), occurs near the base of the sequence and a more discontinuous one, A1, appears towards the middle but neither can be followed for great distances. Montadert and Poag proposed that the base of sequence A coincides with an important unconformity and/or condensed section between the Early Oligocene and Early Eocene (Fig. 3.9). Contrary to this their Figure 15 shows the base reflector as an Early to Late Oligocene unconformity. In their Figure 13 they equated the same reflector with the bottom of lithologic unit 1b at site 550 corresponding to a depth of 309.5 m and a cumulative two-way time of 0.344 seconds (Fig. 3.10). This ties in well with a reflector at 0.33 s TWT on the CM-11 profile and also the lithology versus depth summary in the 550 site chapter (Graciansky, Poag et al. 1985, p255). Unfortunately, Montadert and Poag then went on to pick two different events at 0.3 and 0.4 s TWT below sea

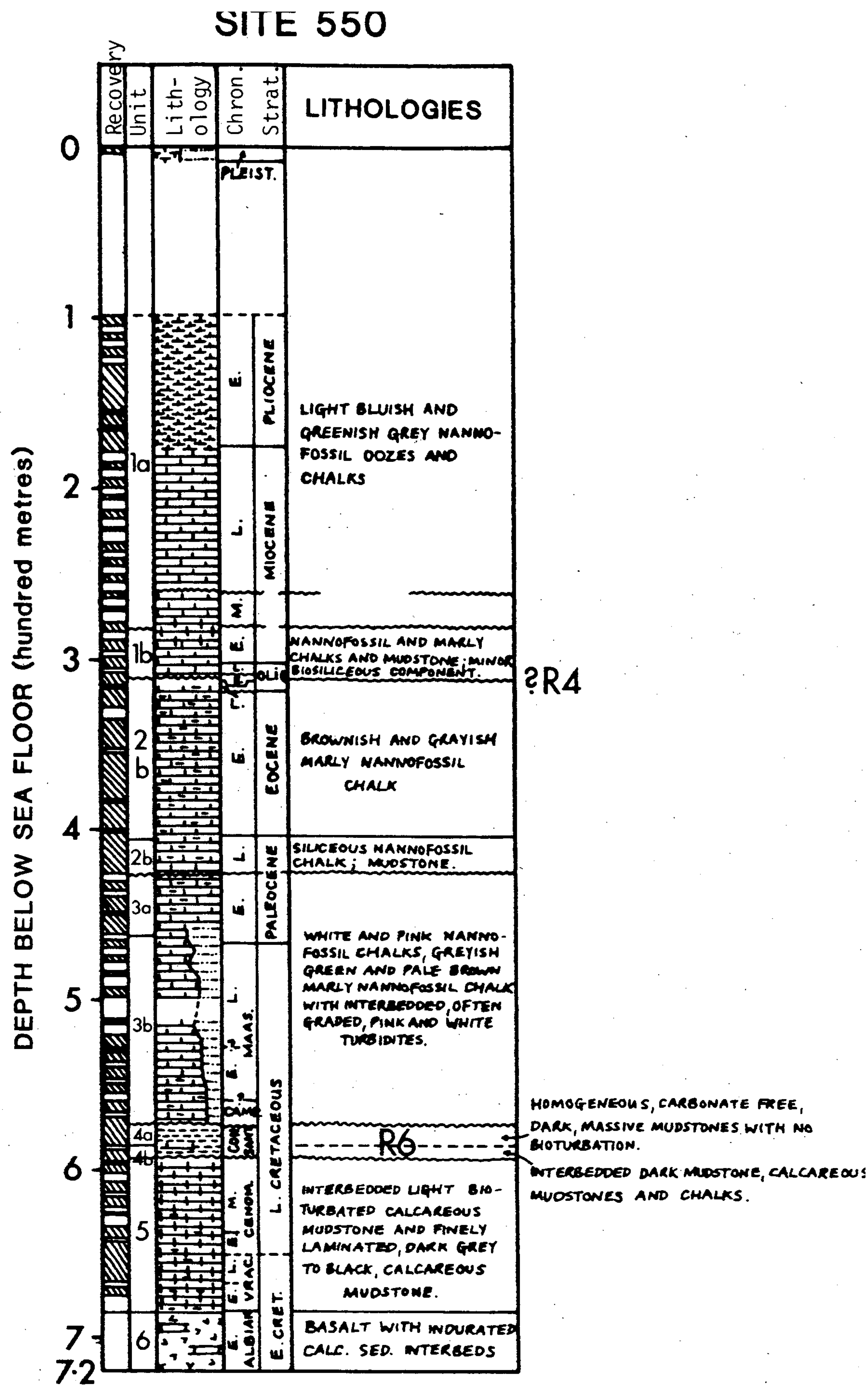


Figure 3.9 Lithological section recovered at DSDP site 550. Redrawn from Graciansky and Poag et al. 1985, p. 265. Refer to text for comparison with CM-11 multichannel seismic profile. R4 and R6 labels mark proposed positions of distinct reflectors observed on the CM-11 profile.

Figure 3.10. Summary of lithostratigraphy, chronostratigraphy and seismic stratigraphy at DSDP Site 550. Compiled from various sources in IR DSDP v80 (Graciansky, Poag et al. 1985).

Lithologic unit	Depth (m)	Lithologic description	Chrono-strat.	Interval thickness	TWT(ms) Site 550	TWT(ms) CM-11	Seismic sequence
	99.5						
1a		Light bluish and grey nannofossil oozes and chalks.	Pliocene to M. Miocene	183.5			
	283						A
1b		Light-coloured nannofossil chalk,marly nannofossil chalk and mudstone; minor biosilic-component.	E. Miocene to L. or M.Oligocene	27			
	310				344	330	R4
2a		Brownish and greyish marly nannofossil chalk.	E.Oligocene to E.Eocene	98			
	408				446	430	
2b		Brownish, grey and olive siliceous nannofossil chalk and mudstone.	L.Palaeocene	18.5			
	426.5				476	460	B
3a 3b		Nannofossil chalks and marly nannofossil chalks with interbedded, often graded, calcareous turbidites.	L.Palaeocene to E. Maas. or L. Camp.	148.5			
	575				591	580	
4a		Homogeneous, carbonate-free, dark, massive mudstones with no bioturbation.	Unknown	9			
	584						
4b		Interbedded dark mudstone, calcareous mudstones and chalks.	Santonian or Coniacian	10			C
	594				614		
5		Interbedded light bioturbated calcareous mudstone and finely laminated, dark grey to black, calcareous mudstone.	M.Cenomanian to L. Albian	91			
	685				689	690	
6		Basalt with thin calcareous sediment interbeds.	?L. Albian	35.5			
	720.5						

floor on the seismic profile through site 550 (their Figures 11 and 15 respectively). Thus the reader is left in considerable doubt as to which reflector they are referring to as the base of sequence A.

The situation is complicated because the M. Eocene to E. Miocene section is represented by only about 10 m of sediment in which the Middle and Late Oligocene are poorly dated. Within this interval unconformities were recognised between the Middle and Late Eocene, and the Early and Late Oligocene; in addition Miller et al. (1985b), in their study of the Late Palaeogene fauna in the same volume, proposed a short break at the Eocene-Oligocene boundary similar to those seen at deep DSDP sites 119 and 400 on the Biscay margin. It seems likely that this last unconformity represents the R4 deep water erosional event and as such might be capable of being traced laterally for large distances.

It is improbable that the CM-11 profile can resolve such fine variations, especially at these depths (in excess of 4.5 km). The interval in question, hiatuses included, is almost certainly represented by just one reflector. Montadert and Poag (1985) stated that the reflector marking the base of sequence A truncates the arched upper reflectors of sequence B (Fig. 3.11). This would fit in well with the assumption that it represents the R4 erosional event in the N. Atlantic (see previous section). And as further support for this correlation the base reflector of Montadert and Poag occurs at 309.5 m at site 550, in exact agreement with the depth assigned to the Eocene-Oligocene hiatus by Miller et al. (1985b). The present author has assumed the base of sequence A to be R4 and this is the correlation followed along the CM survey.

Seismic sequence B is characterised by strong continuous reflectors, the shallower events being truncated by the base of sequence A (Fig. 3.11; Montadert and Poag 1985). It was considered to encompass the interval from E/M. Eocene to E. Maastrichtian or Late Campanian (lithologic units 2 and 3 in Figs 3.9 and 3.10). The sequence is slightly thinner over the basement high beneath site 550 than in the adjacent basins and the basal reflector is seen to erode some of the upper sequence C horizons (Fig. 3.11). The strong reflector marking the bottom of sequence B was thought by Montadert and Poag to represent the 20 m condensed section covering presumed Late Cenomanian to Early Campanian sediments (lithologic unit 4 in Figs 3.9 and 3.10). The equally strong but slightly less persistent

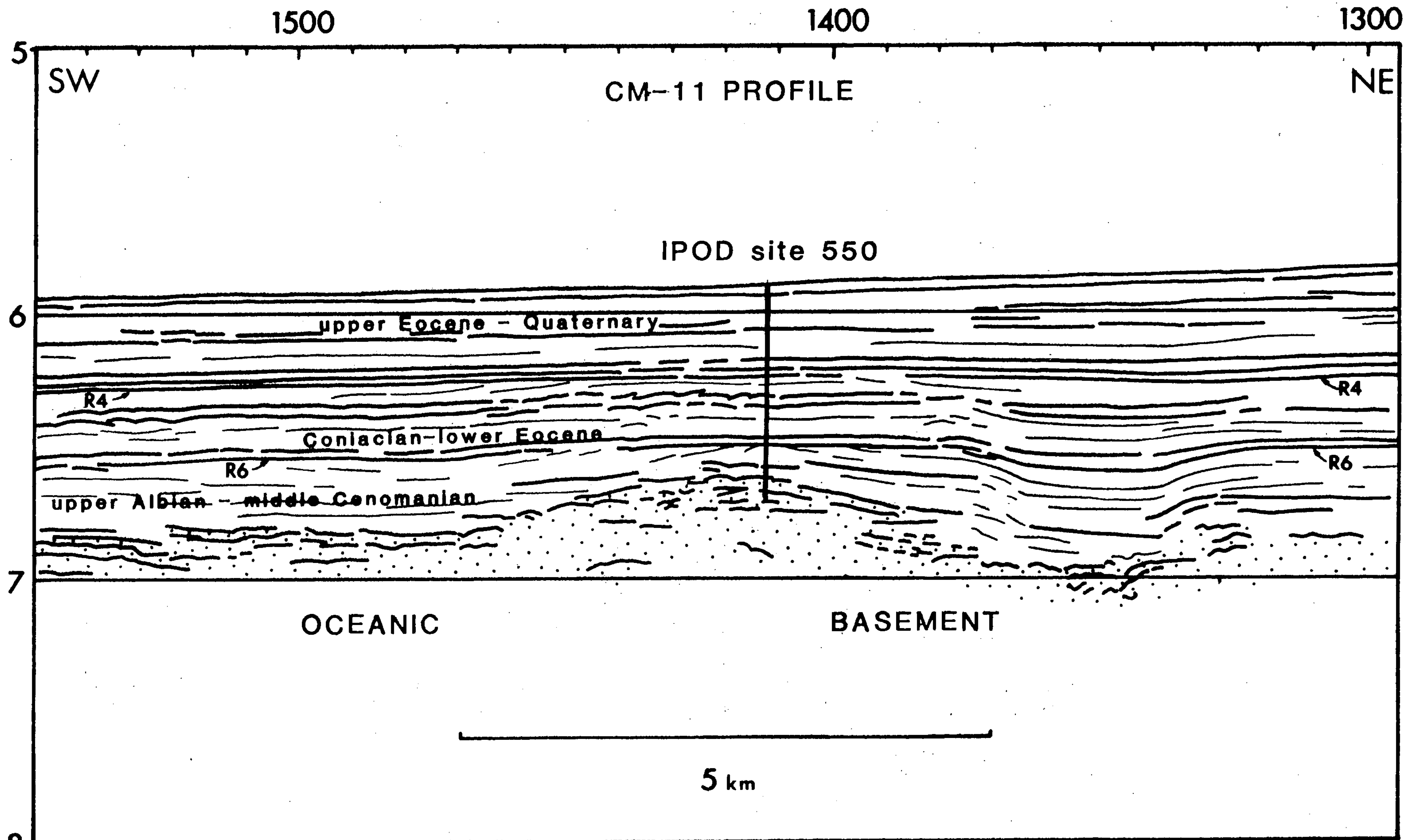


Figure 3.11 Interpretation of multichannel profile CM-11, sps 1300-1550, through IPOD site 550. See Fig. 3.8 for location map. Important reflectors appear as bold lines. Main R4 and R6 events labelled. Original profile courtesy of D.G.Masson.

reflector immediately above the basal event was not remarked upon.

The deepest seismic sequence, C, of Montadert and Poag, in which they included lithologic units 4 and 5 (L. Albian to M. Cenomanian), tends to infill basement lows and does not everywhere cover the basement highs. For this reason parts of the older sedimentary record were almost certainly not drilled at site 550 (Fig. 3.11). Although not strikingly obvious in the vicinity of this site, sequence C is generally transparent with a few faint discontinuous reflectors; it appears to assume a more laminated aspect where the seismic profiles approach the continental margins where the influence of turbidite sedimentation was probably greater.

Seismic stratigraphy at Site 610

Site 610, occupied during DSDP Leg 94, is located on the crest of Feni Ridge in southern Rockall Trough (53°13.3'N, 18°53.2'W; Fig. 3.8). Three holes at this site penetrated predominantly nannofossil oozes, marls and chalks to a total depth of roughly 730 m, though core recovery below 200 m was poor (D.G. Masson, pers. comm.). Masson and Kidd (in press) note that there are no major lithologic changes in the cored sediments except within the glacial cycles which extend to 135 m sub-bottom. In the absence of any seismic profiles passing directly over site 610 they made correlations with the reflectors on the multichannel profile WI-32 (Fig. 3.8), 90 km to the north-east at its closest point but of higher resolution than the nearby single-channel records.

In an attempt to improve the accuracy of the seismic stratigraphy a seismic reflection line (C 84-2) was shot east to west over site 610 (Figs 3.8 and 3.12) during the Challenger 1/1984 cruise (see section 2.2 for details). It was originally hoped to be able to deploy two tuned 40 cubic inch airguns as the acoustic source to achieve high reflector resolution between the intersections with the WI-32 and Discovery 84 lines. The single 160 cubic inch airgun that was eventually employed was initially considered to have provided poor results. But it now appears that the three seismic records obtained along C 84-2 can provide the required resolution.

The display parameters set for the three profiles along C 84-2 and the picks of two-way travel times to distinct reflectors at 05.20 BST Day 165 on each are shown in Figure 3.13. Inspection of the

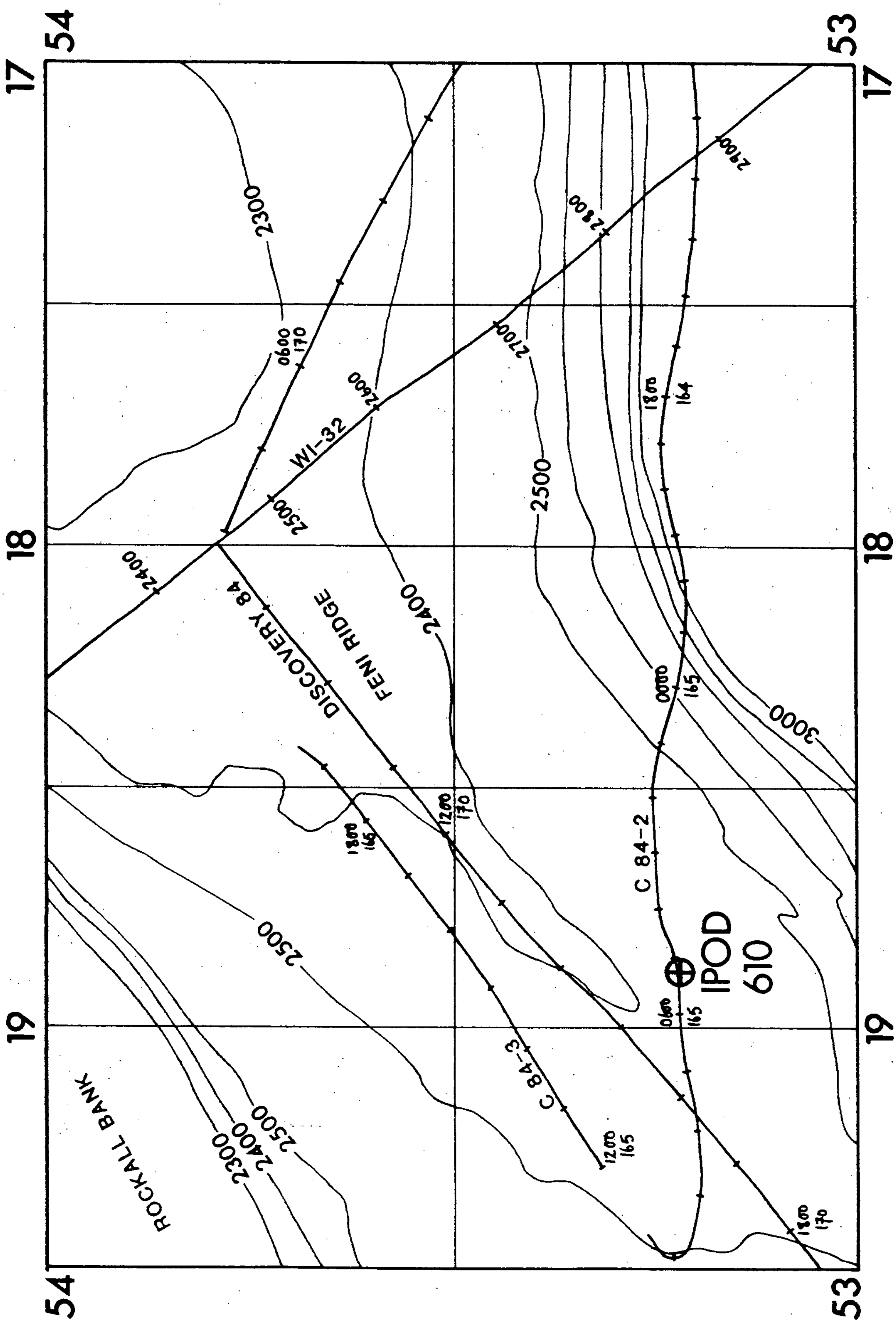


Figure 3.12 Detailed navigation chart of seismic profiles in the vicinity of IPOD site 610. Shot point numbers given for WI-32. Time marks every hour on other tracks. Bathymetry in metres every 100 m.

Figure 3.13. EPC recorder settings and picks of seismic reflectors for profile C 84-2 0520Z/165 at DSDP Site 610. Airgun capacity 160 cu. in. Operating compressed air pressure 1700 p.s.i.

	EPC 1	EPC 2	EPC 3
Time sweep (s)	8	4	8
Frequency Bandpass (Hz)	20-200	40-150	10-50
Time delay (s)	0	2	1
TVG Filter	No	Yes	No
Reflector picks	(seconds Two-Way Time)		
sea floor	3.30	3.30	3.30
R1	3.63	3.61	?3.63
R2	3.90	3.89	c3.97
		to	
	4.12-4.18	4.20	4.12-4.19
?R4	4.44	-	c4.46
R5	5.05	5.03	5.05
?sill	5.26	5.27	5.30
R6 or R7	5.52	No	5.55
?	c5.90	Reflectors	-

Glomar Challenger site survey watergun profile (Masson and Kidd, in press) indicates that the hole was drilled on the south-eastern limb of one of the conspicuous sediment waves in the area; comparison of the bathymetry on this and the C 84-2 records supports the 0520/165 position for site 610 (Fig. 3.14) suggested from the corrected navigation chart (Fig. 3.12 and Chart 1).

Figure 3.15 shows two plots of sonic velocity measured on site 610 samples against depth and two-way time. The stepwise curve was used to calculate the depths of the reflectors picked on line C 84-2 over site 610; the depths of the yellow, purple and green reflectors of Masson and Kidd are also indicated. They projected site 610 90 km to the north-east along the crest of the Feni Ridge to the WI-32 seismic profile under the assumption that the stratigraphy does not vary significantly in that direction. While this is probably the case it seems unlikely that equivalent reflectors at the true and projected sites will occur at the same depths below sea floor. Despite this Masson and Kidd proposed a correlation between their reflectors and downhole velocity increases (Fig. 3.15a). They equated the yellow reflector with a seismic velocity increase between 300 m and 345 m sub-bottom and also an offset in the sedimentation rate curve between 341 m and 415 m. Similarly, the purple reflector was correlated with an increase in velocity between 520 m and 550 m. The low core recovery at both these levels did not allow them to ascertain the lithological cause of the reflectors. The green reflector, however, related to an increase in hardness and velocity due to compaction and silica diagenesis between 625 m and 675 m sub-bottom. The yellow, purple and green reflectors were dated as uppermost Late Miocene, mid-Middle Miocene and uppermost Early Miocene, respectively.

The good correlations of Masson and Kidd (op. cit.) are slightly surprising in view of the distance across which the borehole was projected. It suggests their assumption of constant stratigraphy is a reasonable one, though an element of subjectivity in drawing the velocity-depth curve (Fig. 3.15a) may have enhanced the correlations. In Figure 3.15b the curve has simply been drawn through each velocity analysis (except near total depth where they are poorly clustered) and the C 84-2 picks marked on as broad bands to take account of the ± 0.03 s TWT variation in bathymetry arising from the sediment waves (Fig. 3.14). In this plot the link between the upper reflector and

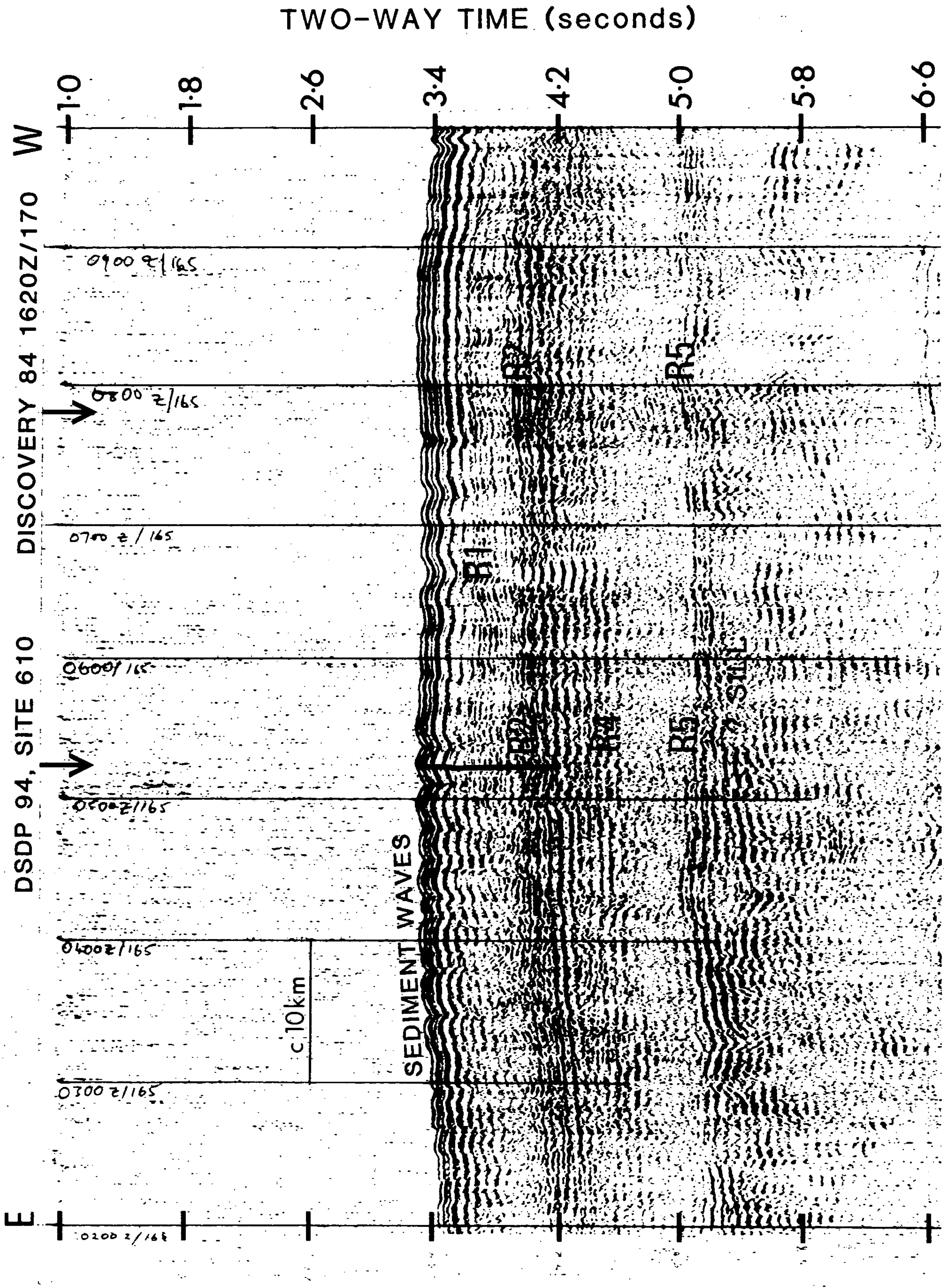


Figure 3.14. Reflector picks between Site 610 and D84 profile

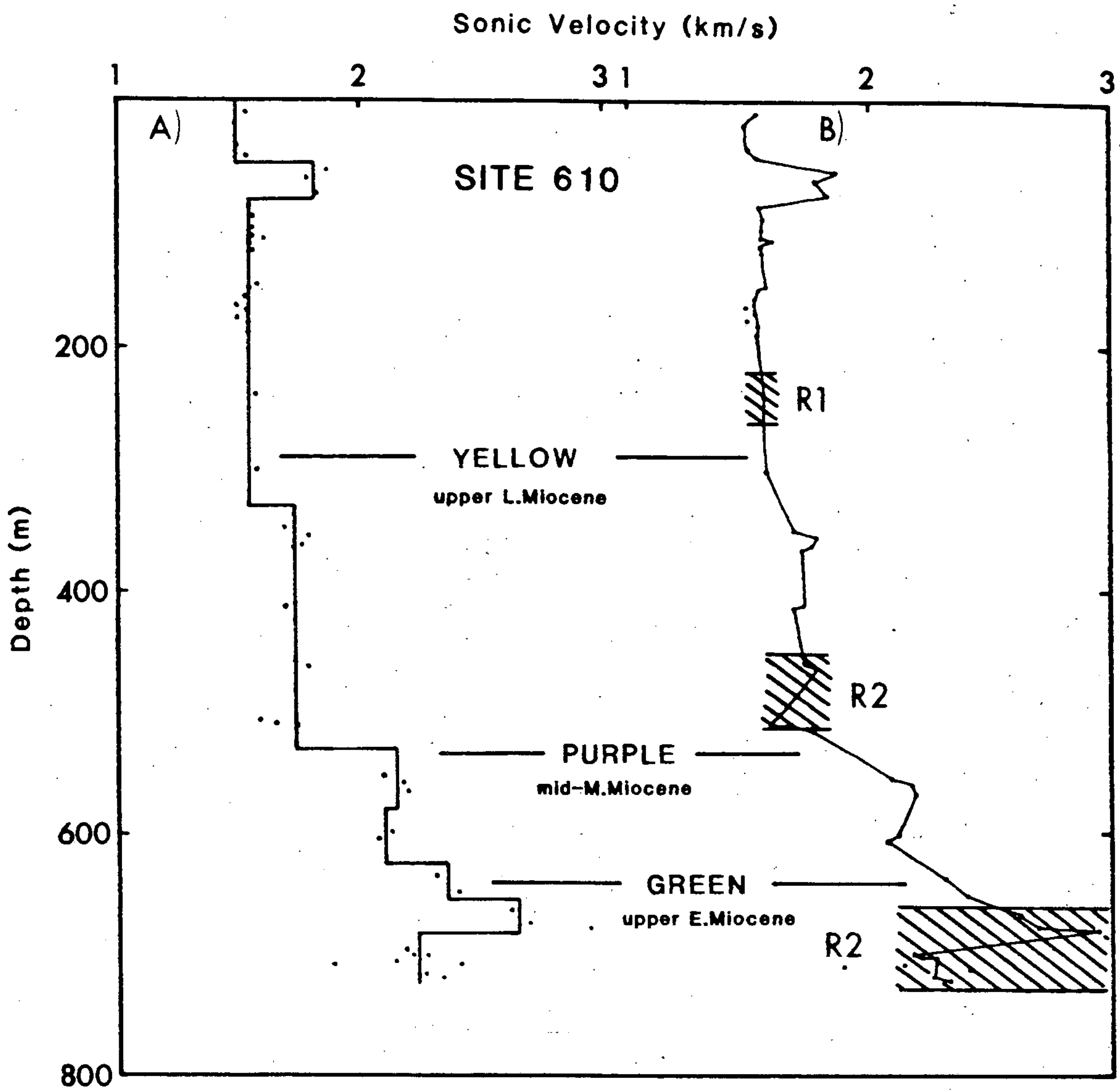


Figure 3.15 A) Stepped sonic velocity versus depth curve for Site 610 (courtesy of D.G.Masson). Position of Yellow, Purple and Green reflectors taken from Masson and Kidd (in press)

B) Continuous sonic velocity versus depth curve for Site 610 constructed by this author. The hatched horizons show the levels of the R1 and R2 reflectors adopted in this study. Refer to text for full discussion.

the velocity rise at c.330 m is tenuous and even the middle reflector appears too shallow. But the lowest reflector has, if anything, an improved correlation with the steep velocity gradient at about 690 m.

The apparent lack of correlation between the upper reflector and stratigraphic and velocity variations at site 610 may be due to a number of factors. Firstly, my pick of this reflector on C 84-2 may be accurate and caused by some event in the Lower Pliocene. Secondly, through navigation errors, lack of exact knowledge of the site 610 location, and poor seismic profile quality, the reflector may have been picked, wrongly, too high (in terms of TWT). And thirdly, the stepped velocity curve of Fig. 3.15a, which is used to convert TWT to true depth, may be too low. The poor velocity control from spot coring below 200 m sub-bottom does not constrain the curve very well; the velocity gradient giving rise to the yellow horizon could comfortably be moved up by 50 m or more. The velocity of roughly 1.55 km/s for most of the interval 0 to 300 m seems rather low. If it was raised by 0.1 to 0.15 km/s the upper reflector (0.31 s TWT) would then occur at approximately 270 m, compatible with the top Miocene age predicted by Masson and Kidd (op. cit.). This would also move the middle reflector into the range of the velocity gradient at about 530 m.

Until the full results of site 610 are published it is difficult to make any definite statements about the lithostratigraphic changes that give rise to the top three reflectors in this area; even then the low core recovery in the interval of interest suggests that any predictions will be highly speculative. Nevertheless there are a number of inferences to be made here. It appears that the correlations established by Masson and Kidd are essentially correct. They ascribed the yellow reflector to reduced sedimentation rates, non-deposition or erosion in the uppermost Late Miocene and equated it with R1 of Miller and Tucholke (1983) (see section 3.2 and Table 3.1). It is difficult to see how an offset in the sedimentation rate curve alone can result in a seismic reflector: the lithification boundary proposed for R1 by Miller and Tucholke seems more tenable, especially since the ooze-chalk transition occurs just below the top of the Miocene.

The middle (purple) and lower (green) reflectors are just two horizons in what is a widespread strongly reflective package. In places this package is thicker than 0.5 s TWT and the velocity structure shown in Figure 3.15 is necessarily a simplification of the true situation. There are numerous other short discontinuous events within the package which seem to relate to compaction and diagenetic effects. Consequently the purple and green horizons are hard to trace laterally on many of the single-channel profiles.

The green reflector, which is here dated as late Early Miocene, is correlated with R2 of Miller and Tucholke (1983), though the widespread unconformity they advocated for R2 is not seen at site 610 (Masson and Kidd, op. cit.). This reflector is also that which Roberts (1975) and Roberts et al. (1981) called R4 in the Rockall Trough and assigned a Late Eocene-Early Oligocene age. This obviously needs to be revised now and the implications for the ages of the deeper reflectors are discussed next.

Seismic continuity between Sites 550 and 610

The picks of seismic reflectors between DSDP sites 550 and 610, a distance of approximately 825 km, are shown in Figure 3.16. The length of the correlation almost certainly means that any reflector will exhibit a degree (indeterminate) of diachronicity; for example, what is picked as the M. Cenomanian to L. Campanian or E. Maastrichtian event beneath site 550 to the south may have a different age 800 km to the north in Rockall Trough.

At site 550 the Eocene-Oligocene (R4) and M. Cenomanian-E. Maastrichtian (R6) unconformities can be traced clearly south-west to the intersection with CM-05. However, less than 5 km beyond this intersection, R4, which is a strong truncating reflector (Fig. 3.11), fades out and a pale crenulated sequence takes its place. This lateral change also occurs on CM-05 roughly between shot points (SP) 4000 and 7000 and consequently the continuity of R4 to the north is not well constrained. Beyond SP 7000 on CM-05 and up to SP 3000 on CM-04 a fairly persistent reflector which occasionally truncates the underlying events, and which usually marks the top of a crinkled semi-transparent sequence, has been equated with R4. In contrast R6 can be followed from site 550 to SP 3000 on CM-04 with a good degree of confidence, though between CM-05 SP 6300 and CM-04 SP 1300 the lower half of the sedimentary section is strongly layered on the profiles and care is required in picking R6. The pre-R6 interval is everywhere noticeably transparent except along this c.140 km section of profile.

North of CM-05 SP 6300, and particularly beyond CM-04 SP 1300, a thin (c.0.2 s TWT) yet strong and persistent group of reflectors is present of which R6 forms the base (see Fig. 4.5). The top of this group has been labelled R5 but little information is available by which to constrain its age. If it corresponds to the strong reflector observed just above the base of sequence B at site 550 then a Maastrichtian age can be proposed. However, tie-ins to two multichannel profiles crossing the south-east slope of Rockall Bank (GSI-1 and NA-1 Ext; see Chapter 5, Figs 5.5 and 5.6) show that R5 and R6 occur, respectively, just above and below a localised strong, irregular reflector concordant with the sediments. This might well correlate with the widespread Palaeocene lavas reported by Smythe et al. (1983), Smythe (1983) and Roberts et al. (1983) around the Faeroe Islands and northern Rockall Trough, and would suggest a younger age for the R5 horizon, perhaps near the Palaeocene-Eocene boundary. Thus the true age of R5 must remain conjectural until improved seismic correlations are made or further drillsites occupied. The younger age is favoured from this work.

Reflector correlations across the Clare Lineament (Figs 3.16 and 3.17) are confused by the abrupt change in bathymetry associated with it. Although this steep sea floor gradient (3-3.5°) does not persist far to the west along the Lineament or the Gibbs Fracture Zone, the changes in the reflector pattern are similar on a number of lines across the discontinuity. All the reflectors step up by 1.0 s TWT or more north of the Clare Lineament, though beneath the region of rough topography between CM-04 SP 3100 and 3200 (c.5 km) the reflectors are heavily disturbed and relationships are not clear. Nevertheless, the transparent sequence below the strong R5 to R6 reflectors is clearly offset upwards, indeed it becomes markedly paler. Hence the continuation of R5/R6 south of the Lineament into R7 north of it and at equal travel time (Fig. 3.16) - a correlation which looks correct on first impressions - is certainly ruled out.

Above R5, along the remainder of the CM-04 profile, the sediments are strongly reflective, not with distinct continuous events, but with irregular crinkled short reflectors. It is difficult to trace any persistent reflectors here: R4 can be picked reasonably well just north of the Clare Lineament but it breaks up and changes character often along CM-04 and can only be traced for a short distance along WI-32 (Fig. 3.16). R4 is lost in a confusion of cuspate reflectors along much of line CM-03 but it may reappear between 4.5 and 5.0 s TWT towards the north-eastern end of that profile (Fig. 3.17). The R4 to R5 interval from the Clare Lineament to CM-04 SP 4600, where it nearly pinches out, is more uniform than the interval above and has a hint of transparency, compatible with its seismic character further south. R4 cannot be picked beyond SP 2900 on WI-32 or on Discovery 84 and Challenger 84-2, that is, beneath the Feni Ridge. There is a good indication that R4 is only ever a good reflector where the influence of bathymetric gradients has been strong. For example, below those sections of CM-11, 5 and 4 close to the continental slope R4 is seen to be a strong truncating reflector. Moving away from the slope R4 passes into a widespread sequence of short crinkled events lacking any coherent structure. Similarly, R4 can be discerned around the Clare Lineament (which is here close to the base of Porcupine Bank) but it is gradually lost as the correlation progresses towards the axis of the Trough.

The upper three reflectors of profile WI-32 over Feni Ridge (Fig. 3.16) are those of Masson and Kidd (in press). R1 cannot be resolved from the bubble pulse reverberations on the 8-second Discovery 84 profile; it is seen clearly however at about 0.3 s TWT sub-bottom on the C 84-2 4-second profile (Fig. 3.13). R2 and the middle reflector are not usually separable on 8-second single- or double-channel seismic profiles and so will be treated here together as R2. The event picked at this level on these profiles can be either of the two reflectors but it almost invariably marks the top of a strongly reflective package which frequently attains a thickness of about 0.5 s TWT.

Roberts (1975) and Roberts et al. (1981) both showed R4 in the position of R2 of Figure 3.16. Since R4 has always been dated at the Eocene-Oligocene boundary, and since R2 has been drilled as a late Early Miocene event at site 610, this correlation is no longer reasonable. Masson and Kidd (op. cit.) revised the Eocene-Oligocene reflector to be R5 of this work which occurs at the base of a transparent sequence beneath Feni Ridge (Fig. 3.14). This is difficult to reconcile both with the correlation from site 550, which indicates that R4 lies somewhere between R2 and R5, and with the large thickness of Oligocene and Early Miocene that would be implied in such a hypothesis. The borehole data from Rockall Plateau, Goban Spur and Biscay (section 3.2) show that the Oligocene was a period of intermittent erosion and non-deposition; whether this manifests itself in the deeper Rockall Trough will remain in doubt until it has been drilled. Furthermore, this author's pick of R4 is in keeping with its seismic character in the vicinity of the DSDP Leg 12 cruise (sites 116 and 117). There is no evidence for the Late Oligocene R3 event of Miller and Tucholke (1983), hence its absence from Figures 3.16 and 3.17.

The deeper reflectors have been labelled as an extension of the R1 to R4 scheme set up by Miller and Tucholke (1983). It is not possible to adhere to the X, Y, Z nomenclature of Roberts (1975) and Roberts et al. (1981) owing to their inconsistent interpretations; their X is variously R4 or R6, Y is R5, R6 or R7, and Z is R7 or sub-R7 of this work. From the seismic descriptions given by Roberts et al. (1981) it is evident that their Y horizon should everywhere be the equivalent of R7 of this study and Shackleton of Dingle et al. (1982). Unfortunately, however, they dated Y where they had incorrectly associated it with supposed Early Eocene lavas. The picks appear to be wrong, the lavas are seemingly mostly sills, and therefore maintaining the reflector as Y cannot be recommended.

In fact it is the R6 horizon that is commonly associated with igneous sills in this area. R6 is largely obscured by sills beneath WI-32 (Fig. 3.16); likewise sills occur at or just below R6 at the end of CM-04 and the beginning of CM-03 (Fig. 3.17), and also a little way south of the Clare Lineament. This consistent appearance of sills supports the picks of R6 in the present work. If the sills are coeval with the extensive Thulean igneous activity (mainly Palaeocene-Eocene) then independent evidence is provided for the Late Campanian or E. Maastrichtian age mooted for R6. The fact that these sills are never observed above R5 means that it (R5) is unlikely to be older than the start of the Eocene. Ridd (1983) describes very similar relationships from the Faeroe-Shetland Channel and comparisons with his figures also indicate an approximate base Eocene age for R5.

The R7 reflector is a strong persistent event that defines the top of a highly reflective sequence which infills the basement topography and which is noticeably absent over the oceanic crust south of the Clare Lineament-Gibbs Fracture Zone. This is well illustrated in Figure 3.17. R7 and the numerous deeper reflectors almost invariably pinch out against basement highs in the Trough. There are in addition, internal unconformities within the pre-R7 sequence (particularly Z of Roberts et al. 1981). If R7 is coeval with the formation of Middle Albian oceanic crust drilled near the continent-ocean boundary west of Goban Spur, as seems possible, then an approximate M. Albian age for the reflector can be proposed. This idea is reinforced by the roughly equal thicknesses of the transparent sequence below R6 seen on both sides of the Clare Lineament (Figs 3.16 and 3.17), the suggestion being that R7 formed at about the same time as seismic basement south of CM-04 SP 3000. If this is so then R7 is in essence a post-rift (drift) unconformity marking the onset of oceanic accretion west of the British Isles. This has important implications for the ages of both the pre-R7 sediments, which are fairly thick in places, and also the basement flooring Rockall Trough: these topics will be discussed in Chapters 4, 5 and 8. The seismic stratigraphy presented in this section is summarised below in Table 3.2.

Table 3.2. Seismic stratigraphic scheme used in the present study. Refer to text for discussion.

R1: latest Miocene
R2: late Early Miocene - Middle Miocene
R3: not identified in this study; Late Oligocene of Miller and Tucholke (1983).
R4: Late Eocene - Early Oligocene
R5: earliest Eocene
R6: ?Late Campanian - Early Maastrichtian
R7: ?Middle - Late Albian

4. THE ROCKALL OFFSET MARGIN AND THE CHARLIE-GIBBS FRACTURE ZONE

4.1 Introduction: bathymetry and early studies

The eastward extension of the Charlie-Gibbs transform fault -which sinistrally offsets the median rift of the North Atlantic ridge by some 350 km - into the abyssal region immediately south of Rockall Trough and west of Porcupine Bank has been known for over a decade (Fig. 4.1; Johnson 1967; Fleming et al. 1970). It results in well documented Cretaceous oceanic crust west of Porcupine Bank being juxtaposed against geological basement of, as yet, unknown affinities in the Trough to the north. If this unknown basement comprises oceanic crust then the label "offset margin" is not strictly true; by definition it should only apply where continental and oceanic basement are contiguous across a sheared continental margin, such as off the southern coast of West Africa. Nevertheless, for the purposes of this discussion the term will be retained as it emphasises the widespread continuation of continental crust to the west below Rockall Plateau. In addition it enables comparison with the similar sized offset margin observed south of Edoras Bank (Fig. 4.1).

The continuation of the Charlie-Gibbs transform fault as a major fracture zone approaching the distant continental margins was reported by Cherkis et al. (1973) and Vogt and Avery (1974). From a detailed bathymetric and magnetic investigation of the eastern basin of the North Atlantic Vogt and Avery detected a major discontinuity as far east as 17°W (their Fig. 2b) which they proposed to be the inactive extension of the transform. Although subsequent workers have made refinements to the position and trend of parts of the fracture zone the fundamental geometry established by Vogt and Avery (1974) is essentially correct. Detailed studies of the active, dextral transform section of the fracture zone, and its intersections with the Mid-Atlantic Ridge, have been presented by Lonsdale and Shor (1979) and Searle (1981). More recently Smoot and Sharman (1985) reported a previously undescribed long narrow ridge within the fracture zone near 52°N 26°W. They ascribed the ridge to some kind of igneous intrusive feature, possibly a serpentinized ultramafic diapir, and equated it with a similar basement high documented further west by Searle (1981) who also expounded an intrusive origin.

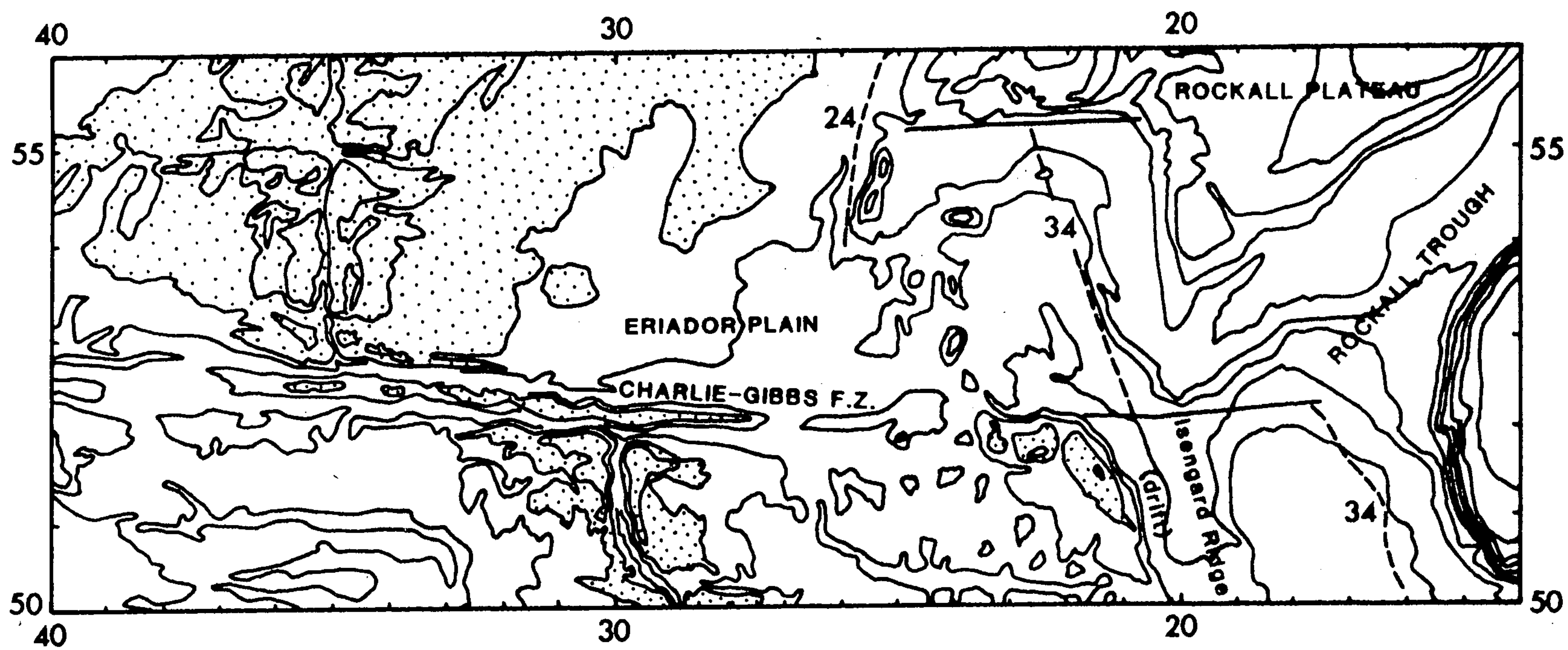


Figure 4.1 Bathymetric map of the NE Atlantic west of the British Isles. 500m to 4500m isobaths at 500m intervals. Stippled regions shallower than 3000m over the Mid-Atlantic and Reykjanes Ridges. Dashed lines are magnetic anomalies 24 and 34, the latter offset by the Charlie-Gibbs Fracture Zone at the mouth of Rockall Trough.

The Charlie-Gibbs F.Z. is the only major fracture zone in the North Atlantic and consequently it has been an important controlling factor throughout the development of the ocean. For this reason the Charlie-Gibbs F.Z. has been the focus of attention in a number of regional geophysical investigations that have been directed at elucidating the opening history of the North Atlantic basins (Vogt and Avery, 1974; Roberts, 1974; Le Pichon et al., 1977; Kristoffersen, 1978; Srivastava, 1978; Roberts et al., 1981; and Hanisch 1984). In particular these studies have concentrated on unravelling the evolutionary story contained within the ocean floor magnetic lineations.

Along the seismically active section of the Charlie-Gibbs Fracture Zone (CGFZ) two conspicuous linear troughs can be seen which are separated by a prominent median ridge (Fig. 4.2; Fleming et al. 1970; Searle, 1981). This distinctive trough-ridge-trough pattern can be followed along a trend of 095° to about 25°W where it continues in a more broken, blocky fashion to 22°W. Beyond this point the feature is obscured by the Isengard Ridge sediment drift, while further to the east, in the mouth of Rockall Trough, the discontinuity has little or no bathymetric expression but can be traced on seismic reflection profiles and inferred from the magnetic and, in part, gravity anomaly signatures.

The characteristic relief of the CGFZ in the deep ocean arises from the presence of two closely spaced fracture zones, rather than just one. A short active spreading segment has been detected at 31.75°W (Searle 1981; Lilwall and Kirk 1985) which is connected by transform faults to the main N. Atlantic median rift at approximately 30°W and 34.75°W (Fig. 4.2). The length of the short spreading centre, and therefore the separation of the transform valleys, is 45-50 km. Despite the limited length of the rift the available evidence suggests that the median ridge between the two transforms has been formed by the normal oceanic accretion process (Searle 1981), and not by one of ultramafic igneous intrusion (Olivet et al. 1974).

The identification of the short spreading centre and of the double fracture zone nature of the CGFZ has an important influence on the recognition and interpretation of the tectonic structures and trends in and around southern Rockall Trough, in particular the major discontinuity called the Clare Lineament as described by Dingle et

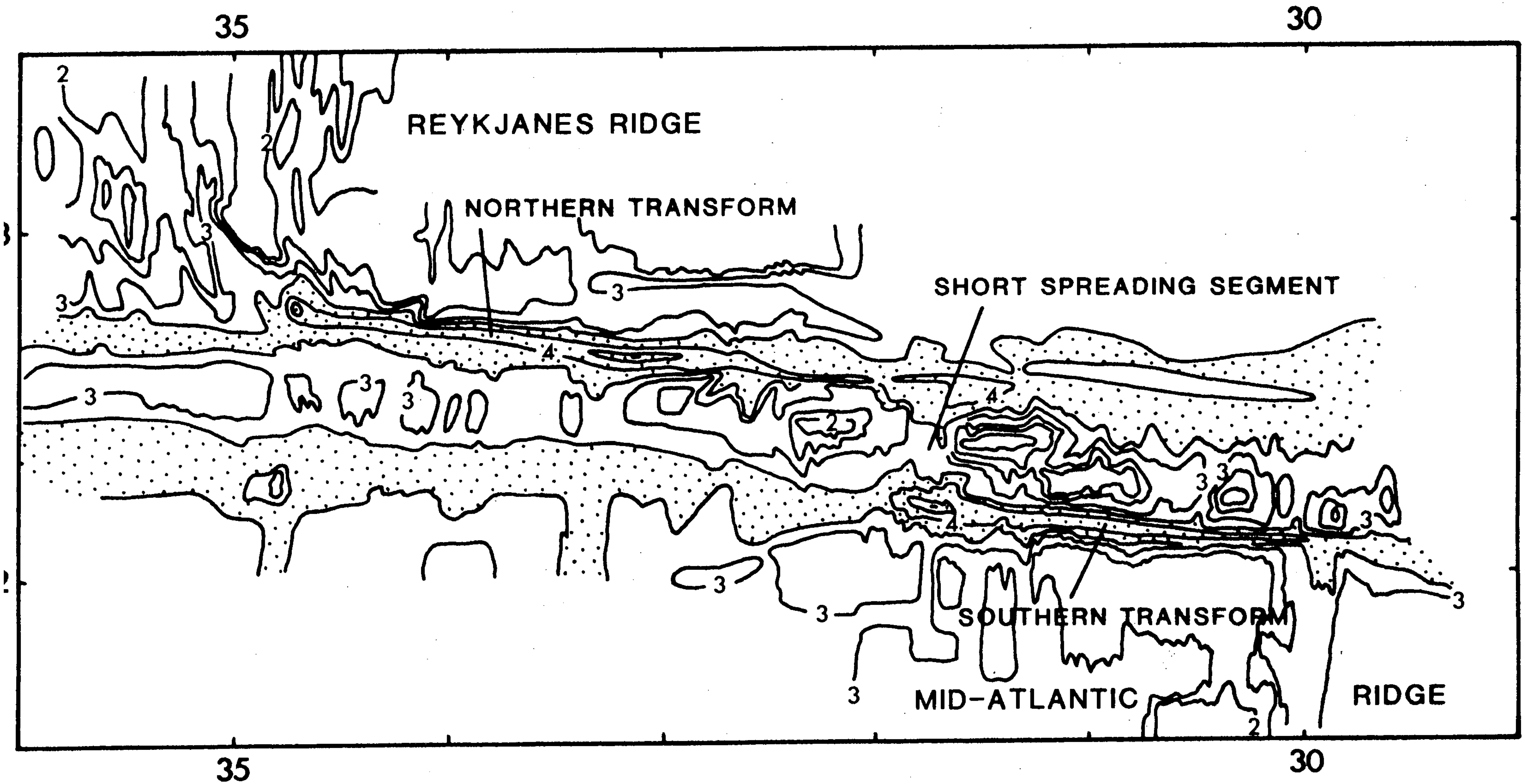


Figure 4.2 Detailed bathymetry of the Charlie-Gibbs Fracture Zone at its intersection with the Mid-Atlantic and Reykjanes Ridges. Spreading direction is roughly east - west. Isobaths drawn every 500m and labelled every 1 km. Stippled regions highlight areas deeper than 3500 m. Modified from Searle (1981; Plate 2).

al. (1982) and Megson (1983). Like previous workers both these studies show the CGFZ terminating between 17° and 18°W, with the Clare Lineament continuing as an additional, separate feature with a more south-easterly trend (see Fig. 4.6). Essentially, however, they appear to be part of the same discontinuity, hence proving the persistence of the CGFZ from the Mid-Atlantic Ridge to the continental margin west of Ireland. The evidence from the conjugate Canadian margin will be considered in the final chapter.

The detailed bathymetry of the southern Rockall Trough is shown in Figure 4.3. Over much of Porcupine Abyssal Plain the isobaths follow the shape of the continental slope. But at 51.5°-52°N, 15.5°-16°W a short, steep east-west gradient is prominent in the sea floor at the base of Porcupine Bank. Where it intersects the continental slope a distinct SW-NE trending trough is present. Some authors (e.g. Roberts 1975; Dingle et al. 1982) have advocated a correlation between the eastern termination of the CGFZ and this conspicuous channel. To the west the gradient soon becomes more subdued but a rough east-west to SE-NW trend is discernible from the isobaths which reflects a similar trend in the buried CGFZ. The area of interest here (up to 20°W) is bounded to the west by the masking effects of the Feni Ridge, and its southern extension, the Isengard Ridge (Roberts, Hunter and Laughton 1979). Thus it is evident that the CGFZ, which is the major discontinuity marking the southern tectonic limit of the Rockall Trough, has no well-defined bathymetric expression in this area. Instead there is a steady northward shallowing of the sea bed that is a consequence of the change from a thin sediment cover above oceanic crust south of the CGFZ to a thick sedimentary sequence in the Rockall Trough.

4.2 Geophysical investigations across the Rockall offset margin

4.2.1 Basement structure and sediment geometry as shown by seismic reflection profiles

The distribution and seismic character of the basement and overlying sedimentary sequence associated with the Charlie-Gibbs Fracture Zone and Clare Lineament were examined on 14 reflection profiles that traverse these discontinuities. The locations of these tracks are shown in Figure 4.3 and line drawing interpretations of

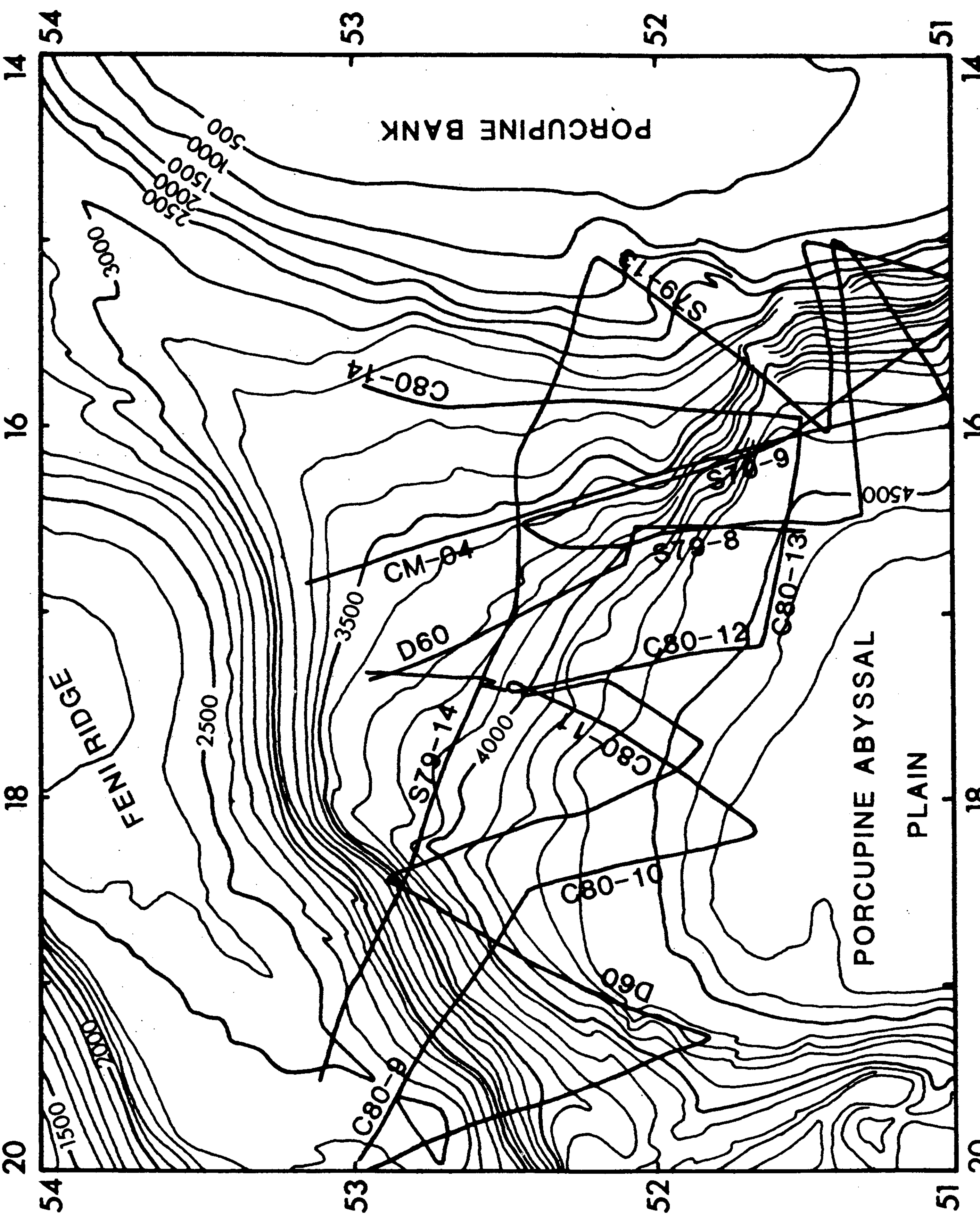


Figure 4.3 Seismic reflection profile coverage over the buried extension of the Charlie-Gibbs Fracture Zone, southern Rockall Trough. See also Chart 1 for more detailed navigation data. Bold isobaths every 500m. Fainter isobaths every 100 m except below the very steep Porcupine Bank continental slope.

nine of the double-channel seismic profiles are given in Figure 4.4. The four Discovery cruise 60 profiles are of poor quality so are not reproduced here, though they do resolve some of the seismic basement highs. An interpretation of the relevant section of the multichannel CM-04 record, in part reprocessed and migrated, appears in Figure 4.5.

Basement structure and distribution

The 14 seismic profiles mentioned above, together with others in the surrounding areas, were used to construct an acoustic basement isochron map (Fig. 4.6; Chart 4 in back pocket) of the buried extension of the Charlie-Gibbs Fracture Zone, i.e. one showing contours of two-way travel time to seismic basement. Similar charts have been produced by Roberts et al. (1981), Dingle et al. (1982) and Megson (1983) which have highlighted similar regional trends. The combined data from these three sources, together with a small amount of new profiles (Challenger 1/84 cruise, section 2.2), were used to draw up Figure 4.6 which should therefore represent a good approximation to the true pattern. Certainly in previous studies errors in contouring have arisen through the mis-identification or mis-representation of intrusive sills within the sediments as seismic basement. Strictly speaking these sills frequently appear as seismic basement (the deepest recognisable reflector), but including them in the contouring scheme distorts the information concerning the deeper more substantial seismic basement - the material that usually corresponds to true geological basement. Because of the absence of good velocity control along the seismic profiles no attempt was made to convert Figure 4.6 from acoustic travel time to true depth; a factor unlikely to affect the dominant trends in this region.

Whereas the extension of the CGFZ to about 17°W at 52°N has been known for some time (Cherkis et al. 1973; Vogt and Avery, 1974; Roberts et al., 1981) the presence of the Clare Lineament was first reported by Dingle et al. (1982) and described in fuller detail by Megson (1983). Both these studies maintained the Clare Lineament as a feature separated and distinct from the CGFZ and considered it to have a NW-SE orientation. Megson (1983) believed that the Lineament manifested itself over the southern Porcupine Bank as a magnetic anomaly low, the Clare Trend, oriented to the south-east. The contour chart of the present work (Fig. 4.6), while substantiating the existence of the Lineament, is partly in disagreement with these previous studies. It shows, at 51.65°N 16°W, two conspicuous basement ridges; the northern ridge has a trend of roughly 110°, the southern ridge and the accompanying linear gradient immediately south are orientated between 100° and 105°. These two highs can be traced west to seismic profile S 79-8 (Figs. 4.3 and 4.4e) where, despite being close together, they are clearly defined; though they are not well picked out by the 7.0 s TWT basement isochron. The northern ridge appears to curve round to point WSW between profiles S 79-9 and S 79-8.

To the north of the Clare Lineament a narrow basin is present that exceeds 7.5 s TWT in depth (Fig. 4.6). This trough opens out to the west into the CGFZ proper and its northern flank is marked by a strong basement gradient G that trends 100°-105°, identical to the southern ridge and slope of the Clare Lineament. These roughly east-west oriented highs and lows of the Clare Lineament are terminated abruptly at 15.7°W by the north-south isochrons of the continental slope of Porcupine Bank.

To the west of seismic profile S 79-8 there is a 50 km wide gap in the Edinburgh data in which the trends and continuity of the seismic basement are poorly constrained. A low resolution single-channel seismic profile shot in this area was discussed by Cherkis et al. (1973). The profile runs approximately north-south along 16.7°W (Fig. 4.6) and inspection of a photographic reduction of this seismic line reveals a strongly reflective broad ridge at about 6.8 s TWT, to the west of the double-ridged basement high observed along S 79-8, C 80-14 and CM-04 (Figs 4.4 and 4.5). Evidently then, the pronounced sea floor and seismic basement topography seen near 16°W becomes progressively more subdued over the 50 km to the west.

The Charlie-Gibbs Fracture Zone from 20°W to 17°W is characterised by two narrow, high linear ridges that are nearly parallel and are 35 to 50 km apart (Fig. 4.6). Between 20°W and 18°W the two ridges are developed along a 90-95° direction. But east of 18°W their bearing increases to 100°, or slightly more. This is more noticeable in the southerly ridge (S) which appears to project into the trend of the Clare Lineament. The wide spacing of the seismic profiles between 16.5°W and 17.5°W makes it impossible to link unequivocally this ridge, where it is well developed on profile C

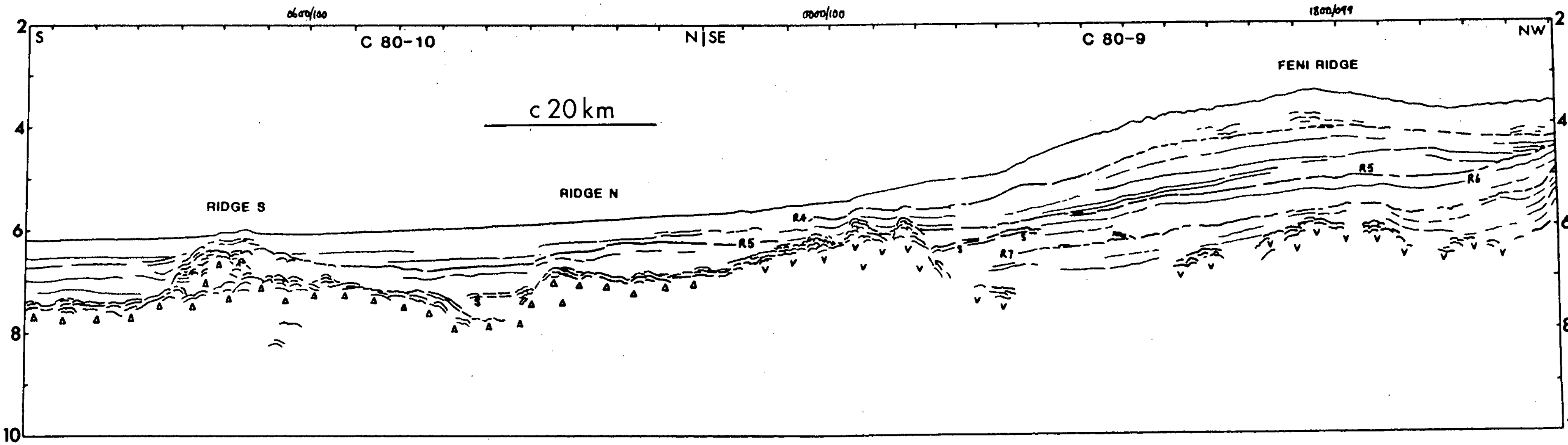


Figure 4.4a. Interpretation of combined seismic profiles C 80-10 and C 80-9. See Fig. 4.3 and Chart 1 for locations. Depth in seconds two-way time. Time marks every ½ hour along horizontal scale. Ridge S and Ridge N define outer limits of Charlie-Gibbs Fracture Zone. Main reflectors appear as bolder lines. Lighter lines indicate geometry within seismic sequences. Sedimentary reflectors labelled R2 to R7 where known. V symbol = volcanic seismic basement. Triangle symbol = oceanic crustal basement. S symbol denotes high amplitude reflectors representing sills or ?lavas.

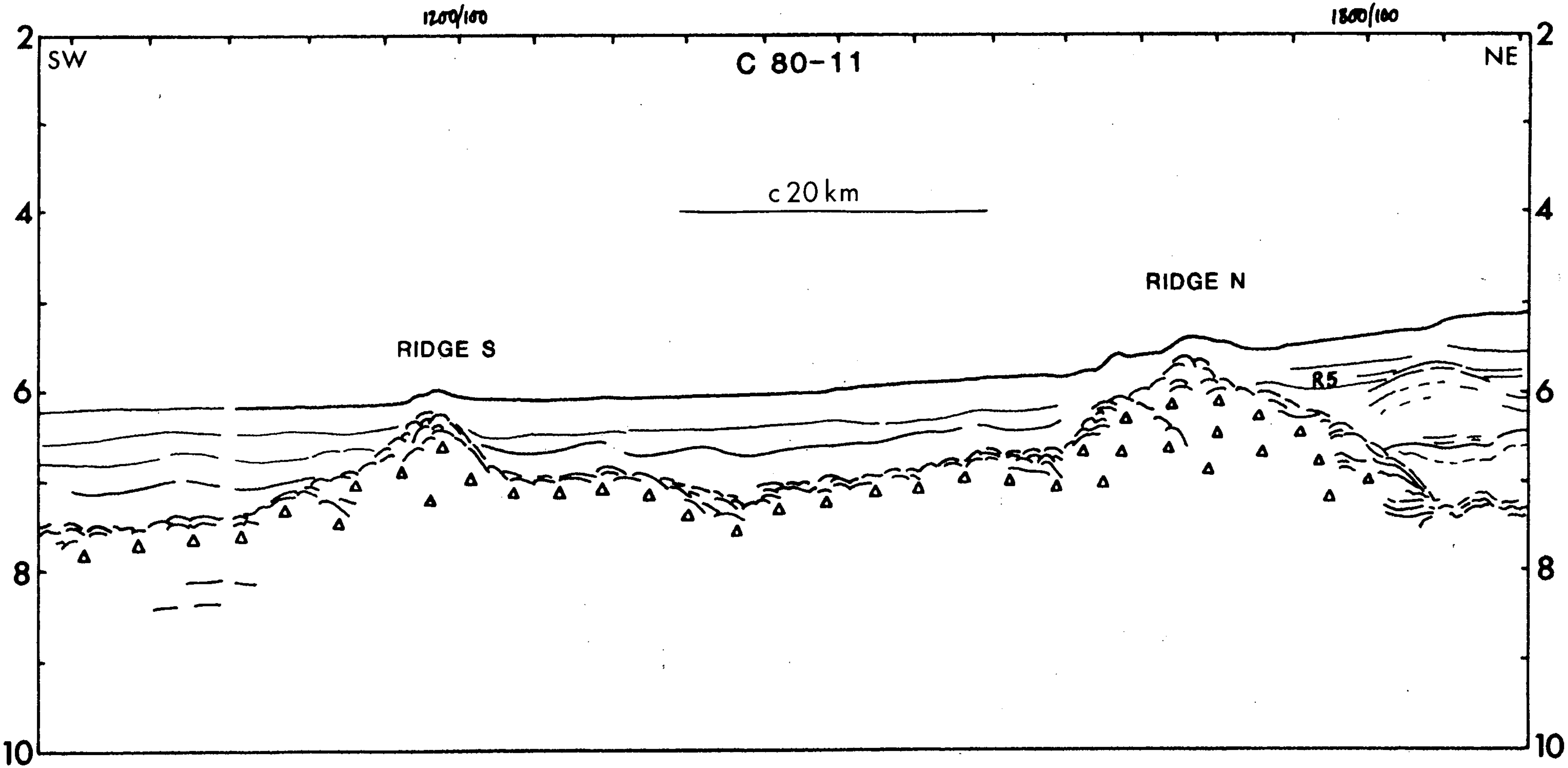


Figure 4.4b. Interpretation of seismic profile C 80-11. Ornament, etc. as for Fig. 4.4a.

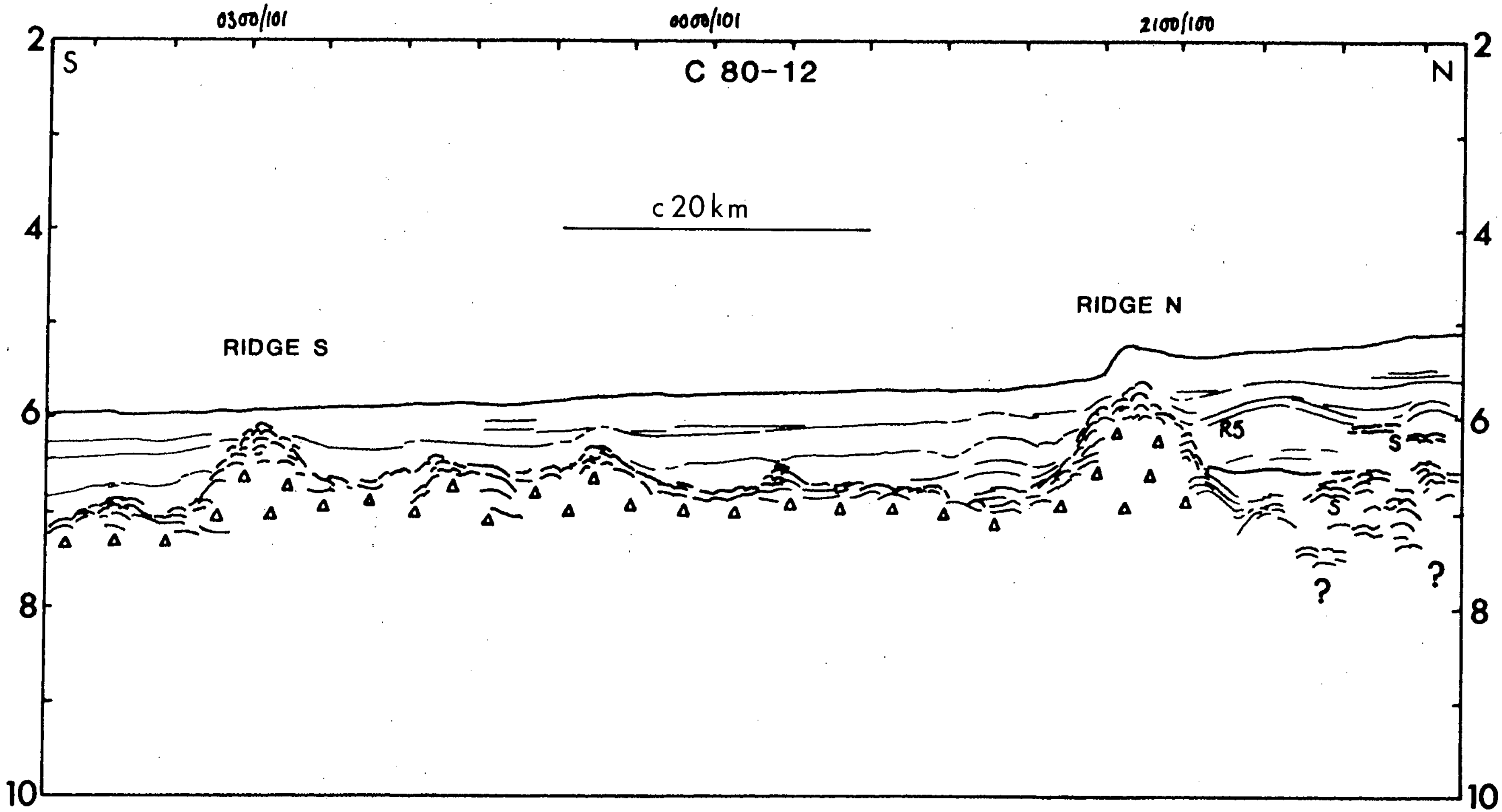


Figure 4.4c. Interpretation of seismic profile C 80-12. Ornament, etc. as for Fig. 4.4a.

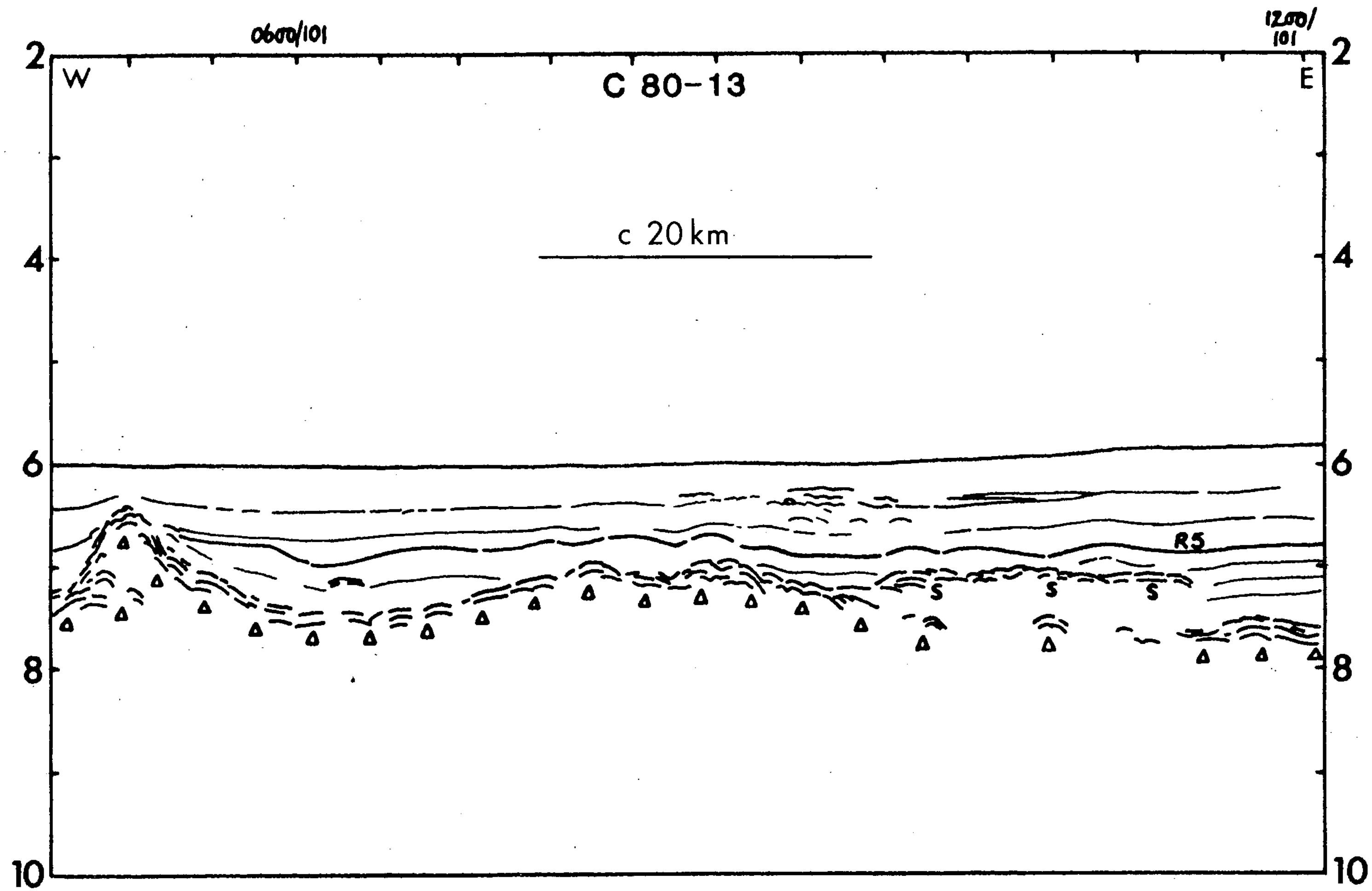


Figure 4.4d. Interpretation of seismic profile C 80-13. Ornament, etc. as for Figure 4.4a.

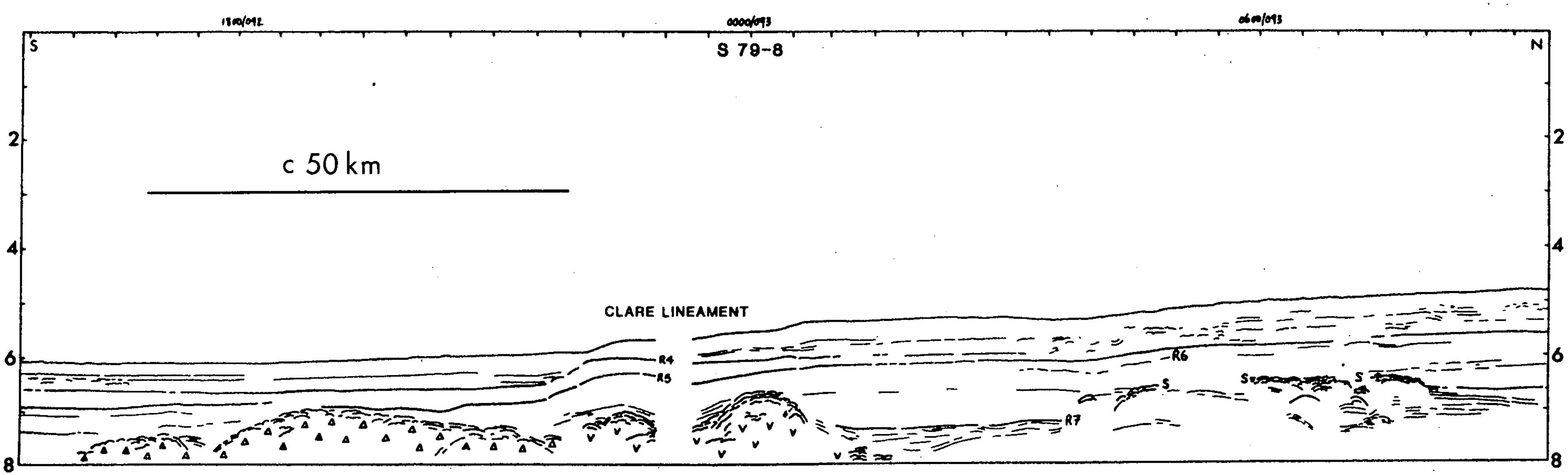


Figure 4.4e. Interpretation of seismic profile S 79-8. Ornament, etc, as for Fig. 4.4a.

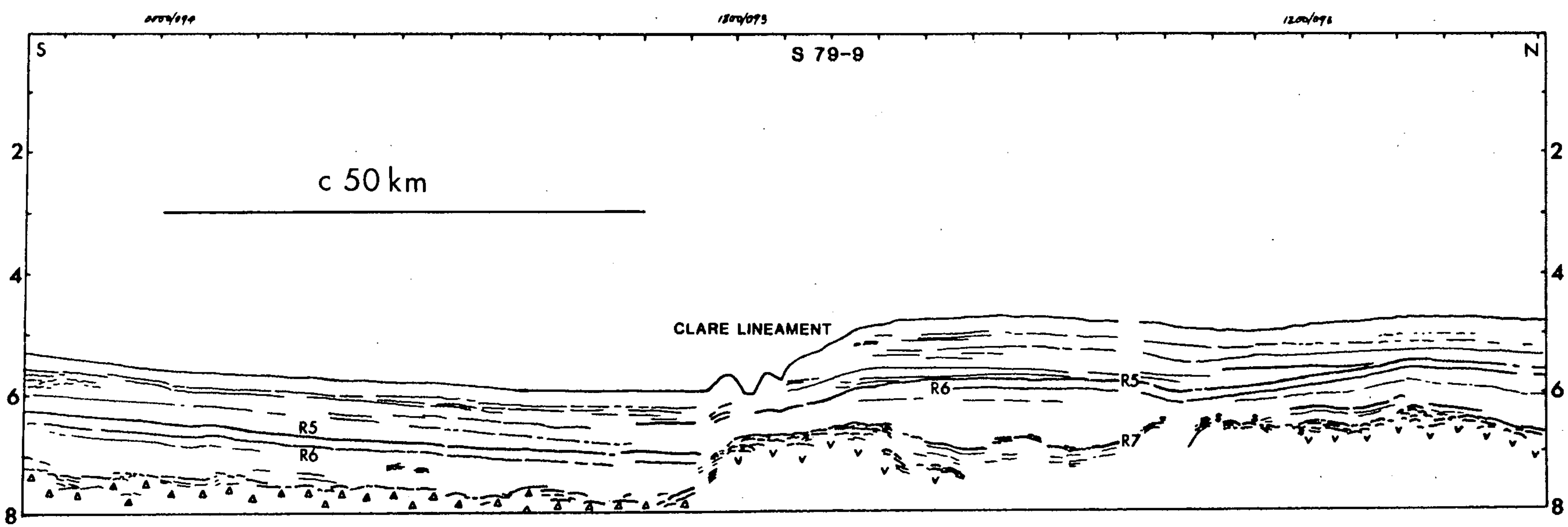


Figure 4.4f. Interpretation of seismic profile S 79-9. Ornament, etc. as for Fig. 4.4a.

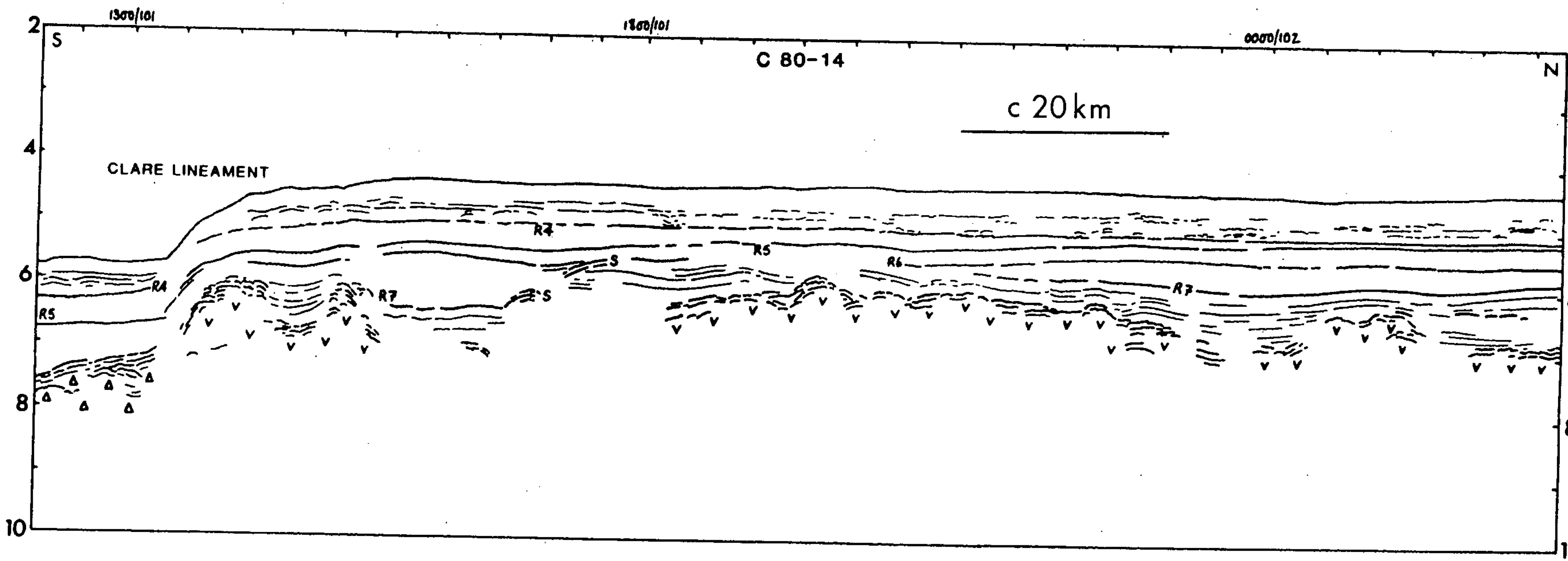
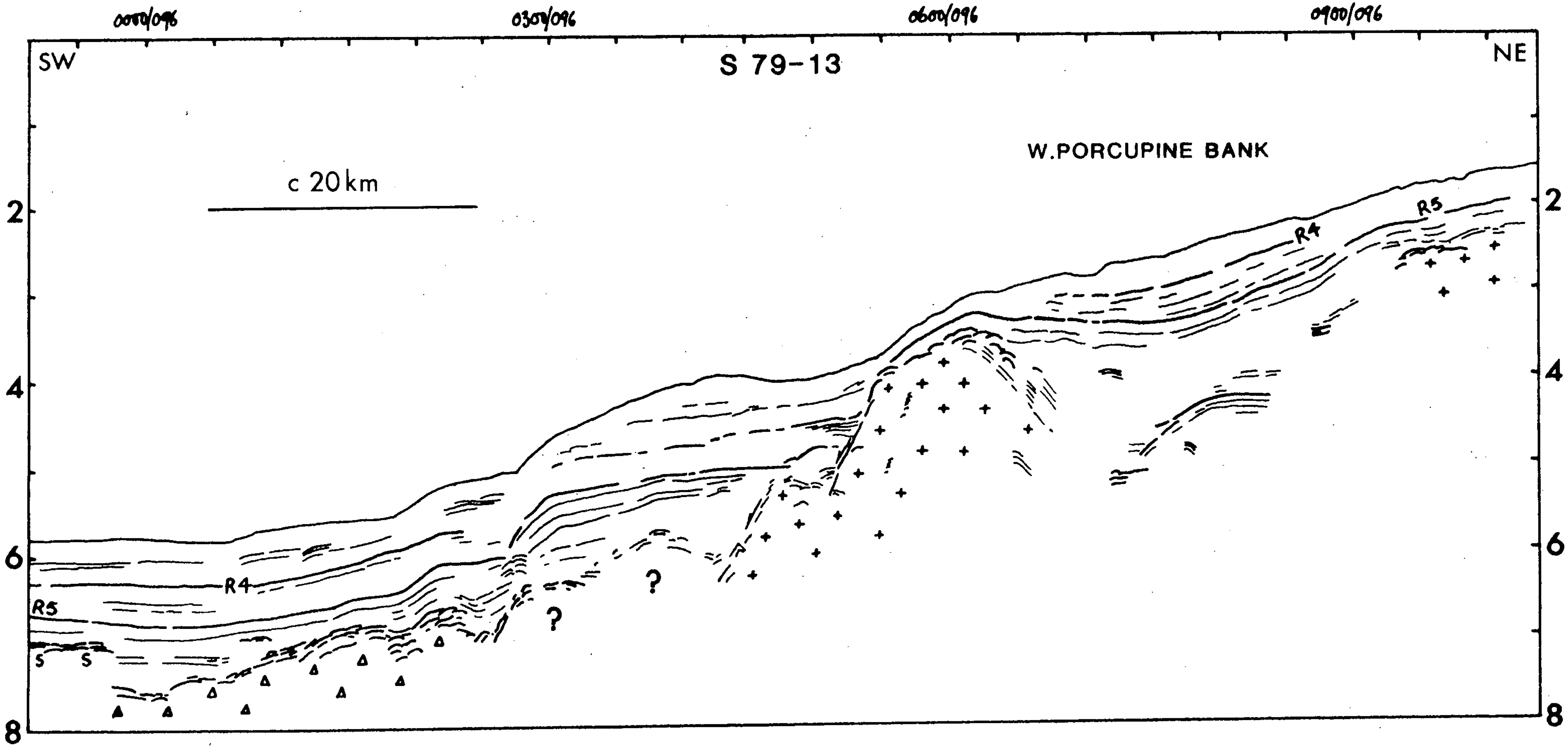


Figure 4.4g. Interpretation of seismic profile C 80-14. Ornament, etc. as for Fig. 4.4a.

Figure 4.4h. Interpretation of seismic profile S 79-13. Ornament, etc. as for Fig. 4.4a. Crosses denote continental basement. ? = presumed continent-ocean transition.



80-12 (Fig. 4.4c), to that of the Clare Lineament on S 79-8, and also on the profile of Cherkis et al. (1973) a mere 30 km distant. Although the continuity of the two ridge systems seems to be an attractive proposition the evidence from the seismic reflection profiles is conflicting. The six crossings of ridge S (Fig. 4.3; three are shown in Fig. 4.4) all show it taking on a distinct triangular peaked appearance; in contrast the basement high at the Clare Lineament is noticeably flatter, and where the ridges are present they are quite rounded and low in comparison (Figs 4.4f and 4.5). Perhaps, instead, ridge S is related to the development of the steep basement gradient and abrupt discontinuity or scarp marking the southern limit of the Clare Lineament. Indeed, this is the most likely explanation, one which will be pursued in later discussions.

The bearing of 90-95° measured for ridges S and N (Fig. 4.6) is identical to that reported for the Charlie-Gibbs Fracture Zone by Cherkis et al. (1973), Vogt and Avery (1974) and Searle (1981) between the Mid-Atlantic Ridge and this research area. It is worth restating here that along the active, transform section of the CGFZ the bathymetry is dominated by two deep troughs on either side of a central ridge (Searle 1981). This pattern of relief progressively diminishes eastwards away from the transform zone termination at 30°W. But there is little evidence in these areas for the two ridge appearance of the CGFZ east of 20°W. The 85 n.m. long but narrow ridge reported by Smoot and Sharman (1985) between 26°W and 27°W, marking the northern edge of the fracture zone, probably represents the younger equivalent of ridge N in Figure 4.6. There is no similar correlative for ridge S west of 20°W. This peculiar change in bathymetric and basement expression has yet to be explained.

In the same way that ridge S projects into the south-facing scarp of the Clare Lineament (Fig. 4.6), ridge N is co-linear with the similar basement gradient G at 52.2°N. However, as with ridge S, the absence of good quality seismic reflection data precludes the accurate mapping of the eastern extension of ridge N. Nevertheless the geometry of the basement scarps and ridges is strongly in favour of their formation by the same mechanism. If ridge S developed through control from the Clare Escarpment it seems likely that ridge N is somehow related to the scarp at G.

The linear ridges and basins of the CGFZ and Clare Lineament, as outlined above, provide a 50 km broad divide between the controversial basement in Rockall Trough and the oceanic crust west of Porcupine Bank and Goban Spur. Immediately north of the CGFZ the trends of the seismic basement structures show a marked change. A number of ridges with a relief of about 1 s TWT strike between SW-NE and SSW-NNE (Fig. 4.6). Two of these ridges continue southwards to join up with ridge N of the CGFZ at 17.5°W and 18.7°W. Although the seismic reflection control is wanting there is a strong indication that the first of these ridges, and perhaps also the second, curves around rapidly to the west near its junction with the Fracture Zone; that is, the intersection between these features is not perpendicular and abrupt but tangential and smooth. The implications of this are treated in later chapters. To the south of the S ridge and Clare Lineament seismic coverage is poor (Figs 4.3 and 4.6). Immediately west of Porcupine Bank a small number of short linear or lensoid peaks and troughs can be defined with a relief of 0.5 s TWT or more which are related to strike fault-blocks on oceanic crust. The presence of shallow acoustic basement beneath the continental slope just north and south of the CGFZ - Clare Lineament is clearly marked by the sudden rise and close spacing of the isochrons in the vicinity of 15.5°W.

The seismic character of oceanic crust in and south of the CGFZ is well illustrated in Figures 4.4 and 4.5. The top of oceanic layer 2 on the profiles is everywhere highly reflective and irregular. It is unusual to observe persistent reflectors below this dense, layered basement. Where they are seen they are not easily accounted for. They may represent deeper volcanic layering or the boundary between extrusive and intrusive material, i.e. the top of layer 2C (sheeted dykes). But most likely they are reflections from oblique crossings of steep normal faults in oceanic crust. Good examples of this phenomenon have been recorded by Masson et al. (1985) west of Goban Spur.

The persistent shape and structural style of the S and N ridges of the CGFZ is exemplified by profiles C 80-10, C 80-11 and C 80-12 (Figs 4.4a-c). The irregularly-layered, steep-sided triangular ridges are easily visible on the Discovery 60 crossings of the Fracture Zone (Fig. 4.3, profiles not illustrated). Between the two ridges the oceanic basement is hummocky but at a fairly constant depth - a level that is slightly elevated compared with the flattish oceanic basement south of ridge S. There is evidence for minor ridges on C 80-12, while at 0400/100 on C 80-10 (Fig. 4.4a) the shape of the reflectors suggests the presence of a sizeable basin nearly 1 s TWT deep. Megson (1983) also noted this feature. However the sill picked here at c.7.3 s may in fact represent the top of oceanic basement, in which case the basin is considerably smaller. The interpretation of the reflector as a sill is favoured because of the close comparisons to the longer slightly shallower sill at 2200/099 on the same profile. This sill appears to show the same relationships to the adjacent basement and sediments.

Further to the east the oceanic basement exhibits identical properties. On profile C 80-13 (Fig. 4.4d) south of the CGFZ and parallel to the early spreading direction, the layer 2 basement is again highly irregular and absorbs most of the acoustic energy. The 1 s TWT tall ridge at 0500/101 has the same appearance as those of the CGFZ but is presumably parallel to the strike of the contemporary mid-ocean ridge. The long (c.20 km) igneous sill at 7.0 s TWT between 0930 and 1130/101 is also seen clearly on the intersecting CM-04 profile (Figs 3.16 and 3.17). South of the Clare Lineament the oceanic basement is generally flat-lying and made up of numerous strong, short reflectors. Examples of broad shallow ridges (S 79-8, 1900/092) and faulted scarps (CM-04, SP 2900) can be found. On seismic profile S 79-13, which traverses the continent-ocean transition near the eastern end of the Clare Lineament, there is a marked difference between the highly reflective, irregular oceanic basement, which rises from 7.5 s TWT near the start of the record to almost 6.0 s TWT at 0300/096, and the shallower rotated continental fault blocks to the east (Fig. 4.4h). The true geometry of the transition between these two geological provinces is obscured by the sedimentary basin present from 0300 to 0500/096.

In a similar manner the relationships between ridge N of the CGFZ and the seismic basement to the north, be it oceanic or continental in composition, are difficult to discern. On profiles C 80-9 and -10 (Fig. 4.4a) the situation is complicated by the intersection of ridge N with a second ridge trending SW to NE (Fig. 4.6). On the seismic profiles there is little observable difference between the two. The oblique ridge has a more irregular, peaked appearance and descends rapidly below the sill at 2230/099 on C 80-9: but there

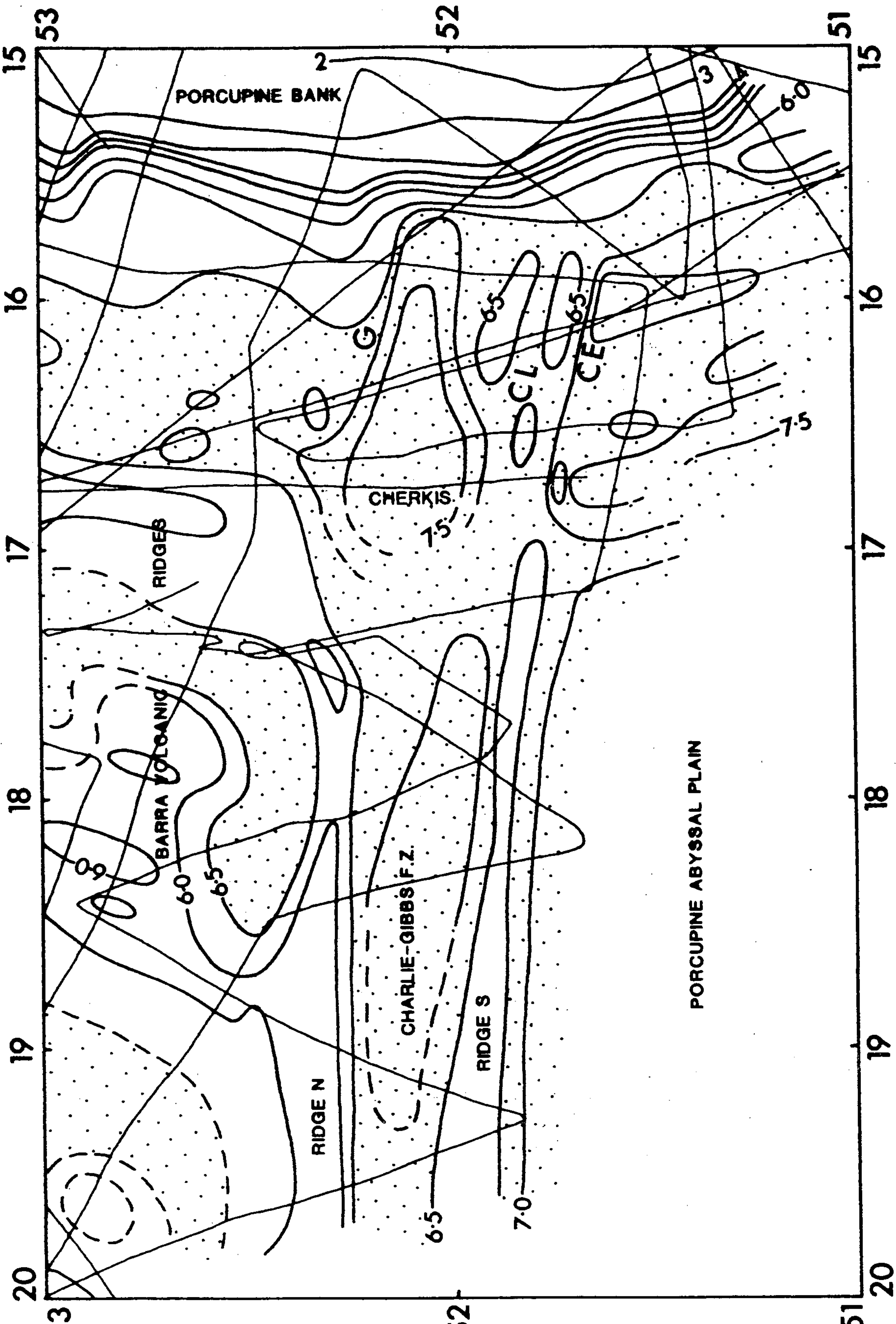


Figure 4.6 Seismic basement isochron map for the CGFZ and Clare Lineament. Contours in seconds two-way time at 0.5 s interval. Stipple highlights areas with basement deeper than 6.5 seconds. CE = Clare Escarpment; CL = Clare Lineament; G = basement gradient. Light lines show seismic control.

is no evidence for any sort of contact or discontinuity from this point south to ridge N at 0300/100. Further north-west along the profile layered (volcanic-type) seismic basement is picked up again beneath Feni Ridge. This material appears to have closely similar seismic qualities to the Layer 2 reflectors around the CGFZ even though the seismic profile here is rather poor. The dome-like aspect of the basement here results mainly from the pull-up effect of the bathymetry; in reality it is almost certainly a good deal flatter.

Rapid variations in seismic basement structure and sediment thickness both immediately north of ridge N and along N are testified by profiles C 80-11 and 12 (Figs 4.4b and d). On the first profile ridge N is broad-based and its northern flank descends evenly to a fairly distinct basement event at c.7.4 s TWT. On the nearby C 80-12 track ridge N is narrower and steeper, and seismic basement to the north is discontinuous and poorly defined. Some of the stronger flat reflectors are clearly intrusive sills. Ridges S and N of the CGFZ are seen to be still influencing the sedimentation pattern in the area; witness, for example, the sea floor perturbances on C 80-11 (Fig. 4.4a).

The structure and nature of the broad basement high beneath the Clare Lineament are less well understood than the oceanic realm to the south and west. The high has a double-ridged appearance on each of the profiles that crosses the discontinuity, with the exception of S 79-9. Even on this line however some relief may be present but masked by sills just above deeper basement (as hinted at by line CM-04 only 2 to 3 km to the east). The ridges are rather different from those ridges (S and N) in the CGFZ, even where the former are most pronounced as on profile C 80-14 (Fig. 4.4g). On profile CM-04, where the vertical exaggeration is about half that of the remaining tracks, the basement relief appears quite subdued (Fig. 4.5). This line also shows clearly how the high consists of short strong irregular reflectors which block out signals from any deeper events. On the double-channel and unprocessed multichannel profiles this layer is accompanied by numerous hyperbolic diffraction events indicative of faults or lavas. There is widespread consent that this seismic basement consists largely of abundant lava flows.

Whereas the profiles S 79-8, S 79-9 and C 80-14 seem to show the deeper sediments draping the basement high at the Clare Lineament the multichannel CM-04 record (Fig. 4.5), which one would expect to be easier to interpret, is ambiguous in this respect. Does the R7 reflector, seen clearly at SP 3800, continue beneath a thin pile of lavas and sills and join up with the near-horizontal layering at a depth of roughly 7.0 s TWT below SP 3500; or does it drape over the lava flows and form the short possibly sedimentary reflector beneath SP 3600 ? This latter explanation is more in keeping with the disposition of the reflectors on the double-channel profiles, especially at 2300/092 on S 79-8 and 1430/101 on C 80-14 (Fig. 4.4) where a package of reflectors infills the trough between the two ridges. The question then is do these sediments belong to the pre- or post-R7 sequence? The abrupt termination of the shallow seismic basement at SP 3070 demands either that there is a steep basement scarp down to oceanic crust or that the deeper sediments persists northwards below the volcanic basement. The gravity data have some bearing on this problem, as will be described in the next section of this chapter.

The exact structural and age relationships between the Clare Lineament basement high and oceanic crust are never clear from the four seismic profiles crossing these features (Figs 4.4e-g and 4.5). Each of the profiles shows a gap in seismic basement and an overlying zone of confused or chaotic reflectors at the position of the Clare Escarpment (Fig. 4.6). On the S 79-9, C 80-14 and CM-04 profiles the oceanic crust appears to stop abruptly, to be replaced by elevated, more variable basement. On profile S 79-8, however, this change in character is much less pronounced and the position of the Clare Escarpment is best marked by the disruptions to the sea floor and sedimentary reflectors. If the other evidence was ignored it might reasonably be assumed that oceanic basement is present from 1600/092 to 0030/093 on this profile.

North of the Clare Lineament the seismic basement is never well imaged on the double-channel seismic records. The existence of large sills in the sedimentary sequence also contributes to the poor definition of the basement. A good indicator of the distribution of basement highs and lows on all four profiles is the draping and infilling by the deeper sediments. Where seismic basement is well seen it has the same character as the ridges of the Clare Lineament to the south. Two important observations can be pointed out: firstly, the seismic basement strongly resembles overlapping lava flows, or interbedded lavas and sediments; secondly, the same basement pre-dates all the sediments seen on the profiles. Both points are nicely illustrated where the pre-R7 sequence drapes and thins out onto the basement high at SP 4600 on CM-04 (Fig. 4.5).

Beneath the trough between the Clare Lineament and the scarp G reflections from basement are indistinct. From SPs 4000 to 4100 on profile CM-04 (Fig. 4.5) a diffuse group of reflectors is present at a depth of c. 7.6 s TWT and below a thick layered sedimentary sequence. It is accompanied by numerous diffraction hyperbolae and displays a remarkable similarity in seismic character and depth to the oceanic basement south of SP 3050. Poorly developed lateral equivalents of this material are visible on the adjacent north-south double-channel seismic profiles.

The sedimentary cover to the Charlie-Gibbs Fracture Zone

A thorough discussion of the seismic stratigraphy of the Rockall area, including the CGFZ, was given in Chapter 3. In addition, the general seismic properties of the main sedimentary intervals and their bounding reflectors were given there. This section provides a closer look at the geometry of, and structures within, the sedimentary sequence around the Rockall offset margin.

One of the more important confirmations of this work, and one that is fundamental to our knowledge of the age of Rockall Trough, is the recognition that the sediment column thickens fairly abruptly moving north across the Clare Escarpment (Fig. 4.5) and ridge N of the CGFZ (Fig. 4.4a and c). This manifests itself as a noticeable shallowing of the isobaths over the whole of Rockall Trough compared with Porcupine Abyssal Plain, and also as the sea floor slopes at the above-mentioned discontinuities. The main reason for these changes is the existence of a pre-R7 sedimentary sequence north of the CGFZ-Clare Lineament and its absence to the south over oceanic crust. This old, deep succession tends to smooth out the topographic variations in the underlying basement. Internally the interval is typically well layered and contains the occasional angular unconformity. The top of the succession, R7, is almost invariably a strong persistent reflector that contrasts greatly with the suprajacent transparent interval (e.g. Fig. 4.4e).

If the interpretation, in Chapter 3, of R7 as a late Early Cretaceous reflector roughly contemporaneous with the formation of oceanic crust south of the Clare Lineament is correct, then the basement north of that feature must predate the oceanic crust and, obviously, the pre-R7 series. Despite the fact that there are frequent thin pockets of layered sediments in small basins on the top of oceanic basement off Porcupine Bank and Goban Spur, it would be necessary to invoke a very large and sudden change in deposition rates across the CGFZ-Clare Escarpment if one wished to maintain all the basement depicted in Figures 4.4 and 4.5 as being of the same age as the oceanic crust. At SP 4100 on CM-04 (Fig. 4.5) there are at least 0.8 s TWT of pre-R7 sediments above basement; assuming a conservative P-wave velocity of 3.0 km/s for the sediments this converts to a thickness of 1.2 km. This compares with no pre-R7 sediments over known oceanic crust on the same profile (south of SP 3050). This contrast is heightened further north in Rockall Trough where the pre-R7 interval is even thicker in places (e.g. 1.2 s TWT + on profile GSI-1, Chart 1 and Fig. 5.6). This evidence, then, could favour an older, pre-middle Cretaceous age for the basement flooring Rockall Trough -an idea that has been mooted independently by D.K. Smythe (pers. comm.).

On the four seismic profiles that traverse the Clare Lineament in a north-south direction the R5 to R7 succession is a conspicuous seismically transparent interval. The upper Cretaceous R6 reflector is reasonably well defined on profile CM-04. But on the poorer quality academic profiles R6 tends to occur at the base of the bubble-pulse reverberations from the overlying R5 event and so cannot be picked with confidence everywhere. R5 and R6 appear to diverge slightly as the profiles progress northwards (e.g. C 80-14, Fig. 4.4g): over oceanic crust they have a fairly constant separation. A persistent though slightly weaker reflector is observed a short way below R6 on the S 79-9 and CM-04 profiles. Otherwise this pale interval, which attains a maximum thickness of 1.4 s TWT (c. 2km at 3.0 km/s) along S 79-8 and -9, contains only a few weak discontinuous reflectors.

To the east and west of the Clare Lineament the picture is not quite as straightforward. On line S 79-13 (Figs 4.3 and 4.4h) the R5 to R7 interval becomes heavily layered and changes thickness often across the continent-ocean transition. The projection of the Clare

Escarpment intersects this profile at 0230-0300/096 where the R4 and R5 reflectors step up by at least 0.5 s, and where the deeper reflectors are rather chaotic. Further north-east the sediments drape and infill a series of continental half-grabens. The weak basement reflectors between 0330/096 and 0500/096 probably represent a highly modified (?intruded) rotated fault block. A sizeable half-graben is present north-east of the continental horst block at 0600/096. The strong R5 event can be picked with confidence at quite shallow depth below this basin but there is little evidence of a deeper post-rift (drift) unconformity or of seismic basement, such as were drilled over Goban Spur (Masson et al. 1985). The strong short reflector at 4.3 s depth at 0800/096 is best explained as a sill.

Over the Charlie-Gibbs F.Z. the pre-R5 succession retains a general quality of seismic transparency, but the presence of strong basement relief here has obviously contributed to the greater variability of this unit. The irregular changes in the shape, amplitude and layering of the reflectors in the R5-basement sequence makes it difficult to correlate across the S and N ridges. North-west of 2300/099 on C 80-9 (Fig. 4.4a) the same interval returns to its more normal appearance. At around 2200/099 a sill intrudes up to the level of a persistent reflector just below R5. This is in support of the R6 age assigned by the present author and is consistent with the association of sills and R6 elsewhere in Rockall Trough.

The seismic appearance of the post-R5 sequence is highly variable both along and between profiles, and the continuity of many reflectors is not convincing. As a generalisation it can be stated that the R4 to R5 interval is considerably more uniform than the sea floor to R4 interval, where the two can be defined. In the vicinity of the Clare Lineament the former interval is typically semi-transparent with an occasional persistent even reflector(Figs 4.4g and 4.5). Reflector R4 is observed as a distinct though discontinuous horizon over the Clare Lineament. Further north R4 deteriorates and may even pinch out against the R5 horizon near SP 4650 on CM-04 (Fig. 4.5). A similar pattern is visible on the adjacent profiles.

The overlying sea bed to R4 sequence exhibits a variable, confused pattern of reflectors, more so than any other interval. Although a small number of persistent flat-lying horizons can be detected this sequence is dominated by numerous short crenate events. These are difficult to represent in their entirety on a line drawing interpretation but Fig. 4.5 gives an indication of their pervasiveness. In places this cuspate pattern is seen to disrupt the comparatively uniform R4 reflector (e.g. SP 3500, CM-04). A relatively flat unbroken reflector, usually 100 to 200 ms below the sea bed, appears to form the upper limit of this chaotic interval; above it there is a return to more even sedimentation.

To the west, above the Charlie-Gibbs F.Z., R4 cannot be identified and the post-R5 sequence is characterised by gently undulating reflectors of varying amplitude (Figs 4.4a-d). Onlap of these onto the basement ridges and R5 occurs, and some internal unconformities are present, but the crenated aspect seen so vividly to the east is only locally developed here (e.g. 0900/101, C 80-13). Of interest, however, is the development of a large lens-shaped mound of mostly transparent sediments immediately north of ridge N (C 80-11 and 12, Fig. 4.4). A similar but smaller body is developed on the southern flank of the ridge on C 80-12. Both features are also picked out on the nearby Discovery cruise 60 profile (0500 to 0700/066). Scrutton and Stow (1984) reported on these features, which they termed contourites, and assigned to them an Eocene-?Palaeocene age. This was based on the recognition of the Eocene-Oligocene R4 reflector draping the mounds. I have tied the main packages of reflectors here into the CM-04 multichannel profile at SP 5000 via the Discovery 60 and Shackleton 79-14 profiles (Fig. 4.3). This correlation firmly favours the positioning of R5/R6 at the top of the transparent bodies and in doing so predicts an older, middle Cretaceous age for their formation. Such an interpretation is also consistent with the widespread transparent nature of the R5 to R7 or basement interval.

Finally, there are very few convincing examples of faulting within the sediments throughout the area embraced by this chapter. To some extent this reflects the poor resolution of many of the available seismic profiles. It does imply, though, that there has been little tectonic activity following the formation of basement in this area. The best evidence for faulting is seen over the Clare Escarpment on profiles S 79-9, CM-04 and C 80-14 (Figs 4.4 and 4.5). Even here the vertical exaggeration is such that what may initially look like a fault may in fact be simply an effect of sediment drape (e.g. 1300/101 C 80-14).

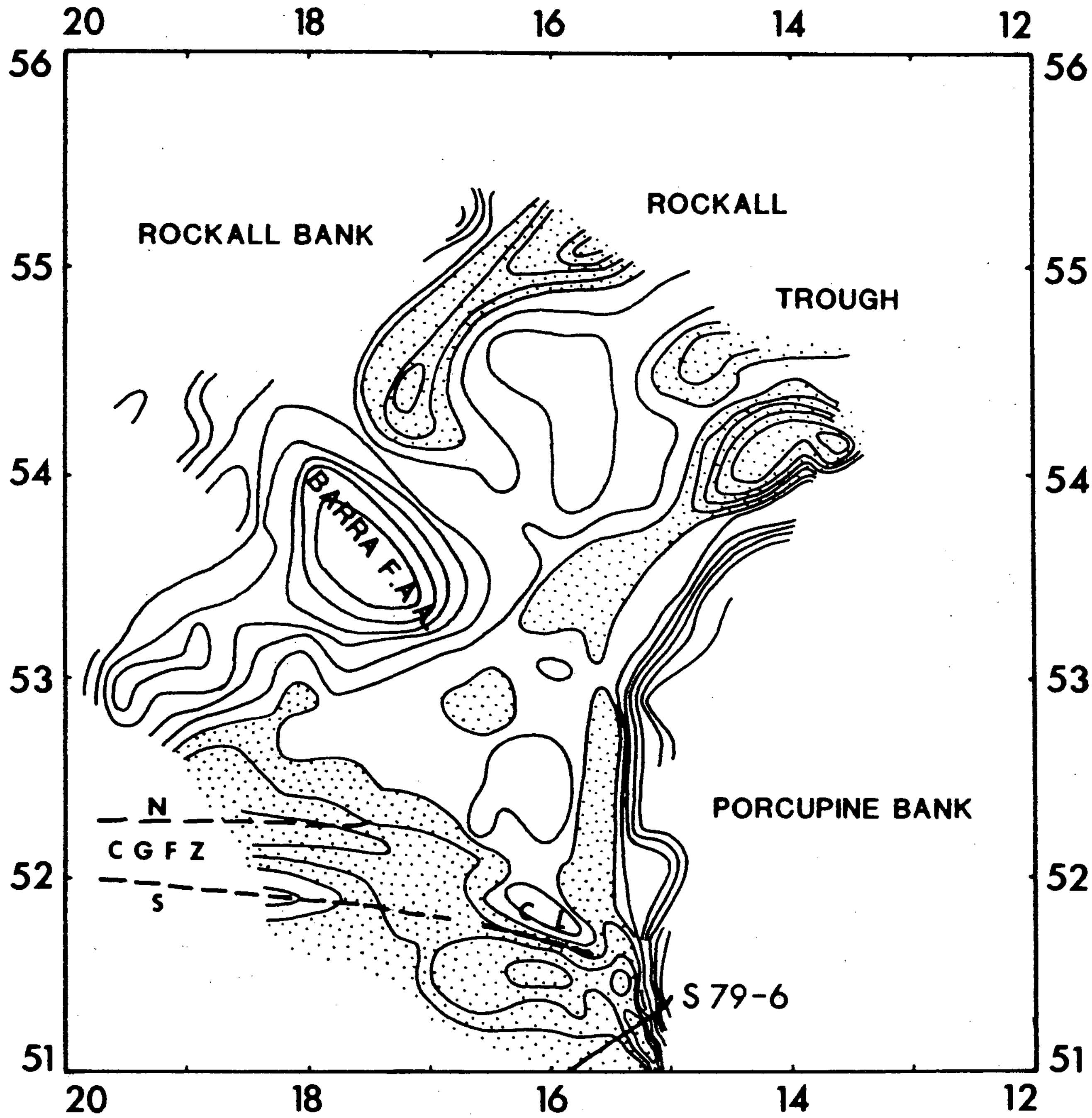


Figure 4.7 Simplified free-air gravity anomaly map of southern and central Rockall Trough, reduced from Chart 2 (back pocket). Contours every 10 mgal, stippled below 0 mgal. BARRA F.A.A. = ovoid Barra free-air anomaly (see text). CGFZ = Charlie-Gibbs Fracture Zone, bounded by Ridges N and S (dashed lines). CL = Clare Lineament. S 79-6 marks location of profile illustrated in Fig. 4.8.

4.2.2 Evidence from gravity observations

The hand-contoured free-air gravity anomaly map of the Rockall area (Chart 2, back pocket) is reproduced at a reduced scale in Figure 4.7. The obvious features on this chart can be discussed initially in a qualitative manner. The gravity field is dominated by two conspicuous gradients, both edge effects over major bathymetric features. A large free-air edge effect is associated with the steep continental slope of west Porcupine Bank which descends from c.500 m to over 4000 m in the Porcupine Abyssal Plain. The free-air anomaly exhibits a concomitant drop from 60 mgal to -30 mgal and gives rise to the north-south trending anomaly low over the base of the continental slope; this pattern is typical of much of the continental margins bounding Rockall Trough.

It is noted that where the Clare Lineament intersects the margin there is a slight but noticeable change in the free-air edge effect (Fig. 4.7). To the south a very steep gravity gradient reaches a minimum of -30 mgal in a narrow trough. To the north the gradient is shallower and the minimum is broader and does not even reach -10 mgal. In addition a conspicuous embayment occurs at 52°N 15.2°W, just north of the above-mentioned junction. These changes are due both to the shallowing of the sea bed in the Trough and to the more moderate attenuation of continental basement north of the Clare Lineament. To the south rapid attenuation of basement across steep normal faults results in the steep free-air anomaly (FAA) gradient here (Fig. 4.8).

The indentation in the FAA contours at 52°N corresponds to a well-defined SW-NE trending canyon on the continental slope (Fig. 4.3). The seismic reflection coverage over this canyon is poor and so it is difficult to establish the depths and trends of the underlying basement (Fig. 4.6). Certainly its relationship to the Clare Lineament is far from well understood. Megson (1983) shows the same embayment as a much more sharply defined, almost rectangular indentation on her free-air anomaly chart, and one that corresponds to a large SW-NE depression in acoustic basement. The present author used much the same (plus additional) geophysical data and finds it difficult to justify such detailed contouring. The acoustic basement chart of Roberts et al. (1981) covering much the same area fails to

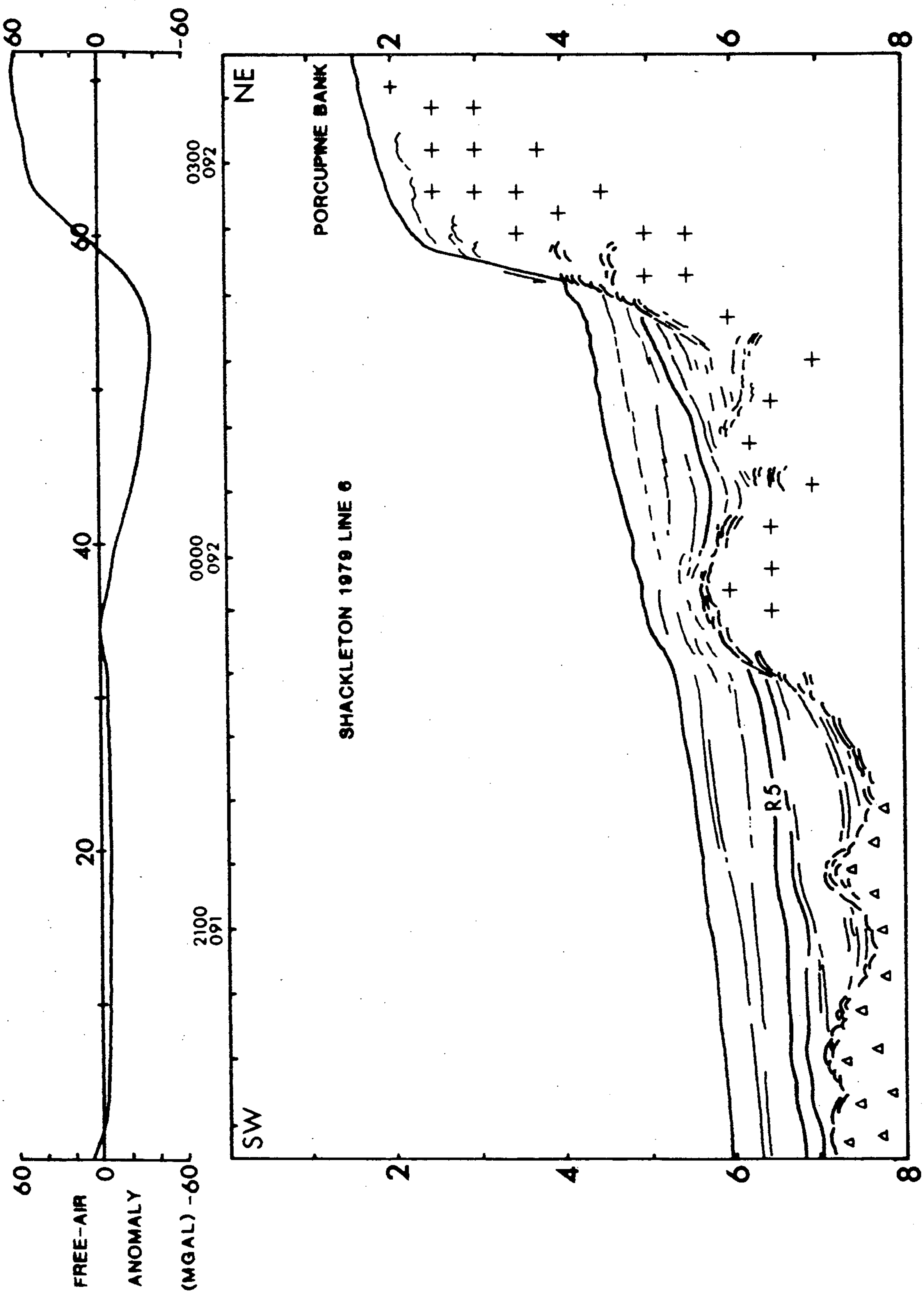


Figure 4.8 Interpretation and free-air anomaly along seismic profile Shackleton 79-6. See Fig. 4.7 and Chart 1 for location. Depth in seconds two-way travel time. Crosses = continental basement; triangles = oceanic crust.

highlight a similar depression. Likewise a FAA chart of Porcupine Bank and the shelf west of Ireland provided by D.G. Masson (pers. comm.) shows a smooth embayment similar to that of Figure 4.7, and does not support the sharply defined contouring of Megson (1983).

However, on seismic profiles C 80-1 and WI-10 (Chart 1) a narrow simple graben is seen that is on a direct north-easterly projection of the canyon. The first evidence for this graben to the south-west near 52°N was provided by Roberts (1975). His NW-SE seismic profile at this latitude shows a single steep normal fault with a large downthrow forming the south-east flank of the graben and three or four smaller normal faults forming the north-west flank. Megson (1983) went on to map the graben for over 100 km in a north-easterly direction across Porcupine Bank between 52° and 53°N. Thus the embayment in the continental slope FAA edge effect is essentially a manifestation of the bathymetric expression of this graben system; the gravity anomaly dies out to the north-east as a result of the graben becoming less pronounced and infilled with sediment (this graben system is further discussed in Chapters 5 and 8).

The second important FAA edge effect is associated with the localised steep bathymetry above the Clare Escarpment (Fig. 4.7). Here a roughly ovoid 10 mgal anomaly high is oriented WNW-ESE over the Clare Lineament and this descends into a broader east-west trending 30 mgal trough to the south. This pattern is clearly disrupted towards the east where there is a complex interference between the gravity effects of the continental slope, the SW-NE canyon at 52°N, and the Clare Lineament. To the west the relative high over the Lineament is offset to the north at 52°N 16.7°W (at the -10 mgal contour); thereafter the 0 and the -10 mgal contours tend to follow the trend of the isobaths. The offset at 16.7°W effectively terminates the FAA high over the Clare Lineament and suggests that a discontinuity exists here, perhaps separating oceanic crust of the Charlie-Gibbs Fracture Zone from unidentified geological basement beneath the Lineament.

Away from the two large FAA edge effects described above the gravity field is smooth and low in amplitude and again is strongly influenced by the bathymetric trends. Over the CGFZ a tendency towards lineations is hinted at (Fig. 4.7) but this may be due partly to poor data coverage and the roughly north-south orientation of the available gravity tracks - a direction that favours east-west contouring. Despite this there is a reasonable indication that the S and N ridges of the CGFZ (Fig. 4.6) are associated with a linear gravity high and a low, respectively (Figs 4.7 and 4.9). The positive anomaly over the S ridge is much as one would expect from an obvious mass excess, as will be shown shortly. But the 20 mgal low slightly south of the N ridge, which on seismic profiles assumes the same appearance as the S ridge, is rather more surprising. It supports the idea, implied earlier from the seismic reflection data, that some change in crustal geology and geometry pertains across the northern edge of the Charlie-Gibbs FZ.

The oceanic crustal areas in and south of the CGFZ are everywhere associated with a negative free-air anomaly, whereas the remainder of southern Rockall Trough, excluding the FAA lows at the bases of the continental slopes, is almost entirely positive. This property is also picked out on the along-track bathymetry and FAA profiles crossing the CGFZ which have been stacked with respect to the S ridge and the Clare Escarpment (Fig. 4.9). Three of the FAA profiles, C 80-9 and 10, CM-04 and C 80-14, have been corrected for variations in the depth to the sea floor to arrive at a Bouguer-type anomaly. Although these are not strictly Bouguer anomaly profiles in the sense that they are applied in land-based studies, the analogy is convenient for this and subsequent discussions. The effect of the water column (1.03 g/cm^3) was added to the FAA profiles assuming a density for the sediments of 2.25 g/cm^3, a value that is a rough average of the densities used in similar studies in the Rockall area.

Hence the three Bouguer anomaly profiles in Figure 4.9 are not influenced by the bathymetry; that is the water column has been replaced by sediments assuming a density contrast of 1.22 g/cm^3. What they illustrate is a rise in Bouguer anomaly over the oceanic crustal regions, particularly on C 80-14 and C 80-9/10 (Figs 4.9a and g). A similar rise is visible at the southern end of CM-04 (Fig. 4.9f) though the picture is partly confused by the two highs slightly to the north. A positive Bouguer anomaly over oceanic regions is widely reported in the literature (e.g. Bott 1982). Notice also the 40 mgal free-air anomaly at the beginning of profile C 80-9 (Fig. 4.9a) which is caused by the Feni Ridge but which disappears on the water-corrected Bouguer anomaly profile.

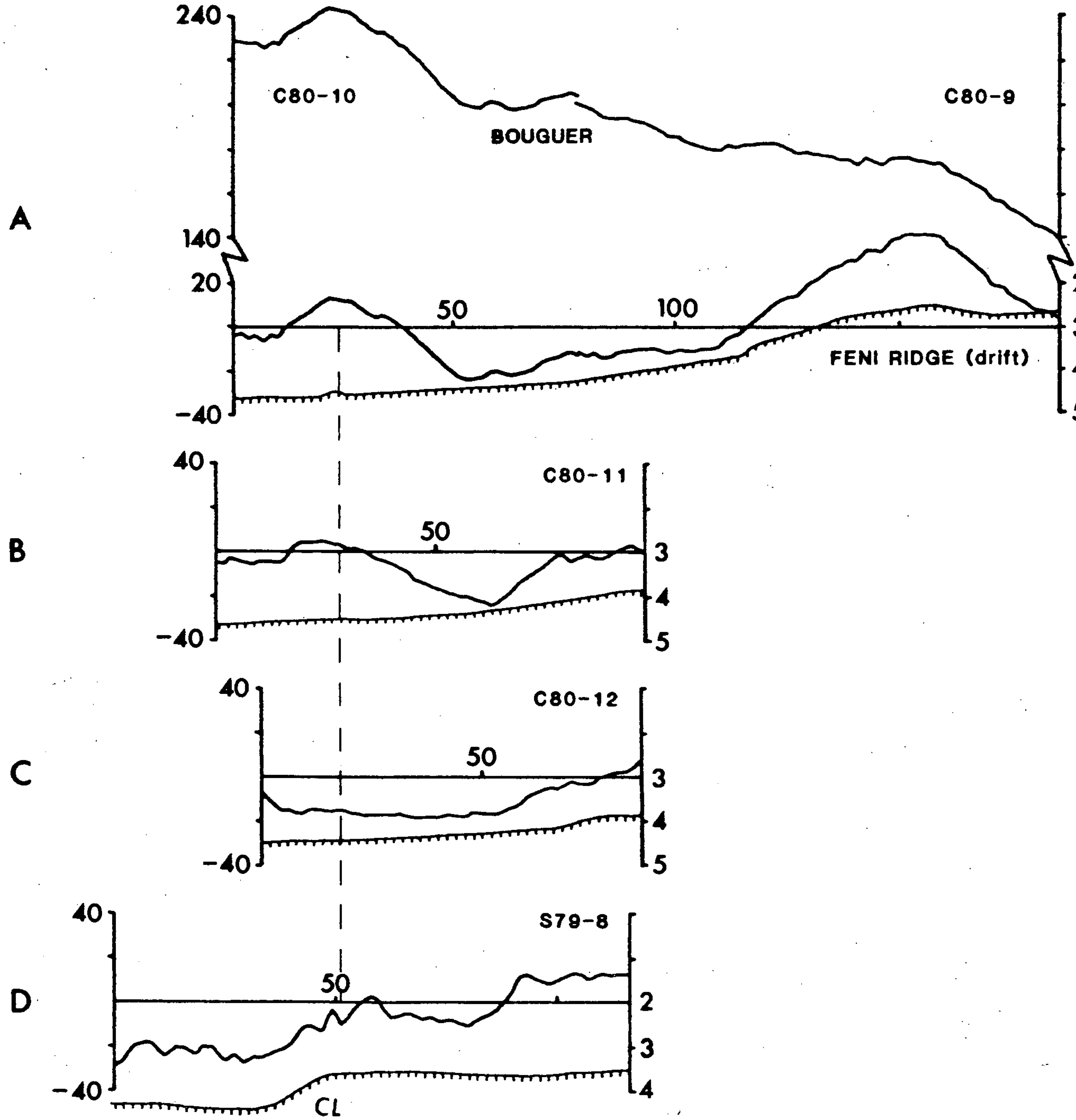


Figure 4.9 Free-air anomaly and bathymetric profiles along selected tracks across the CGFZ and Clare Lineament. All 8 profiles centred about Clare Lineament (CL) or its continuation, Ridge S - as marked by vertical dashed line. Profiles oriented south to left, north to right. Free-air anomaly - bold solid line, scale in mgal on left ordinate. Sea floor relief - light line with vertical marks, scale in km on right ordinate. Bouguer anomaly profiles on A,F and G, scale in mgal to left (see text for discussion). Horizontal scale in km.

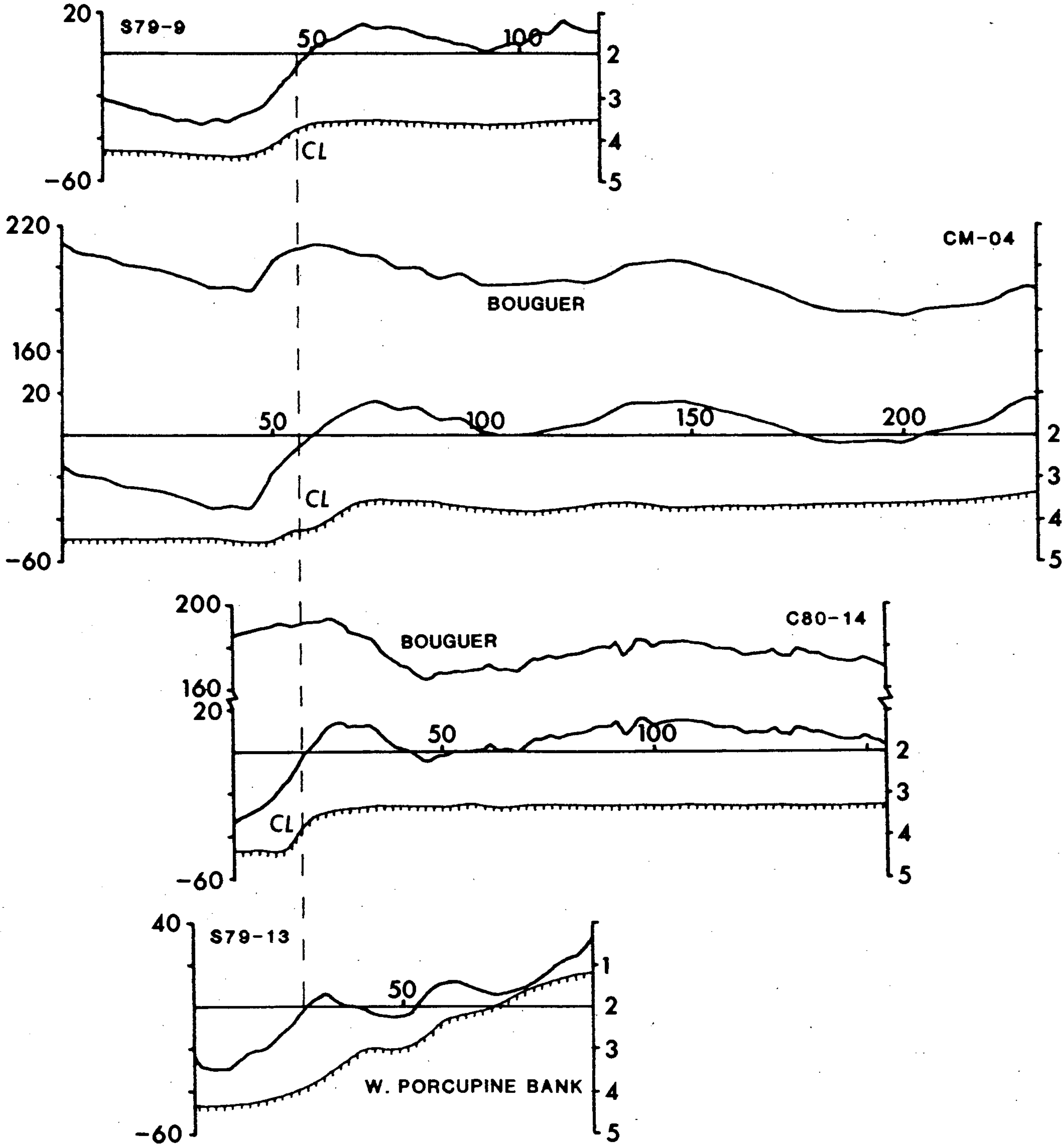


Figure 4.9 (continued). Refer to previous leaf for explanation.

Two-dimensional gravity modelling was performed along two gravity anomaly profiles that traverse the area of interest. A combination of profiles C 80-9 and C 80-10 was considered most suitable for modelling the structure of the Charlie-Gibbs Fracture Zone. Together they provide the only complete gravity profile across the Fracture Zone, though inaccuracies arise both from the two tracks not being exactly perpendicular to the gravity field and from the divergence from a truly two-dimensional potential field. The CM-04 seismic and gravity profile was used to model the Clare Lineament, in preference to the adjacent lines, because it passes over the centre of the main east-west anomalies and because it extends for at least 50 km on either side of the structure. The nearby S 79-9 profile is rather too short in this respect. In addition the multichannel CM-04 profile provides improved control on the depth to seismic basement compared with the poorer quality double-channel academic profiles. To their disadvantage this and the adjacent lines, with the exception of C 80-14 perhaps, are all slightly oblique to the perpendicular direction of the free-air anomalies.

The assumption of a two-dimensional anomaly field and crustal structure in this region is necessarily a gross simplification in view of the lateral termination of the gravity highs and lows of the Clare Lineament. However in the absence of closely spaced gravity and seismic control it is not possible to perform meaningful three-dimensional modelling in such a complicated area. Presumably the continental slope edge effect exerts some unknown, but hopefully small, influence on the CM-04 gravity profile; any such effect may reasonably be supposed to be constant around the Clare Lineament where CM-04 is nearly parallel to the continental margin. The effect on the northern half of the profile is probably minimal.

A crustal model was developed along the CM-04 profile (Fig. 4.10) by first removing the gravity contribution from the water and sediment bodies and then matching the corrected anomaly profile by adjusting the density and geometry of the main crustal layer. This method is described by Talwani and Ewing (1960). The water depths were read off the top of the multichannel record and the approximate depths to seismic (volcanic) basement were computed using a P-wave velocity of 2.5 km/s and 3.3 km/s for the post- and pre-R5 intervals, respectively (see Fig. 4.5). These velocities were picked from the velocity analyses along the profile but they seem to be too high. For

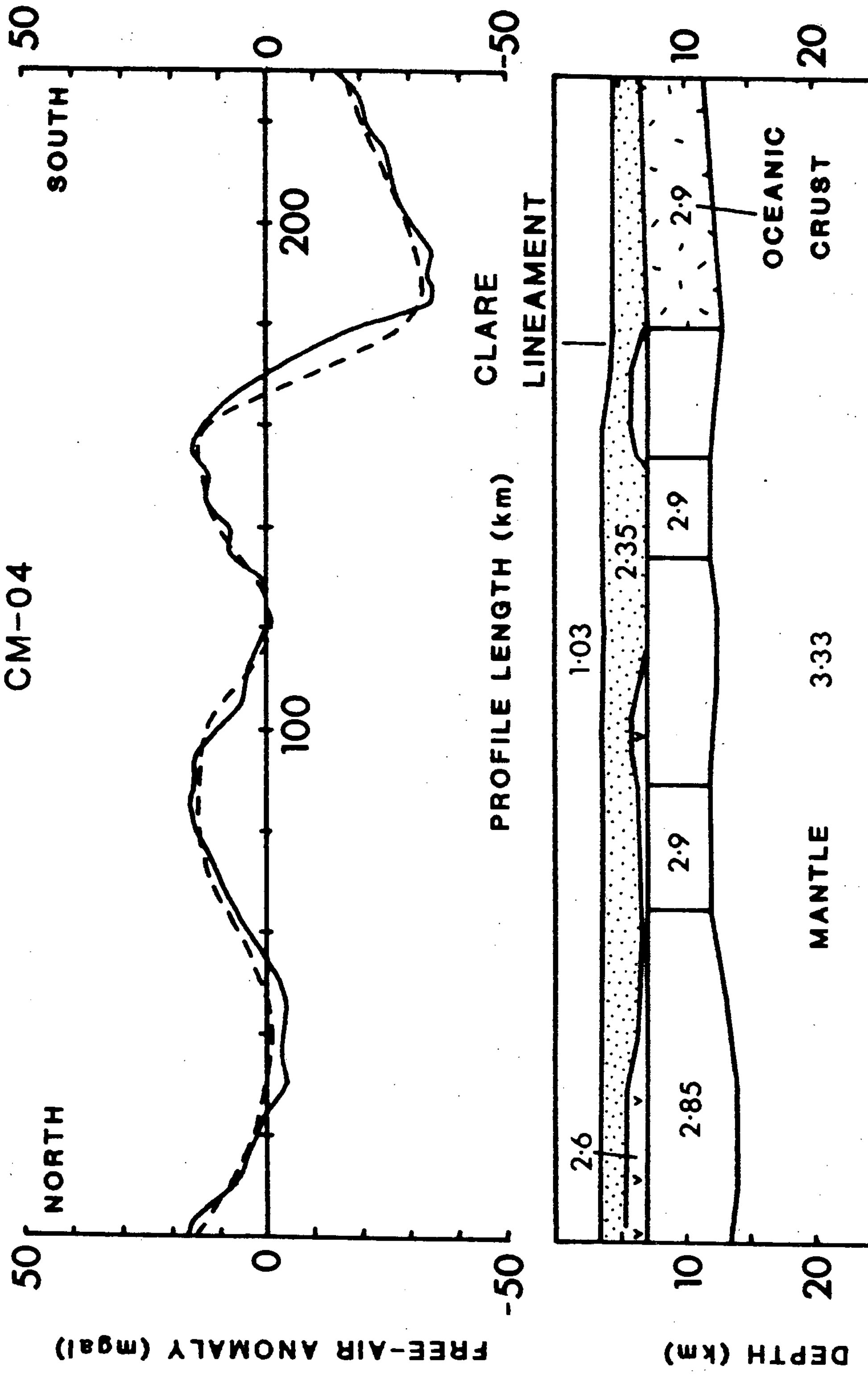


Figure 4.10 Two-dimensional gravity model along CM-04 profile. See Fig. 4.3 for location. Upper figure: solid line = observed anomaly; dashed line = computed anomaly profile. Lower figure: stipple = sediments; V = proposed volcanic basement. Numbers in prisms are densities in g/cm^3. Thin crustal layer mainly 2.85 g/cm^3.

instance, the depth of 7.8 km calculated for the basin beneath kilometre 140 was adjusted to 7.0 km to make it compatible with oceanic crust to the south and also predictions further north in Rockall Trough.

A density of 2.85 g/cm^3 was assigned to the main crustal layer under the starting assumption that it is everywhere composed of oceanic crustal material. The evidence for and against this is considered as a whole in the final chapter. The value of 2.85 g/cm^3 is a compromise for the range of densities between 2.8 and 2.9 g/cm^3 reported widely in the literature for oceanic crust. A density of 3.33 g/cm^3 for the upper mantle is used throughout this work as it is a generally accepted value.

Although a sediment density of 2.25 g/cm^3 was used to produce the Bouguer anomaly profiles of Figure 4.9, using that value in the CM-04 model necessitates thinning the crust to less than 4 km beneath the basins at distances 80 km and 140 km (Fig. 4.10). This was felt to be unrealistic and so the sediment density was raised to 2.35 g/cm^3, enabling the crust to be thickened downwards accordingly. This density corresponds to a P-wave velocity of about 3.6 km/s according to the Nafe-Drake curve (Fig. 2.4), a fact that is difficult to reconcile with the rough sediment velocities established from seismic reflection and refraction work. In the absence of good velocity and density control for the sedimentary sequence, and bearing in mind the simple assumptions alluded to above, it is unnecessary to try and refine this part of the model.

The CM-04 48-channel seismic profile (SPs 2000 to 6600) favours the idea that the topography between 0 and 120 km results from broad ridges of seismic basement comprising volcanic material (Figs 4.5 and 4.10). A fairly consistent velocity of 5.15 km/s for this seismic basement is apparent from the velocity analyses along the track. Hill (1952) computed velocities between 4.94 and 5.12 km/s at three refraction stations over a basement ridge c.100 km north-west of the end of CM-04. These velocities almost certainly refer to the same material which is probably basalt (as predicted by Hill) with a density of roughly 2.60 g/cm^3. A velocity of 5.15 km/s is equivalent to a density of 2.55 g/cm^3 according to the Nafe-Drake relationship and 2.62 according to 0.25 power law relating the two properties ($d = 0.23\ V^{0.25}$, V in ft/s). There is no evidence available to define the basal geometry of this volcanic basement so a level of 7.0 km was chosen to agree with the depth of the basin at 140 km and also the top of oceanic crust to the south. This gives a maximum thickness of 1.8 km for the basaltic layer at the start of the model profile, a figure compatible with that computed by Hill (1952).

The nature of the basement high at 160 km to 180 km is uncertain. The CM-04 seismic profile (Fig. 4.5) here indicates the presence of a number of intrusive sills, often transgressive, and these have a habit of masking the continuity of the deeper reflectors. Inspection of the adjacent double-channel profiles favours the interpretation that a volcanic basement ridge exists between SPs 3100 and 3500 on CM-04 and likely continues to SP 3750.

When the CM-04 gravity profile was corrected for the effects of the water and sediment bodies it was noted that the basins at 40-80 km and 120-150 km (Fig. 4.10) were overlain by gravity highs and the intervening volcanic ridges by lows; if the sediment density is reduced this pattern is enhanced. It became clear that the basins are underlain by either thinned or denser crust, or both. When the profile was modelled assuming a constant crustal density of 2.85 g/cm^3 it was found necessary to construct a highly irregular Moho and to invoke extensive, and questionable crustal thinning beneath both basins - less than 3 km in the case of the southern basin. The model of Figure 4.10 is a compromise between some crustal thinning and the introduction of denser blocks of material (2.90 g/cm^3). The two denser blocks may represent large intrusions of gabbroic material or simply slightly denser regions (+0.05 g/cm^3) of whatever the surrounding crust is - oceanic or continental. But the explanation preferred by the present writer is that they indicate, somewhat crudely, areas of pervasive dyking and minor intrusions within the crust. These could act as the sources for the overlying basaltic layer, in which case one would expect them to be positioned directly beneath the largest volumes of basalt, i.e. the ridges.

At the Clare Escarpment the 50 mgal drop in the free-air anomaly is almost entirely due to the rapid increase in depth of the sea floor to the south (Fig. 4.10). After the effects of the sediments were removed a small anomaly high was still present immediately north of the Escarpment. This can mostly be explained by increasing the density of the volcanic ridge here or, more satisfactorily, by raising the density of the underlying crust to 2.90 g/cm^3. In both instances the FAA edge effect is not well matched but this probably reflects the poor definition in the gravity model of the irregular bathymetry on CM-04 over Clare Lineament. To the south of the Lineament the model predicts oceanic crust with a density of c.2.9 g/cm^3 thinning from 6 km adjacent to the Escarpment to 5 km or less at the end of the profile. This is in agreement with the oceanic crustal thicknesses calculated by Hill and Laughton (1954) and Scrutton, Stacey and Gray (1971) from refraction stations in the Porcupine Abyssal Plain.

Because of the obtuse angle between the C 80-9 and C 80-10 profiles making up the 186 km long gravity model across the Charlie-Gibbs Fracture Zone (Fig. 4.11), the two sections were dealt with separately under much the same assumptions and restrictions as the CM-04 model. The relief on the seismic basement was calculated using the same interval velocities but it is expected to be only a rough approximation to the true topography owing to the poor resolution of the sedimentary and basement reflectors. The presence of the small basin at 135 km, which was questioned in the previous section, appears to be justified by the gravity data. The broad volcanic ridge between 70 km and 110 km is equated with the arcuate basaltic ridges seen to the north (Fig. 4.6) and modelled on CM-04 to the east. The southern edge of this feature was placed at the intersection of the two profiles where there is a slight change in relief and reflector character evident on the seismic profiles (Fig. 4.4a). This position essentially marks the transition between known oceanic crust to the south in the CGFZ and unknown basement to the north in Rockall Trough.

A fairly typical oceanic structure is present south of ridge N. A generally thin cover of sediments occurs above a 4-5 km thick crustal layer, a range that ties in closely with the refraction work a short distance to the south. The S ridge has a relief of about 1 km and is seen to be uncompensated at a horizontal Moho discontinuity which occurs at c.10.5 km. Thus the S ridge is responsible for a large part of the positive anomaly centred around 150 km. The crust starts thickening downwards moving north from the centre of the CGFZ, attaining about 7 km below the subdued ridge N and 10 km below the 85 km mark (Fig. 4.11). As with the Clare Lineament, then, there is good evidence to suggest that the oceanic crust thickens as it approaches the boundary with the Rockall Trough crust.

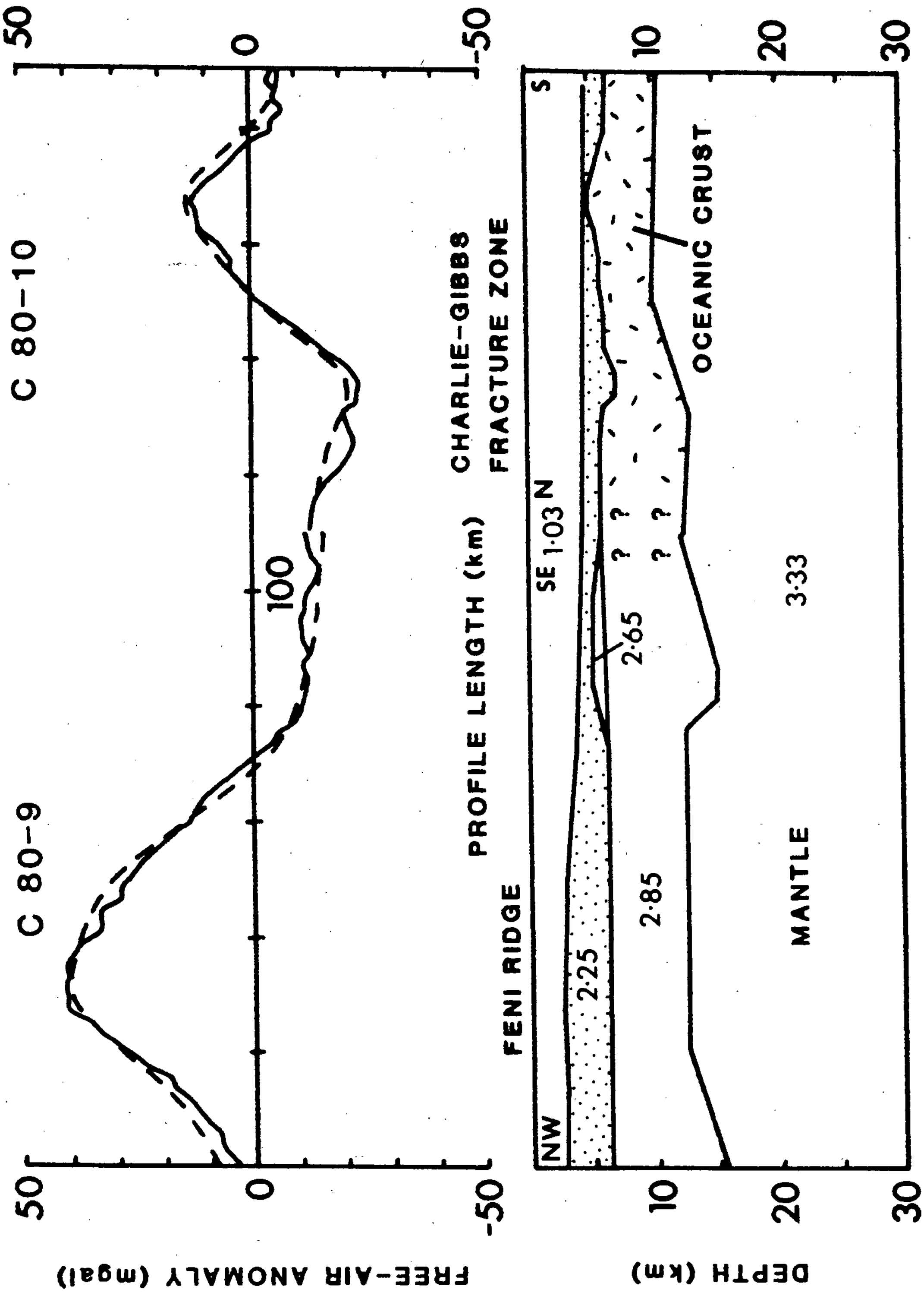


Figure 4.11 Two-dimensional gravity model along combined C 80-9 and C 80-10 profile. See Fig. 4.3 for location. Ornament, etc. as for Fig. 4.10. ? = presumed transitional zone between oceanic crust (dashes) to the south and uncertain basement to the north.

Beneath the Feni Ridge on this profile it is necessary to return to a distended crust with a uniform thickness of c.6.2 km in order to match the observed free-air anomaly and compensate for the thick sedimentary basin below the Ridge. At the north-west end of the profile the crust begins to thicken again as the continental margin off Lorien Bank at the southern tip of Rockall Plateau is approached. This thickening of the crust is probably more rapid than indicated (Fig. 4.11) since the sedimentary reflectors at the beginning of the C80-9 seismic profile (Fig. 4.4a) show early signs of draping and onlap up the margin. The truncation of the profile at this position does not allow a more precise crustal section to be developed here.

4.2.3 Evidence from magnetic observations

Maps of total intensity magnetic values or total intensity magnetic anomalies have been reported by Vogt and Avery (1974), Roberts and Jones (1975) and Max et al. (1982) for the Rockall area. That of Vogt and Avery (1974) forms the basis of the later two compilation charts owing to its close and regular track spacing. The patchy coverage and wider spacing of the profiles constituting the data base for the present work and their frequently large cross-over errors make them unsuitable for producing an improved magnetic anomaly map for southern and central Rockall Trough. However, the good coverage of magnetic profiles in the vicinity of the Charlie-Gibbs Fracture Zone enabled an anomaly map to be constructed for this area which can be compared with the previously published maps (Fig. 4.12). The controlling profiles were recorded on five different surveys and no attempt was made to reduce the accompanying magnetic data to the same International Geomagnetic Reference Field (IGRF). Therefore Figure 4.12 is not intended as a state-of-the-art anomaly map of the CGFZ but rather one that outlines the important trends and amplitudes and supports those identified by earlier authors.

Four magnetic provinces can be defined (Fig. 4.12 and also Chart 3, back pocket): i) a region of high frequency anomalies to the east of the Clare Lineament representing continental crust beneath Porcupine Bank; ii) a field of magnetic lineations within and south of the CGFZ (more obvious on Chart 3); iii) an arcuate zone of oval-shaped anomalies immediately north of the CGFZ; and iv) a magnetic quiet zone between the continental margin and the conspicuous anomalies of ii) and iii) above but including the magnetic low over the Clare Lineament.

The region of NW-SE trending magnetic lineations with maximum amplitudes between 200 and 300 nT overlies mature oceanic crust of the Porcupine Abyssal Plain, as described in the preceding seismic and gravity sections. There is some speculation as to the number identification and therefore age of the oldest anomaly, the only one that is well defined on this author's contour map (Fig. 4.12). Srivastava (1978) assigned it to Anomaly 32 (70 m.y.B.P.) while Kristoffersen (1978) proposed an older Anomaly 34 (83 m.y.B.P.) label. Roberts et al. (1981) used an A32 age in their evolutionary story of the Rockall Trough but recognised that an A34 identification was equally defendable. There is general agreement now that the latter identification is more likely (Masson et al. 1985; Scrutton 1985) though all these interpretations assume that the anomalies are in unbroken chronological order, i.e. there were no ridge jumps.

The east-west fracture zone ridges that are so conspicuous on the seismic profiles (section 4.2.1) are surprisingly poorly manifested in the magnetic field. The southern ridge S is roughly defined by the +100 nT contour west of 18°W (Fig. 4.12). The northern ridge N lies beneath the more obvious east-west positive linear anomaly at about 52.3°N which attains a peak of over 500 nT at an irregular shaped high near 18°W. This linear high is seen to mark the northern limit of the NW-SE sea floor spreading anomalies, the oldest of which, presumed here to be A34, curves abruptly into the eastern termination of the fracture zone high at 52.3N, 17.5°W (Fig. 4.12 and Chart 3). Note that the N ridge doesn't end here but continues east to profiles C 80-11 and C 80-12, where it has a strong relief, and thence curves north-east to join the ridges in southern Rockall Trough (Fig. 4.6).

In fact the magnetic signature of the northern edge of the CGFZ has close similarities to the arcuate set of anomalies immediately to the north (the Barra anomalies of subsequent chapters, named after the island in the Outer Hebrides group). Of particular significance is the way the Barra anomalies appear to curve round to the west and intersect the CGFZ almost at a tangent. This aspect is seen clearly around 52.5°N 19°W on my chart (Fig. 4.12) and more generally on that

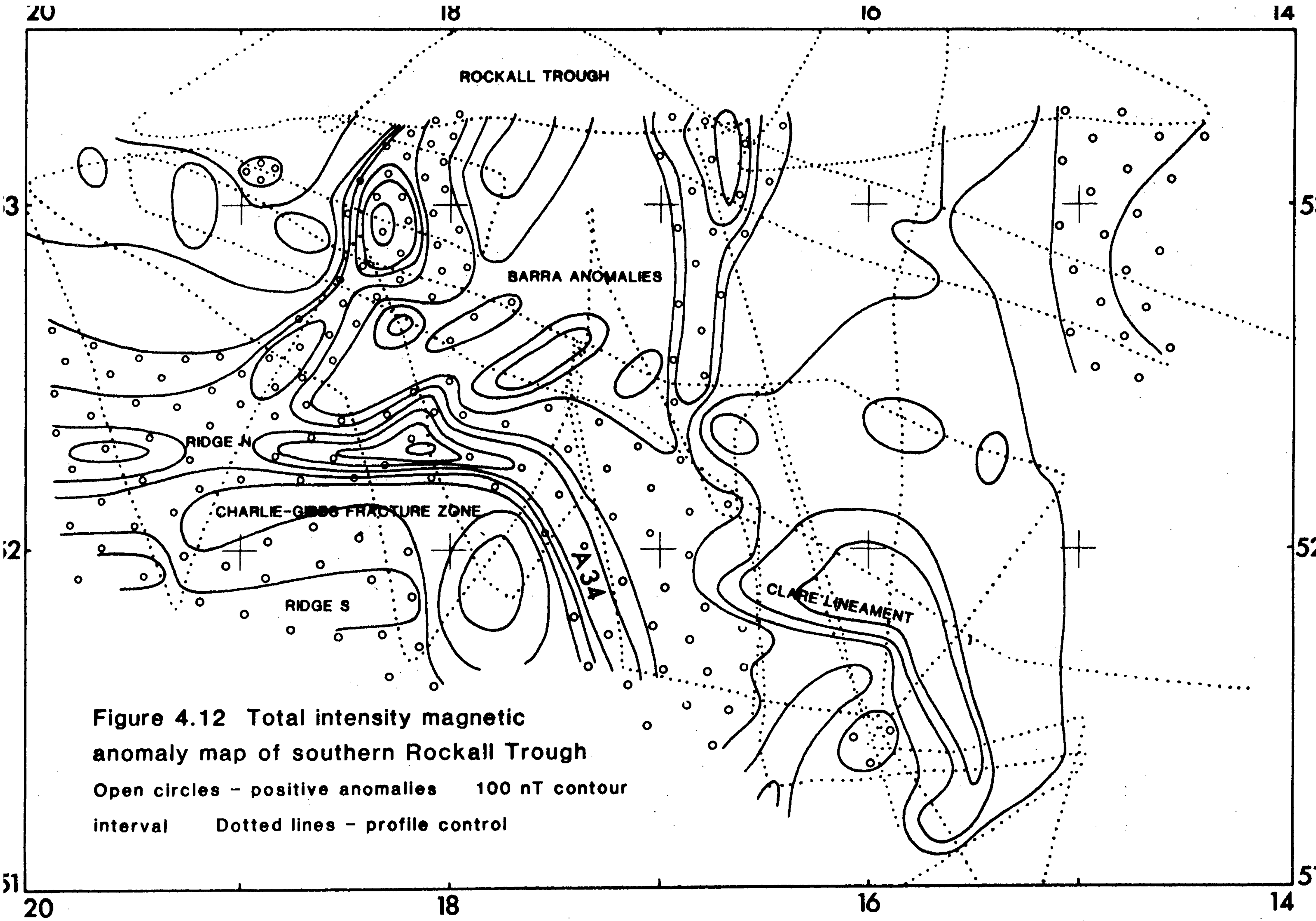


Figure 4.12 Total intensity magnetic anomaly map of southern Rockall Trough

Open circles - positive anomalies 100 nT contour interval Dotted lines - profile control of Vogt and Avery (1984) and Roberts and Jones (1975; Chart 3). The inference to be drawn from this is that the magnetic source or sources that give rise to the Barra anomalies in southern Rockall Trough may also be present beneath the northern margin of the CGFZ and result in the similar shaped anomalies there. This idea is strongly supported by the acoustic basement chart of the CGFZ (Fig. 4.6) which shows the Barra volcanic ridges bending into the N ridge in exactly the same fashion: if anything, this figure is more convincing. These volcanic ridges are accompanied by positive magnetic anomalies up to 1000 nT but their full discussion is delayed until Chapters 5 through 7.

Around the Clare Lineament the magnetic field is fairly quiet, though predominantly negative. But over the Lineament itself there is a conspicuous anomaly low that reaches a minimum of c.-380 nT (Fig. 4.12). This low is hinted at on the maps of previous authors but it is well defined and well controlled here and so its authenticity is in little doubt. The steep gradient forming the southern slope of the low has the same trend as the Clare Escarpment (105°) and lies a little way to the north, directly above the volcanic ridge identified from the seismic profiles (section 4.2.1). The magnetic trough occupies the same region as the deep basin north of this ridge (Figs 4.4 to 4.6) but at its eastern end turns abruptly SSE to form a narrow linear low at the base of the continental margin.

To the south of the Clare Escarpment, in the region where seismic reflection profiles indicate the presence of oceanic crust, the published anomaly maps show a region of weak, discontinuous magnetic lineations (Chart 3). How much this reflects the subjectivity of hand contouring is uncertain. This zone of irregular anomalies is about 70 km wide here but further south off Goban Spur it expands to at least 100 km (Masson et al. 1985). Roberts et al. (1981) believed that this region consisted of pre-Anomaly 32 (or 34) oceanic crust which persisted northwards beyond the CGFZ and on into the Rockall Trough. One would expect therefore to observe a continuation of the weak lineations off southern Porcupine Bank across the Clare Lineament into the Trough; this is not witnessed, though it is unknown how much of an effect the thicker sediments north of the Lineament have in attenuating the magnetic field - probably none.

Scrutton (1985) accounted for the weak anomalies in the region off Goban Spur by invoking topographic and magnetic intensity variations within pre-A34 oceanic crust - no dependency on magnetic polarity reversals was required. The lack of coherent sea floor spreading anomalies in this narrow zone adjacent to the continental margin is rather surprising, for the seismic reflection profiles across it exhibit a classic oceanic layer 2 appearance, especially the CM-11 profile which is perpendicular to the strike of several clearly defined fault scarps (see Chapter 3).

Thus it is seen that the magnetic anomaly field in the vicinity of the Charlie-Gibbs F.Z. and Clare Lineament is complex as a result of the meeting of a number of geological and tectonic provinces. Magnetic modelling in three dimensions was not possible owing both to the inadequate control on the distribution of basement types provided by seismic profiles and to the lack of computational facilities. Although the magnetic anomaly maps (Fig. 4.12 and Chart 3) do not indicate the presence of well developed two-dimensional structures, except perhaps over oceanic crust south of the CGFZ, magnetic modelling in two dimensions was performed along the same two profiles used for gravity modelling in the previous section, namely CM-04 and C 80-9/10 (Figs 4.13 and 4.14). The writer realises the limitations of these magnetic models but they are useful in providing a simple assessment of the distribution of magnetic sources that give rise to the main highs and lows in the anomaly field.

The geometry of the sediments and shallow seismic basement along the CM-04 profile was copied exactly from the gravity anomaly model of section 4.2.2. The shape of the deeper crustal blocks was simplified as their contribution to the anomaly field is certainly small. An intensity of magnetisation of 5 A/m (500 nT) was assumed for a 1 km oceanic crustal layer 2 and also for the volcanic (?basaltic) ridges at and north of the Clare Lineament. This value is in keeping with those intensities measured and widely used in the literature. The magnetic contribution from the deeper oceanic layer 3 is still hotly disputed; a moderate value of 2 A/m was used here. The oceanic crust and basaltic ridges were assumed to have a high Koenigsberger ratio and a remanant magnetic vector of inclination 50° and zero declination. Such an inclination is compatible with their formation and magnetisation during the Late Cretaceous, an age implied by the adjacent sea floor spreading anomalies (for magnetic

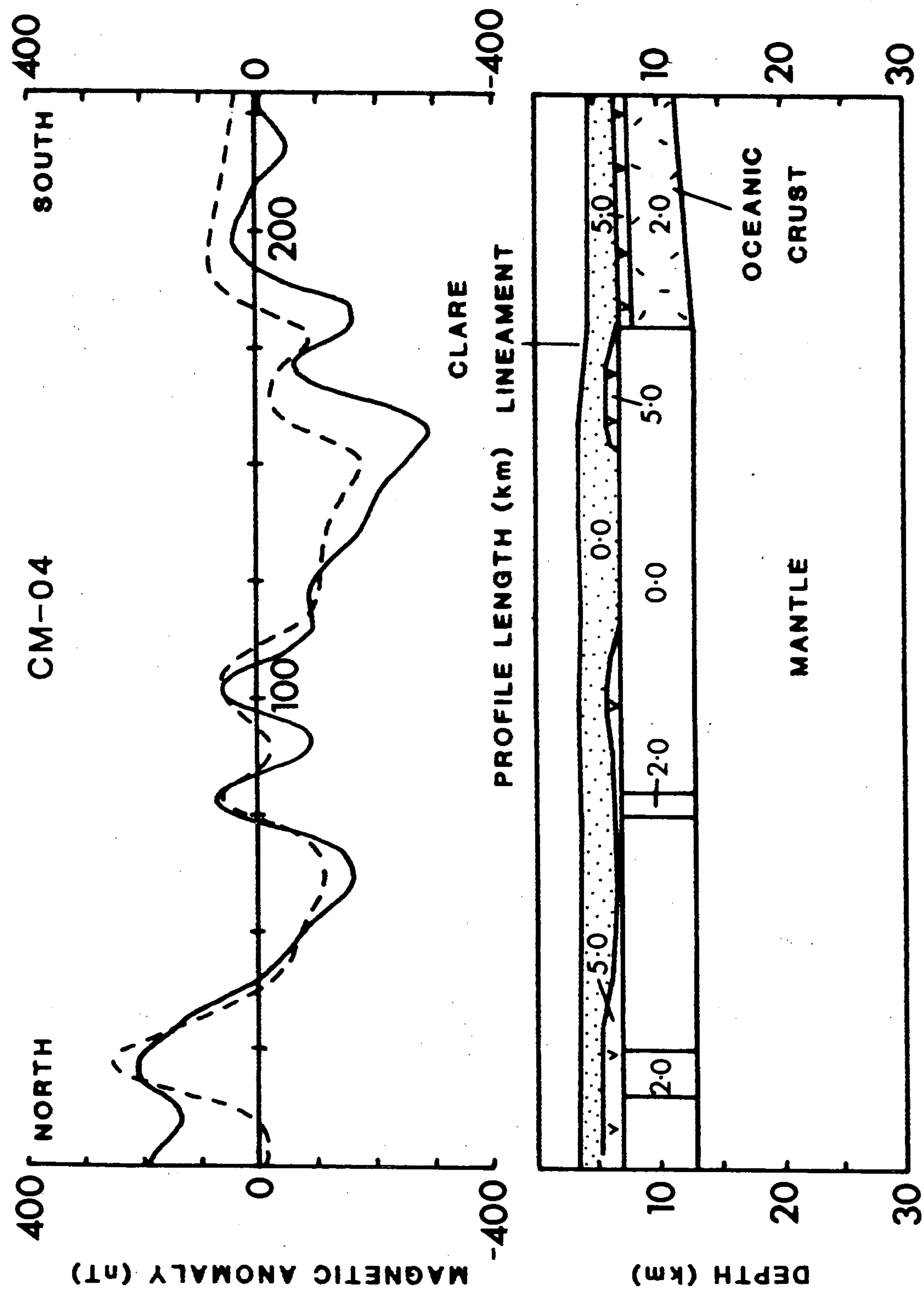


Figure 4.13 Two-dimensional magnetic anomaly model along CM-04 multichannel profile. See Fig. 4.3 for location of section. Upper figure: solid line = observed magnetic anomaly; dashed line = computed anomaly profile. Lower figure: stipple = sediment layer; V = proposed volcanic basement. Numbers in prisms are intensities of magnetisation in Amp/metre. Refer to text for discussion.

pole measurements see, for example, McElhinny and Cowley, 1980). The sediment overburden was considered to have a negligible magnetic intensity.

Over the Clare Lineament (Fig. 4.13) a good approximation to the general shape of the magnetic anomalies was achieved with the initial conditions. The fit could easily have been improved by adjusting the position, depths and shapes of the shallow basement blocks and their intensities of magnetisation. However, the model suffices to illustrate that the broad, roughly east-west trending low above the Clare Lineament could be caused by strongly magnetised oceanic crust being juxtaposed against crust with little or no magnetisation below the Clare Escarpment. The high at c.170 km is then a product of the volcanic ridge at this point. Hence the magnetic anomaly field, like the gravity and seismic data, is in support of some manner of geological and geophysical discontinuity across the Clare Lineament.

Along the northern half of the CM-04 magnetic model the trend of the anomalies from the contour maps is oblique to the profile so no attempt was made to fit closely the observed and computed curves. It was found though that the highly magnetised broad volcanic ridges could not adequately account for the observed amplitudes and steep gradients. The introduction of moderately magnetised vertical-sided bodies beneath the basalt ridges appears to account nicely for these deficiencies. Whether these blocks represent single large intrusions or zones of intense dyking is not clear from this elementary modelling. The same ambiguities were evident in the earlier gravity modelling and it is worth remarking here that the magnetised block at 75 km on the magnetic model (Fig. 4.13) is mirrored by a broader block of high density at the same position on the gravity model (Fig. 4.10).

The combined C80-9/10 magnetic model was approached in much the same fashion (Fig. 4.14). The volcanic basement ridges defined by the seismic reflection and gravity results can account partly for the magnetic anomalies when given a basalt intensity of 5 A/m. The oceanic crustal structure within and south of the CGFZ is certainly an oversimplification; the highly magnetised layer 2 may be of constant thickness. Nevertheless the association between the S fracture zone ridge and the small positive anomaly at 160 km is irrefutable (Fig. 4.14).

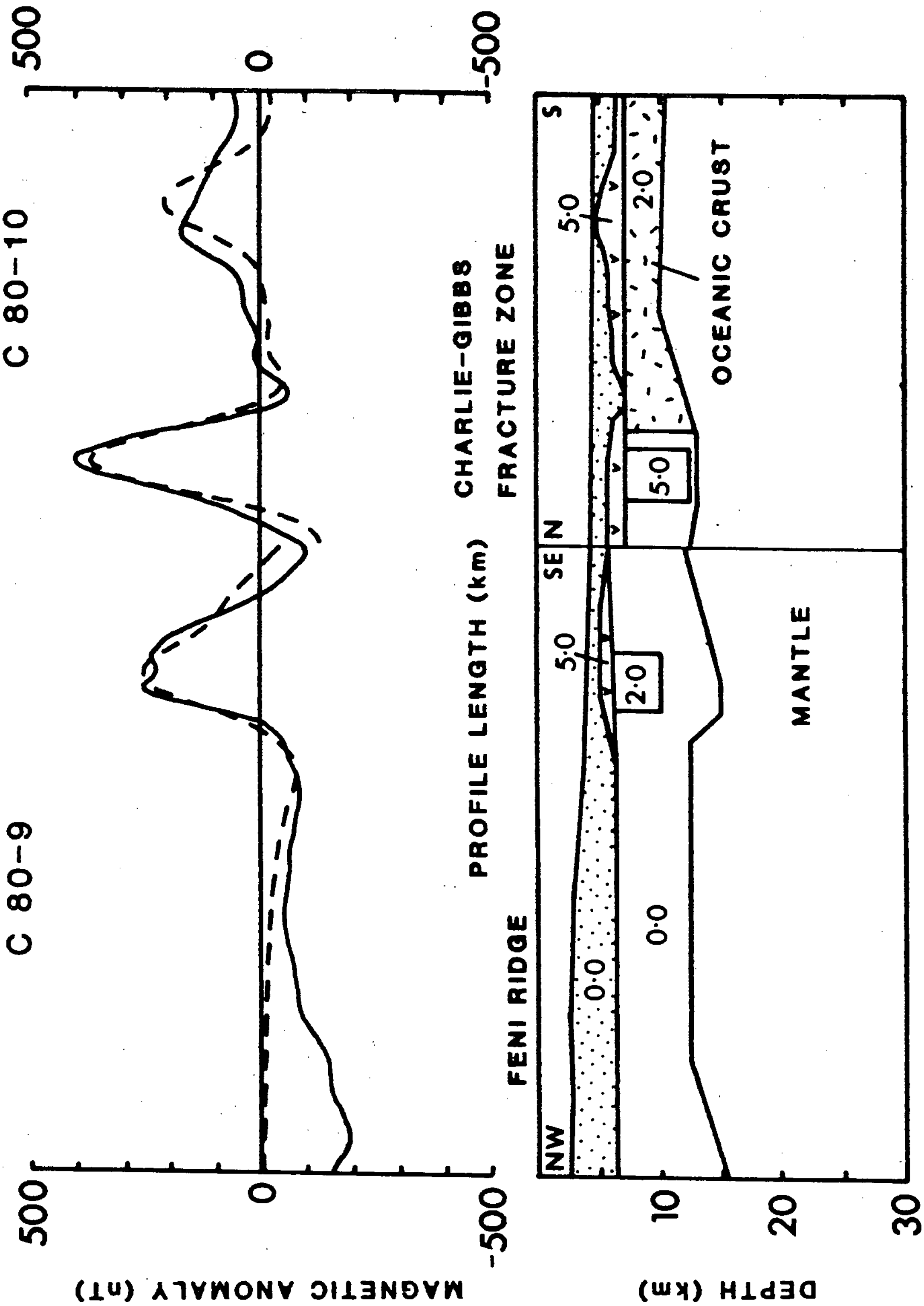


Figure 4.14 Two-dimensional magnetic anomaly model along the combined C 80-9 and C 80-10 profile. See Fig. 4.3 for location. Ornament, etc. as for Fig. 4.13. Crustal layer geometry identical to that of gravity model of Fig. 4.11.

In the middle of the model profile the two near-symmetrical anomalies cannot be accounted for simply with the volcanic ridges; again it is necessary to introduce moderate to highly magnetised intrusive-type bodies into the subjacent crust. In particular, the anomaly at 120 km, which here exceeds 400 nT (and 500 nT a short way to the east), seems to require a high magnetisation contrast below the supposed shallower volcanics. It also requires the termination of the 2 A/m layer 3 of the CGFZ roughly below the weakly defined N fracture zone ridge. If this layer continued to the north it would be necessary to increase the magnetisation contrast of the deeper block even further to produce the observed anomaly amplitude. One would not normally expect such high magnetic intensities in intrusive bodies, especially if they are slow-cooling - perhaps this is evidence that they are zones of pervasive igneous dyking instead? As an alternative, increasing the thickness and/or intensity of magnetisation of the overlying volcanic basement is unattractive since achieving the correct amplitude and matching the gradients is not possible. The seismic and gravity data do not support this hypothesis either.

Since, in depth-converting the seismic basement, this author probably slightly overestimated the deeper sediment velocities (see Section 4.2.2) the relief on the volcanic ridges in the models (Figs 4.13 and 4.14) will be at a maximum. Hence the contribution to the anomalies from the ridges will be at a maximum and it remains to call upon the deeper sources to explain the amplitude and gradient mis-fits. The shape and size of these deeper bodies is hard to determine from this simple two-dimensional modelling. That they are steep-sided is unequivocal though they may be more wedge-shaped (broadening downwards) than rectangular, as depicted here. Lastly, it is thought unlikely that these intrusive features extend to the base of the crust (Fig. 4.13), the somewhat shallower bases of the C 80-9/10 model (Fig. 4.14) seem geologically more reasonable.

5. ROCKALL TROUGH: SEISMIC REFLECTION PROFILING

The geophysical properties of the Rockall Trough and its bounding margins are considered in detail in this and the two subsequent chapters. Here the available seismic reflection data enable this region to be discussed most conveniently in terms of three widely recognizable units or, essentially, major rock types (Fig. 5.1): the deeper seismic basement (section 5.1); the sediments in the Trough and on the margins (5.2); and the widespread igneous intrusive sills, mostly within the sediments (5.3). In addition the first section, seismic basement, is further divided into continental basement, the Barra volcanic ridge system (BVRS), and deep layered basement.

5.1 Seismic basement

Continental basement beneath the margins of Rockall Trough

In the region covered by this work continental crust is observed at the south-eastern margin of Rockall Bank and the northern and western margins of Porcupine Bank (Figs 5.1 and 5.2). In the former area the structure and geometry of the margin are poorly defined owing to the paucity of seismic profiles and the masking effects of later volcanic activity. On the Porcupine Bank margin, in contrast, the comparative wealth of seismic profiles between 51° and 54°N enables a more accurate picture of the basement disposition to be constructed. Naturally in both areas the margin is marked by a conspicuous bathymetric slope and this is reflected in the shallow and closely-spaced isochrons on continental basement (Fig. 5.1). The isochrons show continental basement descending rapidly - over a distance of about 50 km - from 2.0 s TWT and less to about 6.0 s TWT. Occasionally reflectors from presumed continental crust are picked deeper than 7.0 s TWT. The basement isochrons follow the regional trend of the bathymetric contours.

The seismic character of continental basement is highly variable, tending to be shadowy, weak and discontinuous. In particular the basement is difficult to distinguish in the shallower areas,

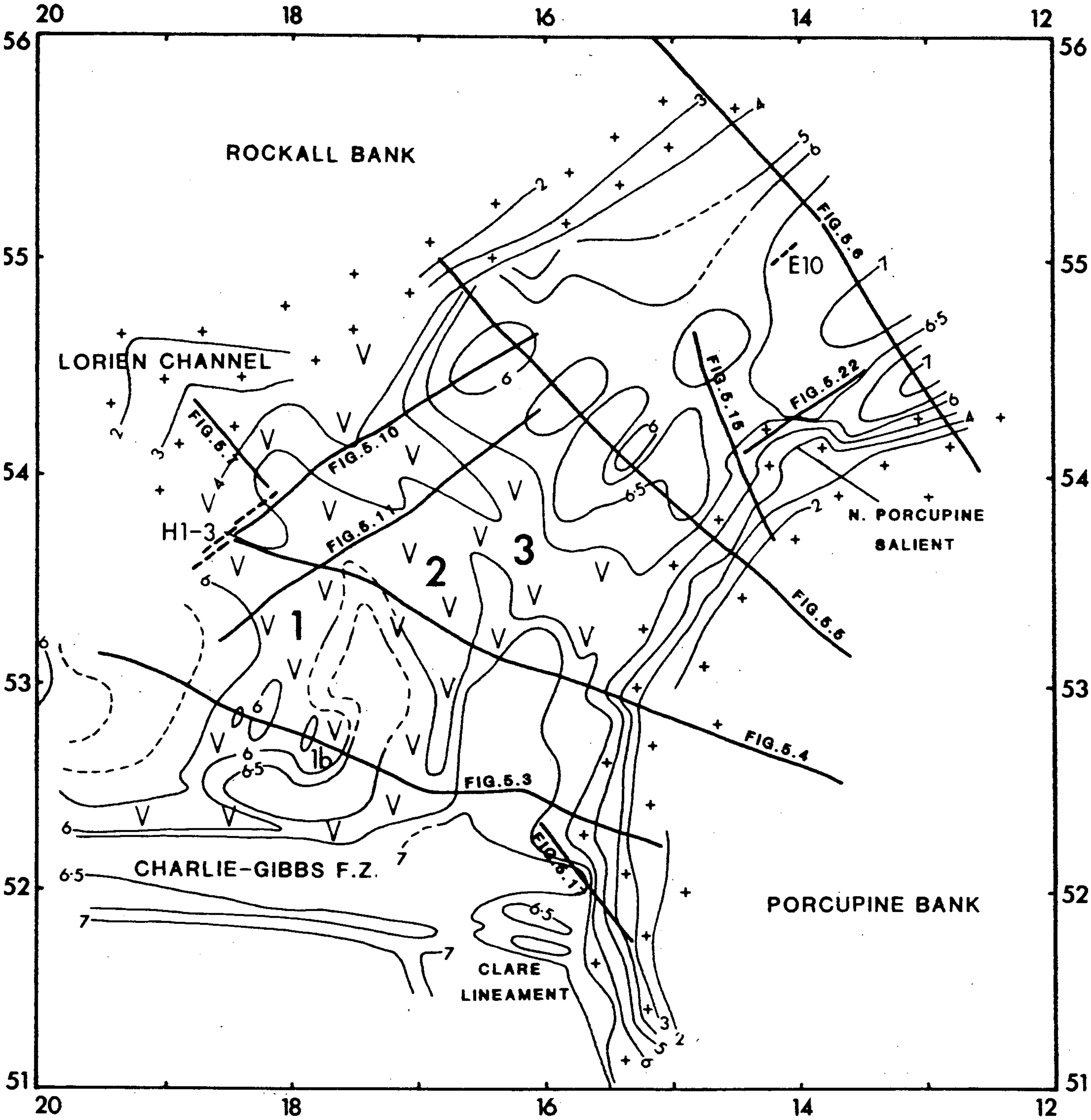


Figure 5.1 Simplified seismic basement isochron map of study area, reduced from Chart 4 (back pocket). Basement isochrons in seconds TWT at 1 second interval but including 6.5 s contour. V symbol shows distribution of Barra volcanic ridges (1, 2 and 3). Crosses show distribution of continental basement at Rockall Trough margins. Bold lines = seismic profiles illustrated in Ch. 5. E10 = refraction station of Ewing and Ewing (1959); H1-3 = refraction stations of Hill (1952).

generally less than 1 km, where numerous strong sea floor and internal multiple reflections tend to obscure the surrounding structures and sediments. At all depths it is rare that good reflectors are imaged below the top of the continental basement. Sometimes groups of subparallel dipping reflectors, suggestive of sedimentary layering, are visible at shallow depths, e.g. SPs 4450-4500, profile NA-1 Ext. (Fig. 5.6). More often though the shelf break and continental slope are characterised below by a pattern of chaotic reflectors, diffractions and side-swipes, as around 1400/096 on S 79-14 (Fig. 5.3).

The commonest structural features of the continental margins are the normal faults which in most cases downthrow towards Rockall Trough. It is, in part, these faults which result in the attenuation and rapid increase in depth of the basement across the continental margin. The number and size of the normal faults varies between the profiles but there are usually one or two faults present with large downthrows that account for the greater part of the basement deepening, as on profiles S 79-14 and C 80-1 (Figs 5.3 and 5.4). On other lines there may be a larger number of faults with smaller downthrows, the combined effect being the same as above. This structural style is highlighted on the simple plot of fault planes beneath the continental margins (Fig. 5.2). Often the crest and parts of the limbs of a downthrown fault block are visible at considerable depth (greater than 6.0 s TWT subsurface) beneath the poorly developed continental rise and are covered by a thick drape of post-rift sediments (Figs 5.3 to 5.6, especially 0700/096 C80-1, and SP 4550 GSI-1).

The precise geometry of the normal faults both near the surface and at depth is uncertain because of the inadequate resolution of the available seismic profiles. Normal faults at continental margins are frequently illustrated as curving round towards the ocean (concave upwards) and flattening out at some depth, often arbitrary, below the sea floor. These so-called listric faults have only rarely been convincingly proven and imaged (see, for example, Wernicke and Burchfiel, 1982); certainly from the data of this work there is no firm evidence for the downward curving of the faults on the seismic profiles which, in any event, probably do not extend to the required depths. Recent seismological work on active normal faults in extensional regimes (Jackson 1985) suggests that the faults are planar throughout the brittle upper crust, and have dips of between 30° and 60°. No evidence was found for seismic motion on faults with inclinations less than 30° at shallow crustal levels. However, other evidence, in particular, the rotation and internal deformation of fault blocks, lends credibility to the idea that the normal faults do flatten out below the brittle upper crust. This is thought to be accommodated by creep in the ductile crust or brittle failure of the ductile crust under large strain rates (Jackson 1985).

In certain instances groups of short, dipping near-parallel reflectors can be observed within the upper parts of the downthrown fault blocks and these may represent back-tilted sediments of the continental basement (Fig. 5.3). Such occurrences tend to support the premise that block rotation took place by motion along curved fault planes. Whatever the exact kinematics and geometry are in the development of extended continental crust it is clear that the pattern of down-faulted and rotated blocks and accompanying half-grabens reported here is typical of classic rifted passive continental margins, even though local differences and complications may result from secondary, antithetic faulting and the like.

The acoustic basement chart (Fig. 5.1) and the structural map (Fig. 5.2) depict the general trend and size of the fault scarps of the Porcupine Bank margin. While the contours of the former chart do not resolve the finer details of the horst blocks and half-grabens they do serve to emphasise the short offsets in the basement of the continental margin, as at 52.2°N 15.5°W, 53°N 15.5°W, and the N. Porcupine Salient at 54°N 14°W (Fig. 5.1). The younger sedimentary cover to the margin precludes their being manifested in the bathymetric contours, and on the structural map (Fig. 5.2) their position and general trend is highlighted by the four lines of dots. These discontinuities, which are transverse to the longitudinal structural elements of the Porcupine margin, are termed transfer faults (Gibbs 1984).

Transfer faults enable dislocations and changes in fault trend and style to be effected along an extensional zone, and this is clearly their role in the present research area. Thus the gently rounded south-east margin of Rockall Trough, as portrayed on bathymetric charts, is in fact seen to consist of a number of short, straight normal fault segments separated by these transfer structures. The transfer faults have elements of transcurrent and vertical

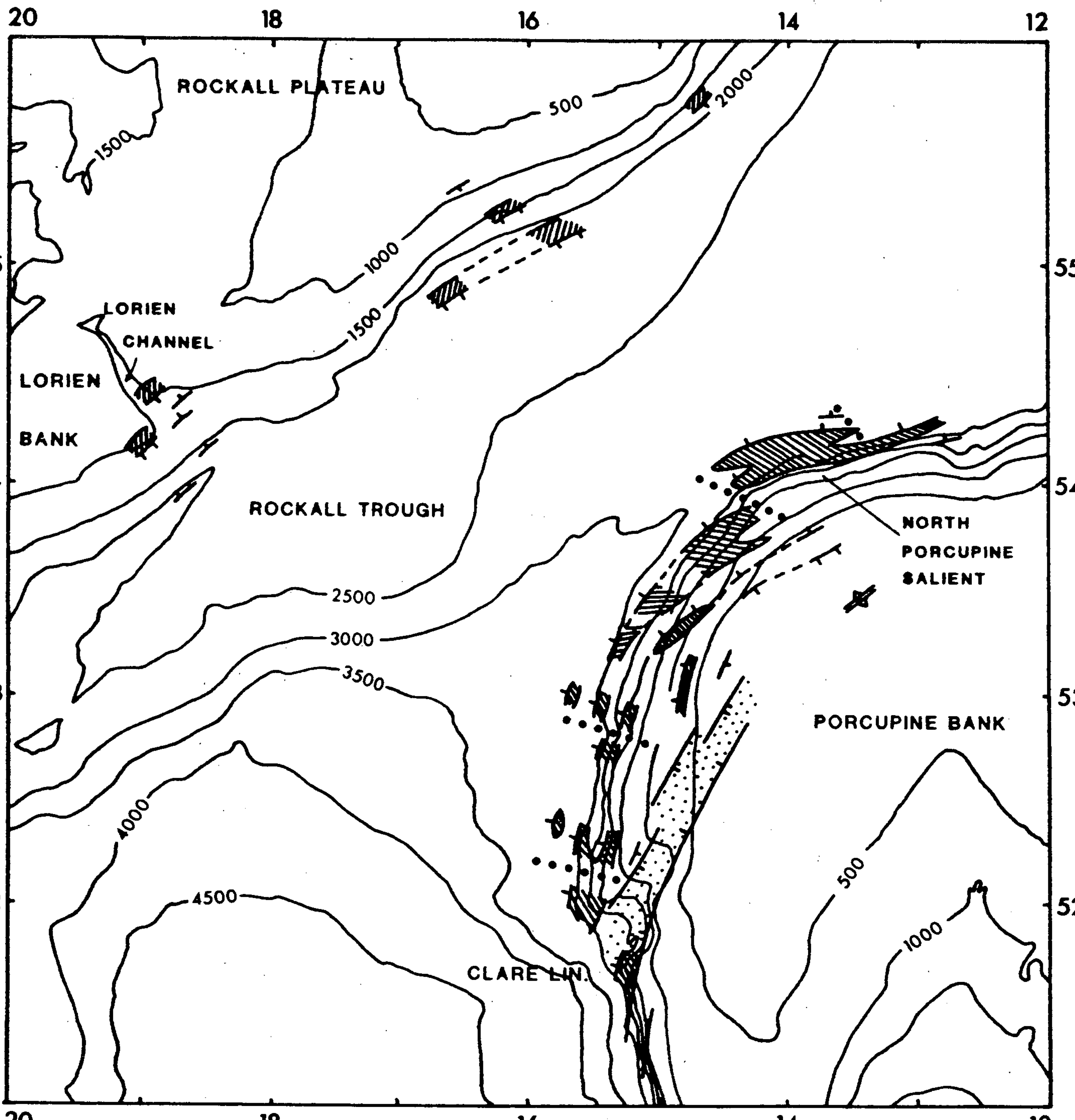


Figure 5.2 Fault patterns below the continental margins of Rockall Trough. Hachured zones indicate large normal faults and inferred continuations (dashed lines). Tick marks sense of downthrow. Smaller faults depicted as short lines with ticks. Lines of filled dots indicate approximate position of transfer structures below Porcupine Bank slope. SW - NE trending graben (stipple) partly after Megson (1983). Bathymetry in metres.

motion, i.e. they are oblique. But their continuity landward is indeterminate owing to the absence of good seismic coverage. Presumably they persist across the width of the distended zone and have counterparts on the conjugate north-west margin of Rockall Trough. In essence the transfer faults, of which there are probably more present here than the seismic reflection coverage can detect, perform the same role at continental margins as transform faults do at mid-ocean ridges - both being extensional environments. Consequently the rough trend of the transfer structures on Porcupine Bank is probably a reasonable indicator of the rifting and opening direction of Rockall Trough: certainly their orientation just south of east compares well to that of the Clare Lineament and Charlie-Gibbs Fracture Zone, two major structures that seem to relate to the opening of the North-Eastern Atlantic Ocean (Chapter 4).

Landward of the shelf break seismic profiling coverage is poor so there is little control on the position or trend of any structures in the continental basement. A prominent horst near the shelf break at 14°W is visible on profiles GSI-1 (Fig. 5.5) and WI-26. To the south-east, also on profile WI-26, a broad (c. 60 km) gentle anticline is present but the trend of its axis is impossible to establish without the presence of adjacent profiles (Fig. 5.2). Both of these structures may form part of the 53°N Flexure Zone proposed by Bailey (1975). He described this feature as a monoclinal flexure defining the southern flank of the Slyne Ridge off Ireland, and projected it across Porcupine Bank at and just north of the 53rd parallel. No evidence for a continuation of this structure is seen west of 14°W where the seismic reflection coverage is comparatively good.

On the western side of Porcupine Bank a sizeable graben is present trending south-west to north-east into the shelf (Fig. 5.2). The graben is largely infilled by sediments except towards its south-western end where a canyon with the same trend opens out onto the eastern extremity of the Clare Lineament. The positions of the bounding normal faults are mainly taken from Megson (1983) though the graben is observed on profile C 80-1 (around 0000/096, Fig. 5.4) and rather better on the nearby multichannel line WI-10 (SPs 1025 to 1100) where it is roughly 20 km wide and contains up to 0.4 s TWT of sediments, equivalent to a depth of 400 m assuming a Vp of 2.0 km/s. No trace of the graben is seen on the WI-8 seismic profile, roughly 30 km north-east of WI-10, therefore it is considered unlikely to continue beyond 53°N. Perhaps it terminates against the 53°N Flexure Zone of Bailey (1975)? Assuming it does end here then its total length is approximately 160 km, similar in size - though narrower -to the exposed parts of the Midland Valley of Scotland.

Although the few seismic profiles in this area do not permit useful contouring of the seismic basement of the graben, those that do cross it, and the basement contour chart of Megson (1983), indicate an increase in cumulative downthrow within the graben from c. 0.5 km near 53°W to 2 km or more near the mouth of the bathymetric re-entrant at 52°N. Several authors have proposed a close structural link between the graben and the location and development of the Charlie-Gibbs F.Z. and, before that, the Clare Lineament. Megson (1983) suggested that the graben system resulted from tensional stress reactivating Caledonian faults. Indeed this seems the most plausible explanation given the ubiquitous SW-NE trends of the many Caledonide structures around the Rockall region. Whether the graben can be followed into one of the major discontinuities on the U.K. mainland remains to be seen; the present author is pessimistic in this respect.

Whether the graben is a reactivation of an ancient crustal discontinuity or an independent younger structure it seems likely that the tensional forces that created it were also responsible for the other normal faults on Porcupine Bank (Fig. 5.2) - that is, the main rifting event which structured the Rockall Trough. It almost appears as if the graben formed in response to the tensional stresses attempting to follow a straight line, least path of resistance between the Clare Lineament and north Porcupine Bank. The orientation of the structural elements of the graben are similar to those of the nearby Porcupine Seabight as reported by Lefort and Max (1984) and Masson and Miles (Structure and development of the Porcupine Seabight sedimentary basin, in prep.); it is possible, therefore, that they are coeval.

The pattern of structural elements within the north-west continental margin of Rockall Trough, comprising Rockall and Lorien Banks, is rather less well known compared with the Porcupine Bank (Figs 5.1 and 5.2). This is because of both the poorer geophysical coverage and the complications arising from later lower Tertiary igneous activity around this north-west flank. The latter factor is a problem on the margin of Rockall Bank north of 55°N where apparently extensive lavas mask the underlying continental basement structure (Figs 5.5 and 5.6). Consequently the acoustic basement contours in this area, and also to an extent around the top of Lorien Channel (Fig. 5.1), do not properly represent true seismic (continental) basement and so should be interpreted with caution accordingly.

Between Rockall and Lorien Banks a bathymetric canyon with a relief of over 1000 m, the Lorien Channel of this work, trends at right angles to the north-west Rockall margin (Fig. 5.2). The Channel appears to be controlled by tectonic constraints and is not simply an erosional feature in an otherwise straightforward down-faulted continental margin - although erosion may be operating within it at the present. A close analogy can be drawn with the tectonically controlled canyon and graben beneath west Porcupine Bank, described above.

Three seismic profiles that run parallel to the Lorien Channel (C 84-4, RH 111 and WI-32) and one profile crossing it obliquely (C 84-5) each show presumed continental basement rising gently towards the north-west and also a number of short steep antithetic normal faults at the mouth of the Channel (Fig. 5.2). Two such faults appear on the multichannel profile WI-32 (Fig. 5.7) where the simplest interpretation is to infer downthrow towards the Rockall microcontinent, i.e. in the opposite (antithetic) sense to the standard down-to-the-basin normal faults elsewhere. The trend of the faults is difficult to determine with so few seismic profiles, especially when none of the tracks cross the channel in a perpendicular direction. It is probable that the two styles of faults are deflected to the north and north-west, away from the SW-NE trend that dominates the remainder of the margin.

An alternative interpretation is that the seismic profiles in fact represent highly oblique crossings of synthetic normal faults oriented roughly parallel to Lorien Channel. Neither interpretation is favoured over the other in the absence of further seismic reflection coverage. However, the occurrence of small antithetic (or counter) faults near the margins of extensional terranes is widely described in the literature (Fig. 5.8; Gibbs 1984) and therefore their appearance here is not unexpected.

The origin of Lorien Channel has not previously been discussed as part of the multiphase evolution of the Rockall area. The NW-SE trend of the Channel in a region where the SW-NE structural grain is

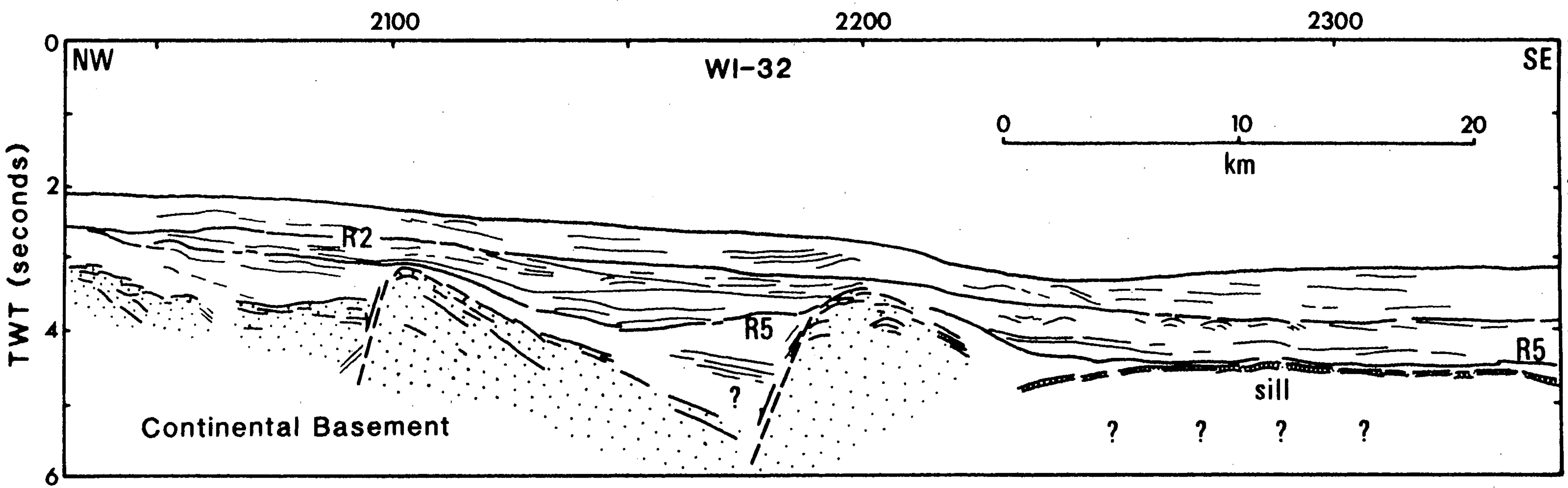


Figure 5.7 Interpretation of multichannel seismic profile WI-32, sps. 2050-2350. Refer to Fig. 5.1 for location. Original profile made available courtesy of D.G.Masson. Bold lines indicate important reflectors, lighter lines show form within seismic sequences. Note draping of sediments over faulted continental basement (stipple) to the west. ? = seismic basement not imaged beneath sill or lava flow.

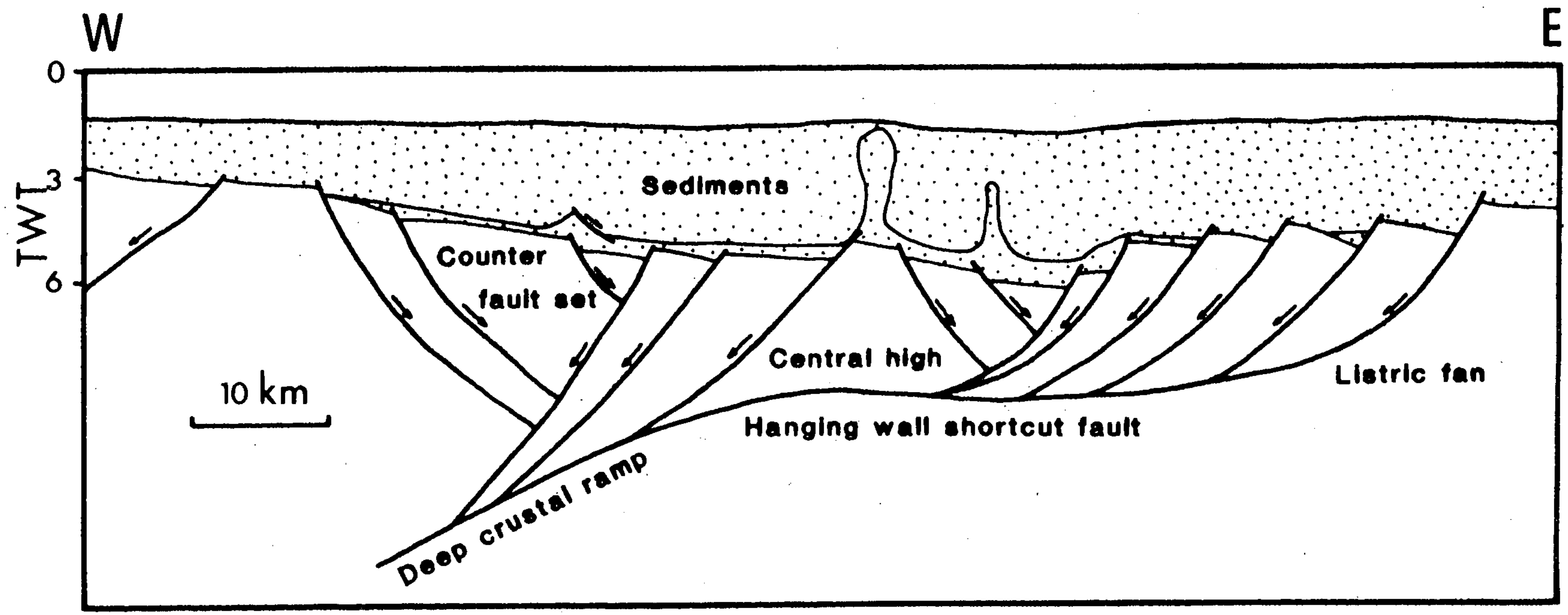


Figure 5.8 Schematic structural cross-section across the North Sea illustrating the occurrence of counter (or antithetic) faults at extensional basin margins. Redrawn from Gibbs (1984; Fig.1).

so influential and widespread is difficult to reconcile. This writer suggests that the structures and trends beneath the south-west Rockall Plateau, including Lorien Channel, may have been closely controlled by the position and early evolution of the two major oceanic fracture zones in this area - the Charlie-Gibbs F.Z. and the unnamed fracture zone to the north (55°N) below Edoras Bank (refer to Fig. 4.1). Such inferences of regional tectonic control and development are pursued further in Chapter 8.

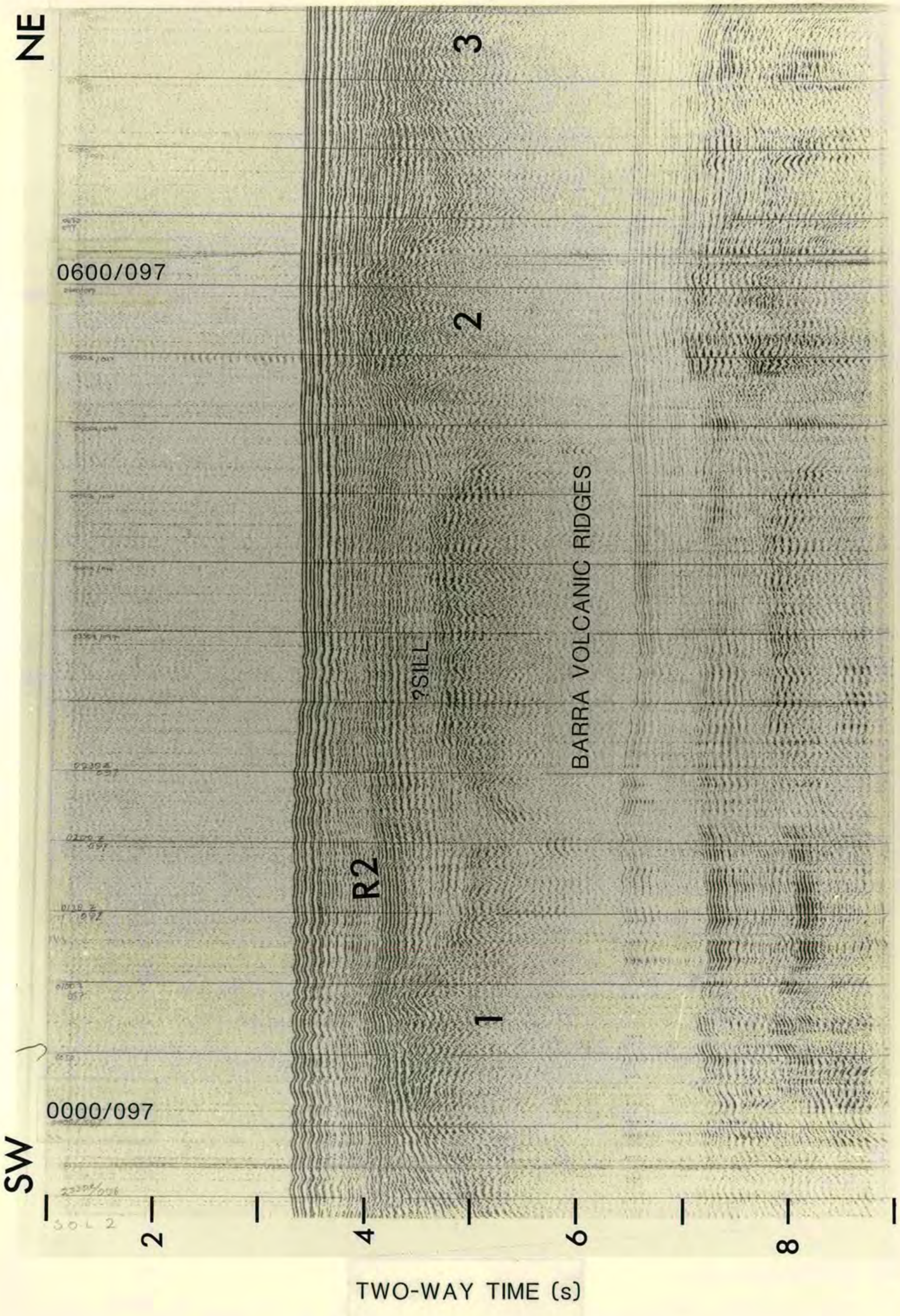
The Barra Volcanic Ridge System

In the southern Rockall Trough a previously undescribed complex of arcuate seismic basement ridges, the Barra volcanic ridge system (BVRS), has been mapped out from this work (Fig. 5.1). The presence of probable volcanic material in the Trough was first reported by Roberts et al. (1981) who outlined the distribution of supposed Palaeocene lavas in the region 53°N to 54°N, 16°W to 18°W. However, inspection of their seismic records, and correlations with nearby multichannel profiles, suggests that their intra-sedimentary lava sheet is composed largely, if not wholly, of intrusive sills (see section 5.3). This author has re-interpreted the single-channel seismic profile of Figure 6 of Roberts et al. (1981) and believes the feature they depicted as a lava sheet is in fact a basement ridge similar to and continuous with the one they picked a short distance to the south-west. The best evidence in support of this is the drape and onlap of the sediments in the flanking basins onto the proposed basement ridges.

The appearance of the ridges of the BVRS on the seismic reflection profiles is usually quite distinct. On the single-and double-channel profiles, because of the strong vertical exaggeration (about 6 times), the ridges have a typical peaked aspect (Figs 5.3, 5.4, 5.10 and 5.11). This is particularly noticeable on the S 79-14 profile (Fig. 5.3) from 2300/096 to 1200/097. On the same profiles the ridges are seen to consist of a thick layer of short, strong irregular reflectors which absorbs much of the acoustic energy, such that few deeper events are ever recorded (Fig. 5.9). This highly reflective layer is replete with diffraction hyperbolae which account for the crenate appearance of many of the reflectors on the line drawing interpretations.

Figure 5.9.

Section of Challenger 80-2 seismic profile across the BVRS



LINE DRAWING INTERPRETATION OF CHALLENGER 6/80 LINE 2

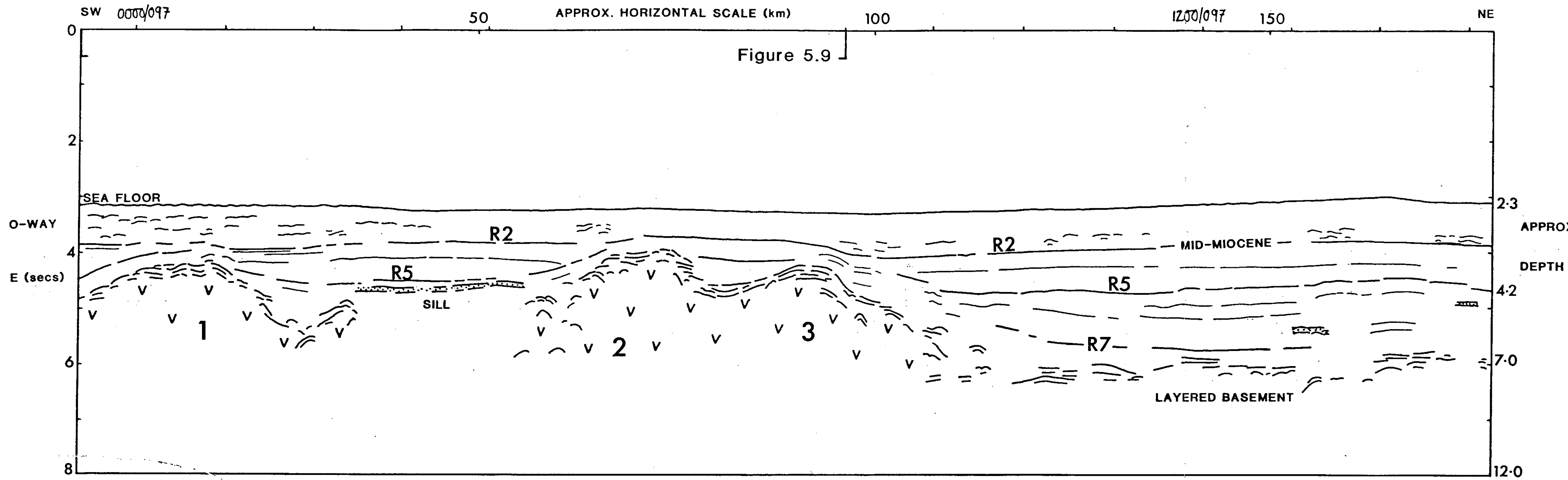


Figure 5.10 1,2 and 3 denote three Barra volcanic ridges, marked by V symbol. Refer to Fig. 5.1 for location

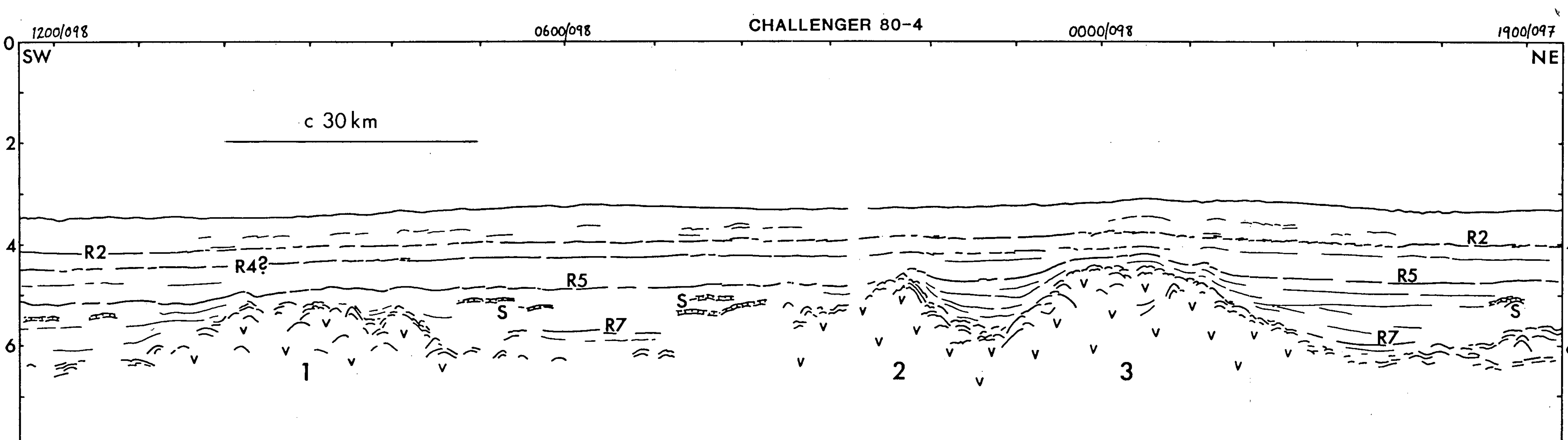


Figure 5.11 As for Fig. 5.10. S denotes sills in sediments

On the improved quality multichannel profiles that cross the BVRS, such as WI-32, CM-03 and CM-04, the diffraction interferences are largely absent and it is possible to observe that the true seismic character of the ridges is one of numerous overlapping short reflectors of highly variable dip (see, for example, profile CM-04 north of SP 3400, Fig. 4.5). Occasionally a more persistent reflector can be picked within, or a short depth below, the topmost reflective layer; but generally the internal structure of the ridges is poorly imaged. Hence it appears that the abundant diffraction events, which obscure the basement structure on the simple double-channel seismic records, are originating at the abrupt terminations of the irregular and discontinuous reflectors picked on the multichannel records. Interestingly this author has found little convincing evidence for faulting within the ridges.

The rapid interlayering of short reflectors, together with their high reflectivity, are a good indication that the rocks constituting the BVRS are largely extrusive in origin, hence the descriptive title. It is still open to speculation as to whether the presumed extrusive rocks (almost certainly lava flows) are interbedded with sediments or not, and if so to what extent. On most of the profiles inspected which traverse the BVRS the sedimentary fill to Rockall trough appears to drape over the rough surface of the volcanic ridges: this is especially marked by the drape and thinning of the pre-R5 sequence over these seismic basement highs (Figs 5.10 and 5.11). There are other instances, on the other hand, where the situation is much less clear cut. On profiles CM-04 and WI-32, for instance, it is sometimes difficult to determine where the interface is between the base of the sediment column and the top of the volcanic basement. This is particularly true of the deep pre-R7 succession (and occasionally the R5-R7 interval) which tends to be strongly layered and gives the impression of gradual interbedding of lavas and sediments. However, this hazy zone is restricted to the lower flanks of the ridges and the intervening troughs.

Little evidence is available regarding the geometry and structure of the volcanic ridges below their strongly reflective rough upper surfaces. On none of the double-channel seismic profiles is there a clear relationship between the deeper flanks of the ridges and the surrounding deep layered basement. On profiles S 79-14, C 80-1 and C 80-2 (Figs 5.3, 5.4 and 5.10) there is an ambiguous zone of chaotic or, more often, very weak reflectors at the transition between the ridges and the deeper flatter seismic basement (see, especially, 0900 to 1000/097, profile C 80-2). This confused zone makes it impossible to state, unequivocally, whether the ridges rest on top of a continuous deep layered event or on top of some thickness of old sediments.

Nevertheless, on profile C 80-4 (2000 to 2300/097, Fig. 5.11) the disposition of the reflectors is such as to suggest that even the deepest sediments onlap onto the flanks of the ridges, with no strata persisting below them. This is supported by the interpretations of the WI-32, CM-03 and CM-04 multichannel profiles which indicate continuity between the crests of the ridges and the start of the deep layered basement. Indeed on the WI-32 record, which passes obliquely over the BVRS and continues onto the continental margin below west Porcupine Bank, there is no noticeable change in seismic character between the volcanic ridges and the contiguous deep layered basement, other than that of topographic relief. This leads the present author into thinking that these two types of seismic basement are in fact made up of the same material - lava flows - and that the BVRS is merely a region of deep layered-type basement with enhanced relief, reflecting more voluminous extrusive activity.

The rather fortuitous location of Hill's (1952) three refraction stations over ridge 3 of the BVRS (Fig. 5.1 and profile C 80-2, Fig. 5.10) provides some indication of the depth to the base of the volcanic ridges and the local P-wave velocity structure. The depths and velocities of the main refractors below the three stations (Fig. 5.12) illustrate the thinning of a 4.94 km/s layer to the south-west and this corresponds to the crest and flank of ridge 3 on the seismic profiles (e.g. C 80-2). Hill (1952) calculated a maximum thickness of 3.38 km for this layer beneath station 3 and placed its base at between 7.7 and 7.9 km. A similar thickness can be computed from the seismic reflection profiles by performing approximate conversions of two-way travel time to depth, as indicated at the end of track C 80-2 (Fig. 5.10). The depth of 7 km calculated for the flat deep layered basement on this profile is significantly shallower than the base of the volcanic layer as reported by Hill. This disparity could be due either to down-warping of the underlying basement owing to the large weight of the volcanic ridges, or to a velocity inversion beneath the ridges caused by a layer of low velocity sediments. Alternatively

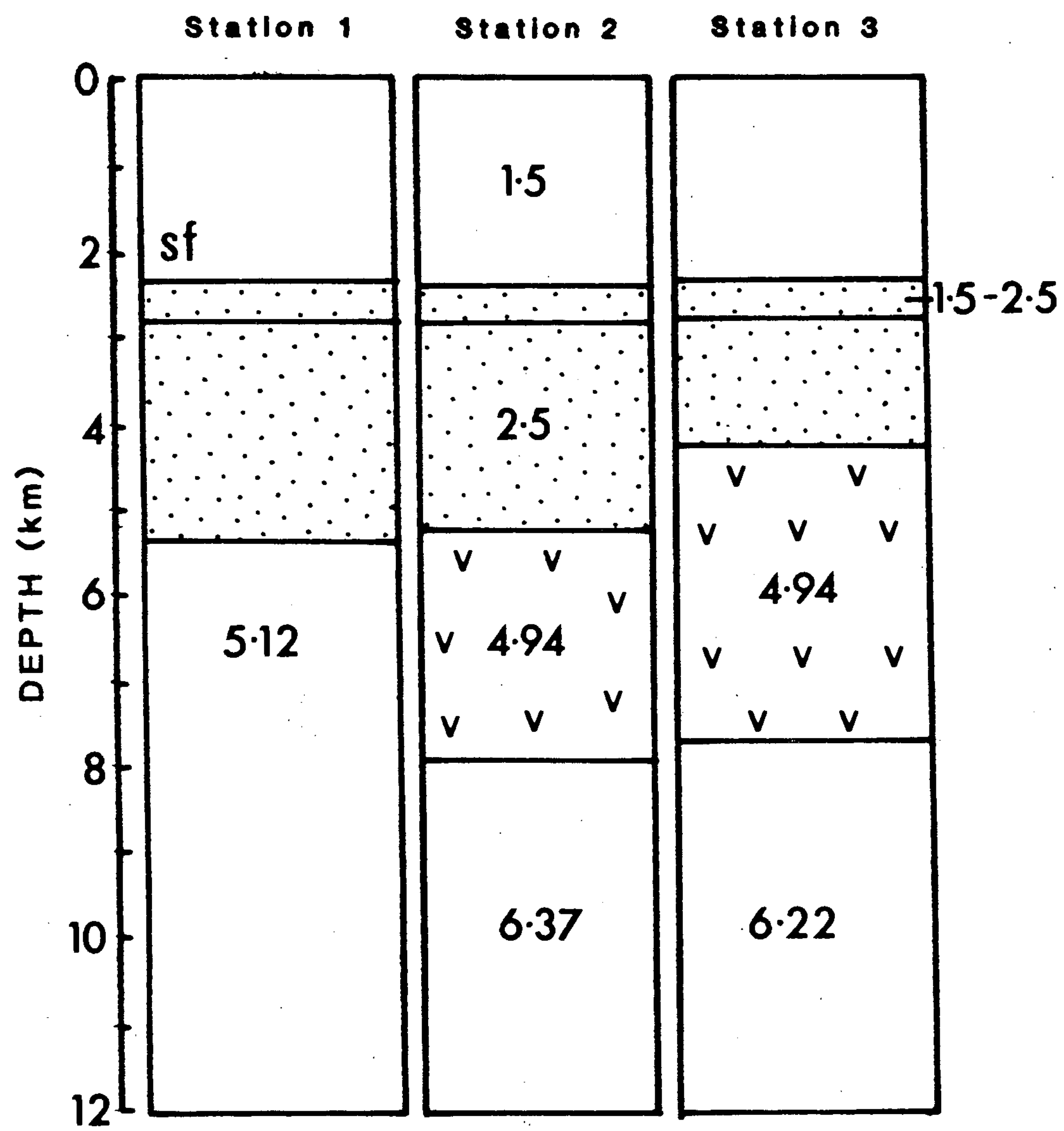


Figure 5.12 Hill's (1952) three refraction stations recorded in southern Rockall Trough. See Fig. 5.1 for location of stations. Stipple = sediments; V = volcanic basement. Numbers in layers are approximate compressional wave velocities in km/s.

Hill's 7.7-7.9 km basal refractor may persist across the whole area such that the deep layered basement of this work is simply a thin extrusive skin over a deeper geological basement - Layer 3 of oceanic crust for instance.

The refraction data indicate a velocity of 4.94 to 5.12 km/s for the volcanic layer (=BVRS). This is the typical velocity range for ocean-type (tholeiitic) basalt, an observation which supports this author's interpretation of the BVRS as volcanic edifices from the seismic reflection profiles. If the volcanic ridges contained a sizeable percentage of sediments interbedded with the lavas then the refraction velocities would be expected to work out somewhat lower than the whole basalt velocities computed by Hill (1952; Fig. 5.12). Again this is in support of the reflection data which favour a largely volcanic composition for the ridge complex. That Hill arrived at much the same conclusions in 1952 concerning the composition and regional significance of this basaltic layer, with so little local geological information then available, is tribute indeed to his astuteness and innovation.

The BVRS, as mapped out here (Fig. 5.1), forms a continuous complex of ridges stretching from the Charlie-Gibbs F.Z. northwards to the Rockall Plateau. Three main ridges are evident from the acoustic basement chart (see Chart 4, back pocket) - 1, 2 and 3 -with a further short segment, 1b, being poorly constrained. The manner in which ridges 1 and 2 curve abruptly into the northern ridge of the CGFZ was discussed at length in Chapter 4. North of the fracture zone these two ridges, and also 1b, parallel each other in a NNE orientation for roughly 80 km. This pattern of straight, parallel, peaked ridges (see S 79-14, Fig. 5.3) is reminiscent of that observed about mid-ocean ridges, an analogy that is strengthened by the associated lineated magnetic anomaly signature (see Chart 3, back pocket).

At approximately 53.3°N ridges 1 and 2 change direction to trend NW-SE; this change is quite sudden for ridge 1. Ridge 3 immediately to the north-east assumes a similar trend and forms a complete barrier across Rockall Trough from 54.4°N on the Rockall Bank margin to 53.3°N on Porcupine Bank, a distance of over 200 km. Together the three ridges trend north-west onto Rockall Bank where the distinction between volcanic and continental basement on the seismic profiles becomes harder to establish, hence the rather confused basement

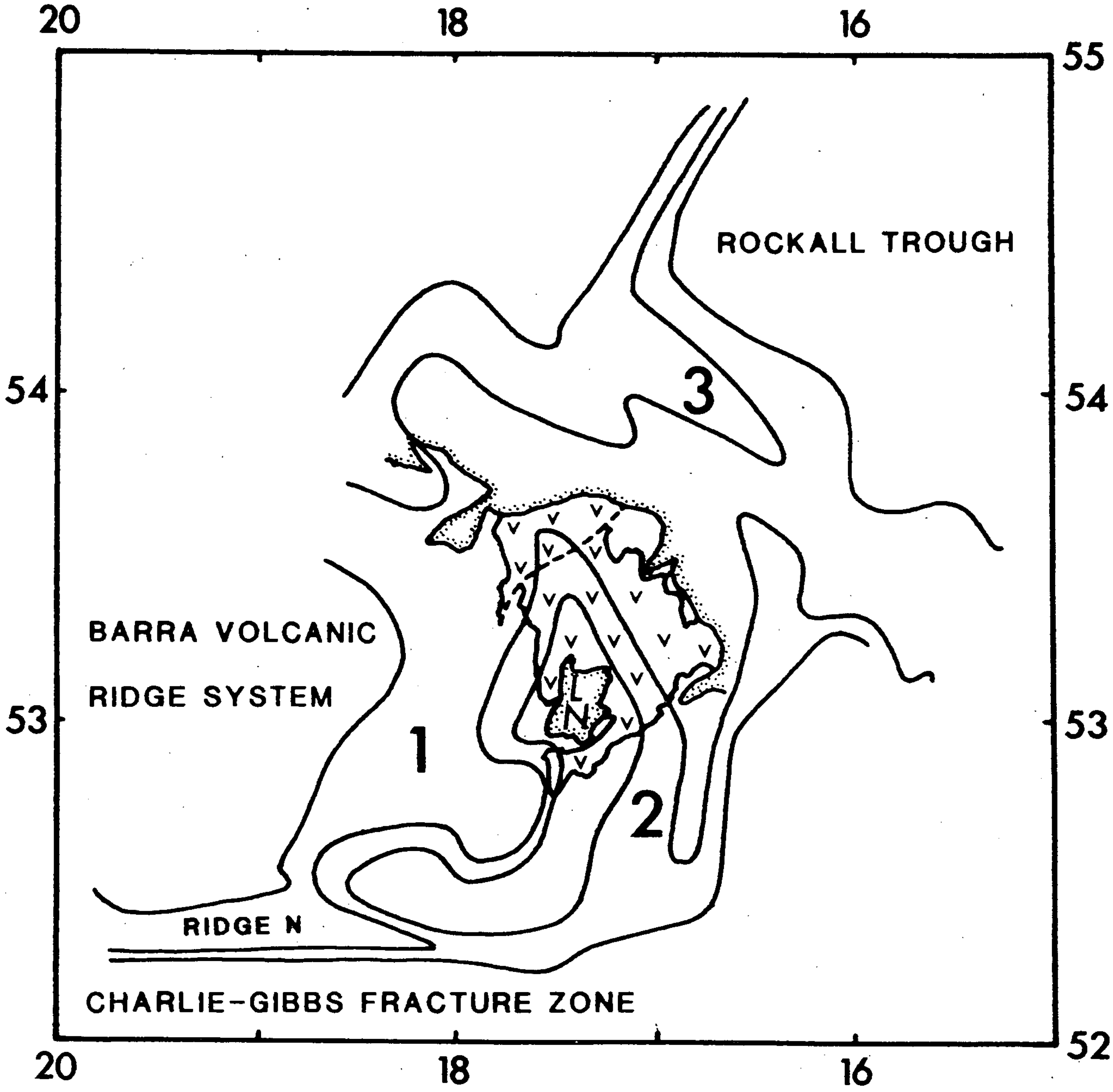


Figure 5.13 Scale comparison between the areal extent of the Barra volcanic ridges (1, 2 and 3) of this study and that of the Antrim plateau basalts (V ornament). Northern Irish coast stippled for clarity. LN = Lough Neagh.

isochrons in this area (Fig. 5.1). The coincidence in location and trend (colinearity) of the northern BVRS and Lorien Channel suggests that they have some common origin.

Earlier a comparison was made between the southerly ridges of the BVRS and those reported at and near the mid-ocean ridges. No suitable analogy has been found with any existing continental structures. The large areal extent of the BVRS and the moderately large aspect ratio (width: thickness of the body) of the separate ridges -in the order of 40:3 km - are not typical of igneous features in the continental realm. The extensive plateau basalts of India and Africa are of the correct composition and scale but are unlikely to exhibit the relief reported here for the BVRS. The BVRS is considerably bigger, for example, than the Antrim plateau basalts of Northern Ireland (Fig. 5.13) or the Caledonian intrusive province of the Grampian Highlands, Scotland. The appearance of the BVRS on the seismic profiles and its aerial distribution do not support the idea, proposed by Megson (1983), that the component ridges are large igneous intrusions, although these may be present at greater depths below the BVRS. Thus the initial likening to the sort of large scale processes that occur in the oceanic realm is still considered most realistic.

Deep layered basement in the Rockall Trough

Between the continental margins of Rockall and Porcupine Bank and away from the Barra volcanic ridges the floor of the Trough is characterised by a deep and generally even layer of short, strong, though often irregular, reflectors (profiles GSI-1 and NA-1, Figs 5.5 and 5.6). In many respects the seismic appearance is similar to that of the BVRS, the main difference obviously being the absence of such marked relief. Naturally then the same considerations apply regarding the probable volcanic nature of the deep seismic basement as they did for the BVRS. Comparisons can also be made to the oceanic basement west of southern Porcupine Bank and Goban Spur, though the deep seismic basement in the Trough differs through being more discontinuous and lacks the regular fault scarps observed in the former areas.

On the double-channel seismic profiles the layered basement is everywhere poorly imaged and typically appears as an impersistent band of short reflectors and numerous diffractions, both of variable intensity. This shadowy aspect probably results from its occurrence at considerable depths (over 5 km in places) below thick sediments and also because of its irregular top surface. Where the overlying sediments are thickest, notably below Feni Ridge, the layered basement is not usually imaged as, for example, between 1500 and 1800/097 on profile S 79-14 (Fig. 5.3).

Along the better quality multichannel reflection profiles the seismic character of the layered basement is much the same as described above, though the confusing diffraction events are generally absent. The relationship between the basement and the deepest sediments varies markedly from one profile to another. On track GSI-1 (Fig. 5.5) the top of layered basement in the Trough (SPs 5100 to 5350), while being irregular on a small scale, is well defined and clearly separable from the suprajacent well layered pre-R7 succession. In contrast, the deep layered basement below profile NA-1 (Fig. 5.6) is noticeably rougher and its top surface is difficult to distinguish from the highly reflective thin pre-R7 sedimentary drape. The impression gained is that of interlayering of lavas and sediments within the upper parts of the deep layered basement. This point is well illustrated between SP 750, NA-1 and SP 100, NA-1 Ext. (Fig. 5.6), and also seems to apply to the north-western half of profile GSI-1 (SPs 4600-4900) although the seismic basement is not well seen here.

The lava flows, and any trapped sediments, that constitute the deep layered basement do not dip consistently in any one direction. On profile GSI-1 the short basement reflectors are mainly flat-lying, whereas on NA-1 and NA-1 Ext. there is a tendency for the lavas to occur in shallow overlapping mounds. Both styles are compatible with fissure-type eruption within the Trough. In addition, on profile GSI-1 gently rounded ridges or domes occur in the layered basement at SPs 5120, 4950 and 4770. The double-ridged feature between SPs 4880 to 5000, the largest one mapped out north-east of the BVRS, is approximately 20 km across and rises c.1.5 km above the surrounding layered basement. Its near central location within the Trough and the roughly symmetrical disposition of the smaller ridges on either side are rather reminiscent of the geometry and bathymetric relief encountered at mid-ocean ridges - a similitude that was earlier mooted for ridges 1 and 2 of the BVRS. The seismic basement contours for the central Rockall Trough (Fig. 5.1 and Chart 4) highlight a grain within the layered basement, i.e. ridges and troughs, that is parallel to the basin margins. Unfortunately the seismic reflection coverage here is sparser than over the southern parts of the Trough so the isochrons are not closely constrained. Some measure of support for these basement trends is provided by the presence of like-trending (SW-NE) positive magnetic anomalies overlying the basement highs on GSI-1 and other profiles.

The junction between the deep layered basement and the base of the continental margin is with few exceptions vague and confusing, as exemplified by profiles S 79-14, C 80-1 and NA-1 (Figs 5.3, 5.4 and 5.6). The geometry of the basement and sedimentary reflectors beneath the base of the Porcupine Bank continental margin observed on these three profiles is fairly representative of the remaining tracks crossing it. Typically the layered basement deepens appreciably and becomes weaker and less layered a short distance before the front of a down-faulted and tilted block of continental basement. On the three profiles illustrated a gap exists in the seismic basement between the diffuse edge of the deep layered basement and the continental block. It is uncertain therefore whether there is an abrupt vertical contact between the two types or if the volcanic basement tapers out onto the flanks of the fault blocks. In either event a rather characteristic deep triangular-shaped sedimentary basin is often observed over this junction below Porcupine continental rise.

A notable exception to this pattern is observed at the south-eastern end of profile GSI-1 (SPs 5300 to 5450, Fig. 5.5). Here there is no evidence for a rotated fault block beneath the base of the continental slope. Instead the deep layered basement appears to continue upwards from the floor of the Trough until it overlaps what is probably the crest of a fault block midway up the slope. Again the surface separating layered basement here from the overlying highly reflective pre-R7 sediments is difficult to define clearly. If a local source of eruption for the presumed lava flows was situated midway up the continental slope then a continuous slope of lavas might be expected to result, with infilling perhaps of any pre-existing half-grabens.

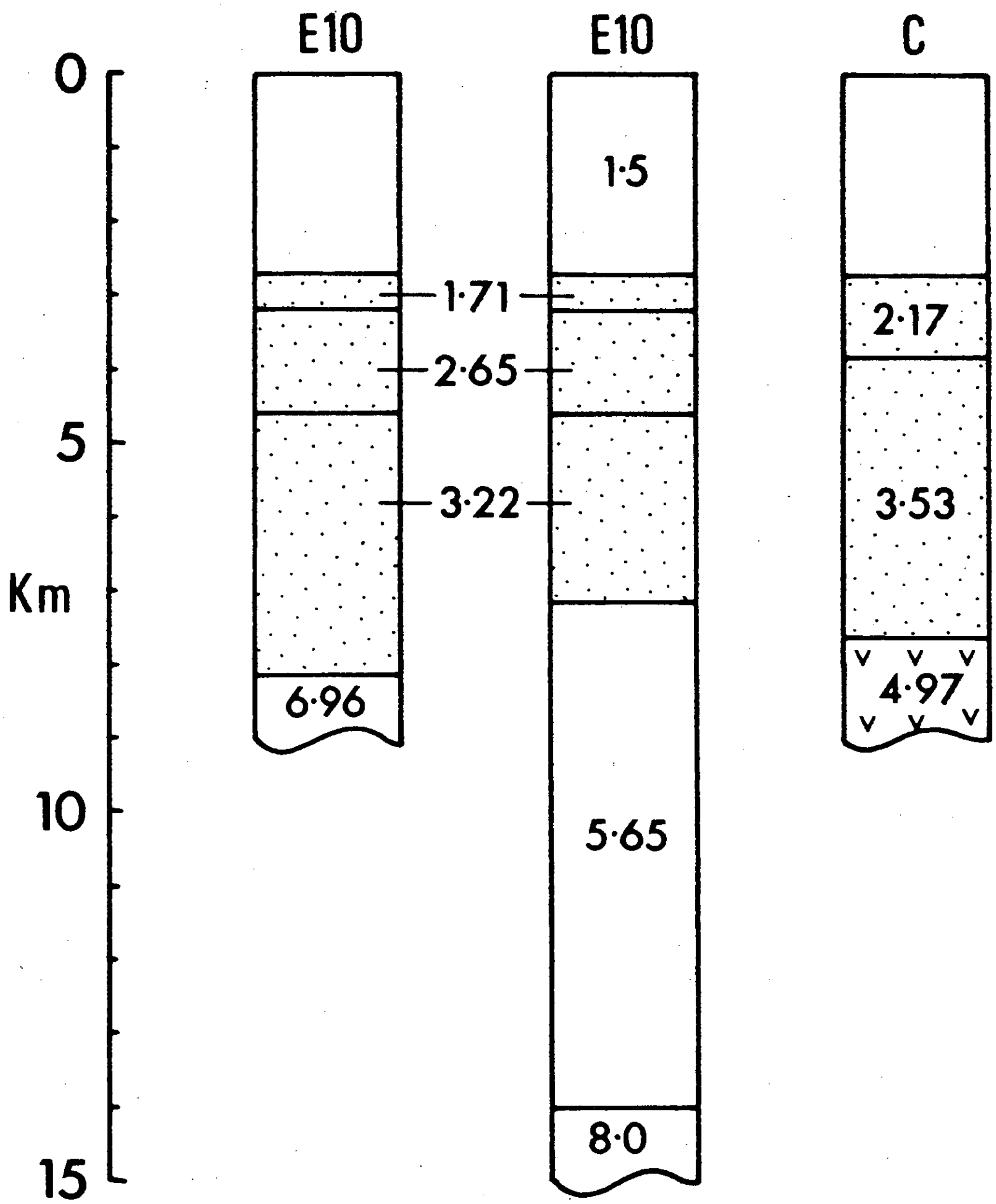


Figure 5.14 Left: E10 refraction station of Ewing and Ewing (1959). Stipple = sediments. Numbers are P-wave velocities in km/s. See Fig. 5.1 for location.

Centre: Reinterpretation of E10 refraction station by Jones et al. (1970).

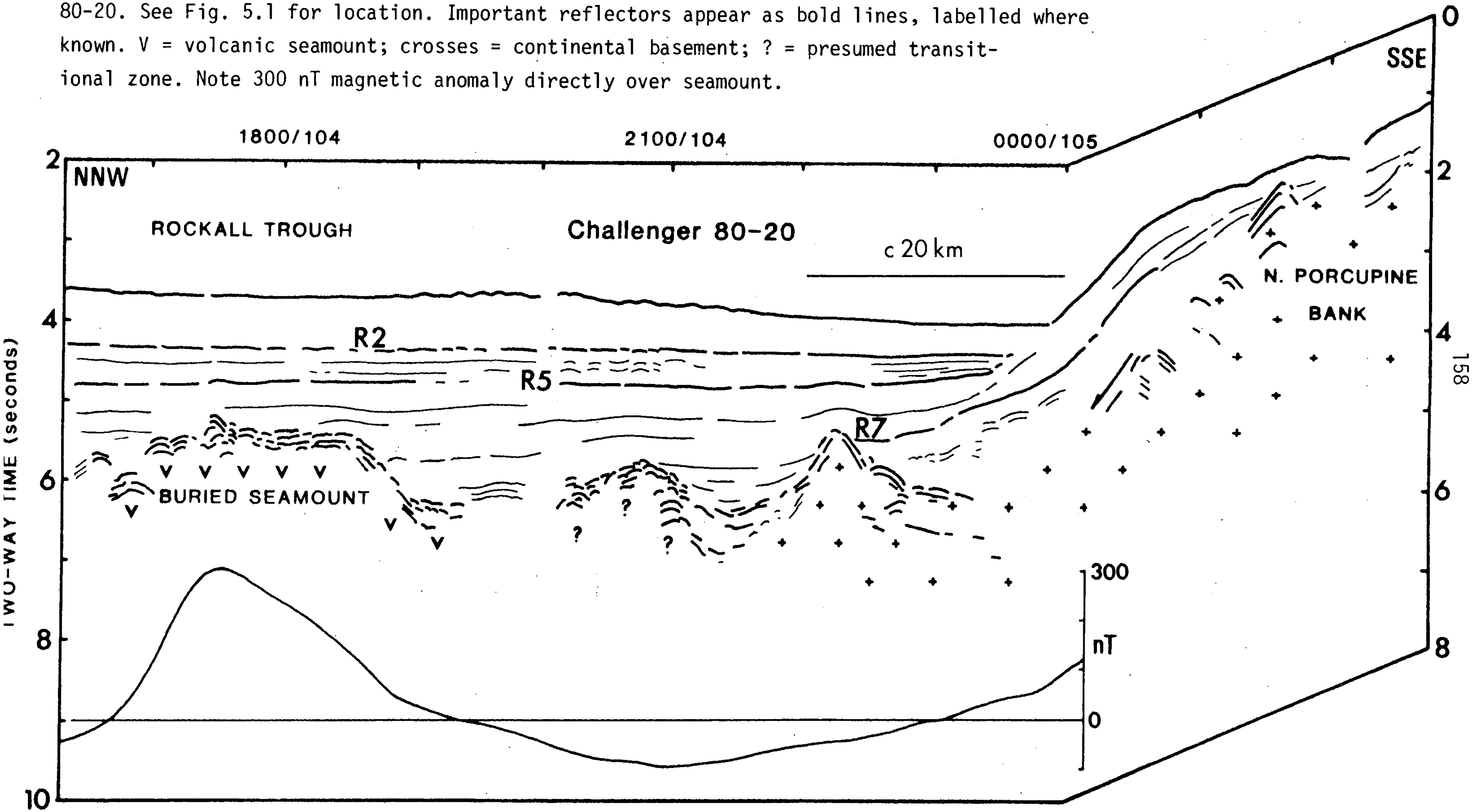
Right: C refraction station of Scrutton (1972). See Fig. 1.6 for location. V symbol = volcanic basement.

Beneath the north-west margin of the Trough the relationship between deep layered and continental basement is obscured further by the continuation of the Barra volcanic ridge system in the south onto the Rockall Bank and by the occurrence of intra-sedimentary lavas straddling the margin to the north (Figs 5.1 and 5.6).

Two refraction stations within the Rockall Trough, apart from those of Hill (1952), have some bearing on the depth to the top surface of the deep layered basement (Fig. 5.14). Ewing and Ewing (1959) gave two interpretations of their short unreversed refraction line E10, which was also re-interpreted by Jones et al. (1970). The latter interpretation which places a 5.65 km/s refractor at a depth of 7.14 km is favoured here as this would correlate closely with profile NA-1 (SP 1000, Fig. 5.6), roughly 25 km to the north-east. After converting the depths of the refractors to two-way travel time using the calculated velocities in Figure 5.4 it is seen that the 3.22 and 5.65 km/s refractors occur at 5.23 and 6.81 seconds TWT, respectively. On the nearby NA-1 profile R5 is a prominent reflector at 5.25 seconds and seismic basement first appears at 6.8 seconds. Presumably there is a velocity inversion at the base of the igneous sill at 5.8 s TWT on the same profile but this does not seem to have any noticeable effect on the correlation.

A depth to deep layered basement of 7.0 km or slightly more is commonly found when applying the sediment refraction velocities of E10 to the seismic reflection profiles. D.K. Smythe (pers. comm.) calculated depths to the same deep basement in the vicinity of E10 at a little under 7.0 km, or about 6.0 km when stripped of the loading effects of the suprajacent sediments. Scrutton's (1972) 4.72 km/s basement refractor at 7.6 km (Fig. 5.14) was recorded in the north-eastern parts of Rockall Trough where the seismic basement is known to be generally deeper and the sediments correspondingly thicker. Ewing and Ewing (1959) calculated an average of 5.11 km/s for the basement Vp at all their N. Atlantic stations; the average was reduced to 4.96 km/s if two dubious stations, including E10, were omitted. This ties in well with the basement velocities of 4.94 and 4.72 km/s found, respectively, by Hill (1952) and Scrutton (1972) and is strong evidence that the deep layered basement, like the BVRS, is composed of basaltic material.

Figure 5.15 Seismic interpretation and total intensity magnetic anomaly along profile Challenger 80-20. See Fig. 5.1 for location. Important reflectors appear as bold lines, labelled where known. V = volcanic seamount; crosses = continental basement; ? = presumed transitional zone. Note 300 nT magnetic anomaly directly over seamount.



At 54.6°N 14.7°W in the Trough a crudely oval-shaped high is contoured on the acoustic basement isochron chart (Fig. 5.1) and is interpreted as a seamount by virtue of its flat-topped seismic appearance and central location within the Trough (Fig. 5.15), and also its coincidence with a steep oval-shaped 400 nT magnetic anomaly (Chart 3, back pocket; Roberts and Jones 1975; Max et al. 1982). As with the volcanic ridges of the BVRS it is uncertain whether the seamount sits on sediments or deep layered basement, or perhaps is equivalent to the latter but has a more conspicuous relief. On the two Edinburgh seismic profiles that cross the seamount, C 80-20 (Fig. 5.15) and C 84-7 (Fig. 5.21), the relationship between it and the deeper basement is never clear.

The appearance of small positive magnetic anomalies near the centre of Rockall Trough at 55°N 13°W and 56°N 11.7°W, similar to that over the seamount illustrated here, suggests that buried seamounts are also present at these positions. In the north of Rockall Trough the Hebrides Terrace, Anton Dohrn and Rosemary Bank Seamounts, with their characteristic bathymetric expressions, complete a SW-NE trending chain of six axially-located seamounts spanning 5° of latitude. The complicated magnetic signatures of the three above-named seamounts are believed to arise from the occurrence of abundant Tertiary lavas and sills in the surrounding sediments. The chain of six seamounts at or near the axis of the Rockall basin, together with the weakly lineated grain of the adjacent deep seismic basement, are good indicators to some form of nascent oceanic accretionary process having been in operation. Fuller discussion of such inferences is deferred until the final chapter.

5.2 The sedimentary infill to Rockall Trough

The general distribution of basement depressions within the Rockall Trough is outlined in the seismic basement travel time chart (Fig. 5.1; Chart 4, back pocket). Chart 5 shows isochrons for the total thickness of sediments measured along the 8600 km or so of seismic reflection profiles over southern and central Rockall Trough. The irregular, locally sparse, seismic coverage and the absence of good control on the P-wave velocities of the sediments do not permit either of these charts to be realistically converted to true depth or thickness. Nevertheless it is expected that the important basin trends will be little affected by this shortcoming.

Despite the fact that the Rockall Trough presents itself as a straightforward deep basin on the bathymetric charts the contours for total sediment thickness are, at first, surprisingly complicated. They reflect the complex interplay between the NW-SE to N-S trending Barra volcanic ridge system, the SW-NE trending Rockall Channel as a whole, and more particularly the prominent Feni Ridge sediment drift. Consequently there are few resemblances between the isobaths on the bathymetric charts and the sediment thickness isochrons of Chart 5 inside Rockall Trough. Bounding the entrance to the Trough in the south the Charlie-Gibbs Fracture Zone and Clare Lineament have clearly strongly influenced the comparatively thin sedimentary cover here; this is particularly obvious in the east-west orientation of ridge S of the CGFZ and the overlying sediment isochrons. Likewise to the south, over L.Cretaceous oceanic crust, the sedimentary basins mirror the NNW-SSE trend of the basement scarps and troughs (Fig. 5.1).

Three major depocentres, A, B and C (Chart 5) are identified in the Trough that correspond to similar-shaped depressions in the subjacent seismic basement. Basins A and B contain extensive areas of sediment thicker than 2.5 s TWT, or 3.0 s TWT in the case of B, while basin C is mostly thinner than 2.5 s TWT. A further depocentre D is similar in extent and thickness to C but is remarkable in that it sits over the main ridges of the BVRS. This configuration is made possible because of the build-up, by differential deposition, of the Feni Ridge drift against a large stretch of the north-west margin of southern and central Rockall Trough (illustrated well along profile C 80-1, Fig. 5.4).

That the BVRS has influenced the pattern of deposition in the Trough since its inception is indicted by the way in which the contours for total sediment thickness and various interval thickness (Figs 5.16 to 5.20) tend to follow those of the underlying ridges. This influence also persists up to the present sea floor where it results in the south-easterly bulge in the Feni Ridge isobaths around 53.5°N 17°W; it would be an unexpected fortuity if this was not the case. This evidence for control on the distribution of sediments by some pre-existing structure reinforces this author's belief in the volcanic nature of the BVRS. If the three ridges were of intrusive origin, as intimated by Megson (1983), it would be difficult to explain the unequivocal proof of draping and moulding of sediments around them.

In the central parts of the Trough, north of Porcupine Bank, the large basin B appears to consist of two equally-sized depocentres, the one to the north-east being offset to the south-east (Chart 5). This divided, though erroneous, basin outline is brought about by the projection of a lobe of thin sediments around the supposed seamount (54.5°N, 14.7°W) towards what appears to be a platform area beside the Rockall Bank margin. The abrupt south-eastern termination of the latter feature as shown on profile NA-1 Ext. (Fig. 5.6) has been described as the Jean Charcot Fault Zone in the past (Pitman and Talwani 1972, Roberts 1975). But the even, persistent reflector marking the top of the platform is now strongly believed to be a well marked lava flow, by analogy with the Palaeocene and Early Eocene lavas reported around the Faeroe-Shetland region (Roberts et al. 1983; Smythe 1983, and pers. comm.). Hence the contours shown on Chart 5 represent the sediment thickness on top of the lava, while beneath it the deeper sediments probably continue for about another 30 km.

The BVRS acts as an abrupt and effective barrier between the three main basins, A, B and C. The large thick basin at A results largely from the presence of the Feni Ridge and there is a marked contrast with the thinly sedimented ridges of the BVRS immediately to the east (Chart 5). Away from the influence of the Feni Ridge the sedimentary column is, not surprisingly, noticeably thinner as in basin C west of Porcupine Bank - despite the extensive areas of deep basement here. The lateral changes caused by the varying influences of the Feni Ridge and BVRS are well illustrated by the 285 km long S 79-14 profile (Fig. 5.3).

Some measure of the considerable thicknesses of sediment in parts of the Rockall Trough is afforded by the limited seismic refraction data, as mentioned in the latter part of the previous section. The E10 refraction station of Ewing and Ewing (1959; Fig. 5.14) indicates that the R5 reflector of this work is also a strong refracting horizon, separating sediments above with a Vp of roughly 2.65 km/s from those below with a velocity of 3.22 km/s. If these interval velocities are applied to the NA-1 profile (Fig. 5.6), along which the deepest basement is observed in the current study area, then maximum sediment thicknesses of 4.81 and 4.87 km are calculated for shot points 600 and 750 respectively, corresponding to basement depths of 7.72 and 7.85 km. These are in good agreement with the 4.9 km sediment thickness reported for refraction station C by Scrutton (1972) in the northern Rockall Trough (Fig. 5.14). The two interpretations of station E10 give total thicknesses of sediment at either 5.39 or 4.39 km of which the latter is favoured here (section 5.1).

To the south-west, along the GSI-1 profile, total sediment thicknesses of 4.56, 4.43 and 4.72 are arrived at for shot points 4850, 5000 and 5075. Here the seismic basement is shallower than to the north-east but this is compensated for by the influential Feni Ridge drift. Thus the greatest development of sediments occurs towards the north-eastern parts of this research area, the south-western end of the main Rockall Trough depocentre, where total thicknesses attain a maximum of about 5.0 km, these generally being restricted to the western flank of the Trough, as indicated by the 3.0 s TWT isopach on Chart 5.

At the other extreme areas of thin sedimentation are evident over the BVRS, Charlie-Gibbs F.Z. and, in particular, beneath the shelf break and continental slope. Along the S 79-14 seismic profile the irregular, layered sediments draping the peaks of ridge 1 of the BVRS may be as thin as 200 m or less (0800/097, Fig. 5.3). Beneath the continental margins the sediments soon thin landwards to about 0.5 km or less, as outlined crudely by the 0.5 s TWT isopach (Chart 5). Around the Porcupine Bank margin a few basement horsts or crests of fault blocks are almost or completely absent of sediment, as on profile GSI-1 (Fig. 5.5). Below the north-west margin of the Trough the same pattern of thinning over continental fault blocks is observed but, in addition, the seismic profiles display a small number of sharp bare peaks which are thought to be of volcanic (Early Tertiary?) origin.

Pre-R7 seismic sequence

An isochron chart was contoured showing the distribution and thickness of the R7 to seismic basement sedimentary interval in the main part of the research area (Fig. 5.16). Owing to the poor resolution of the R7 and basement reflectors on most of the single-

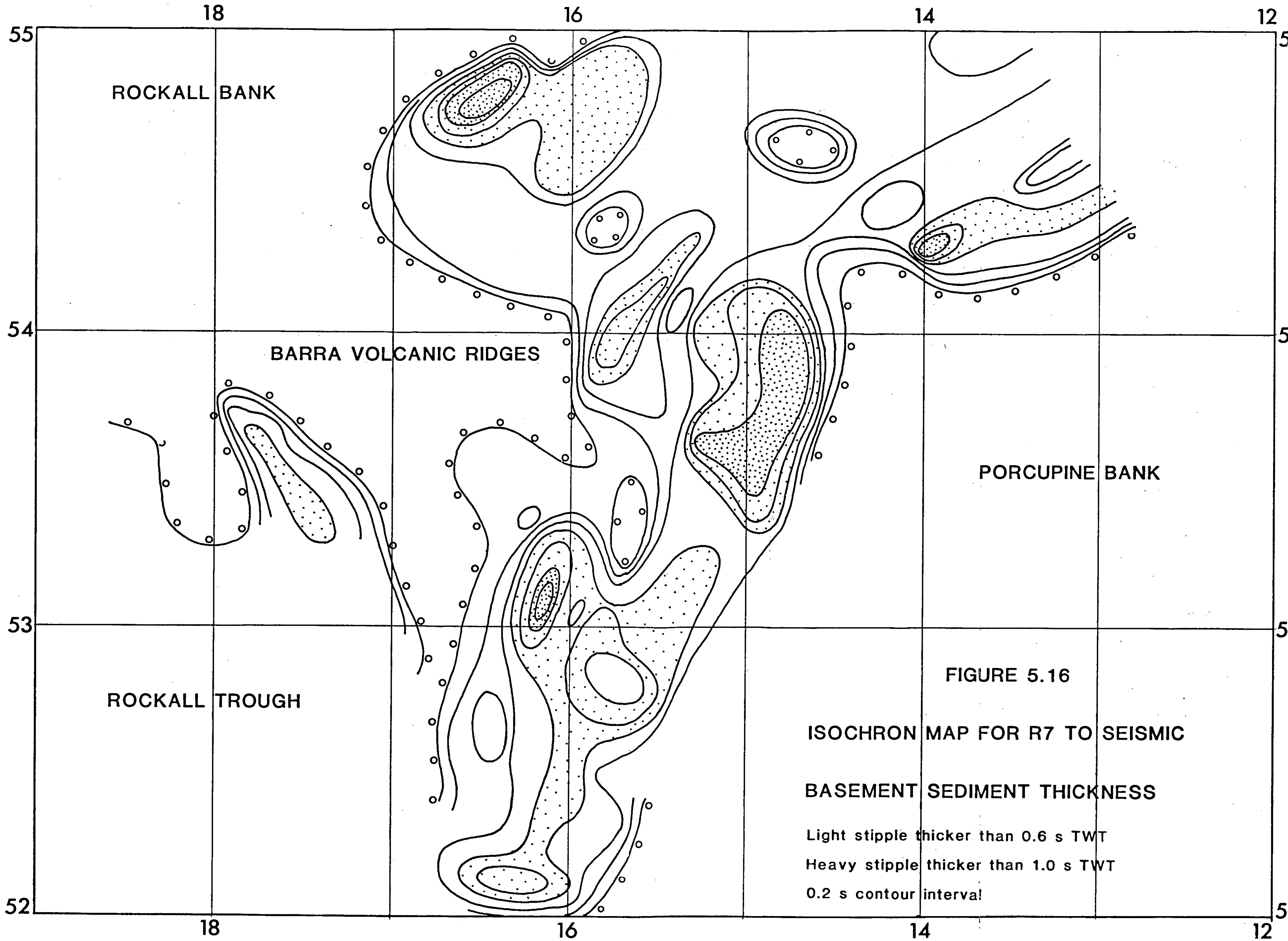


FIGURE 5.16

ISOCHRON MAP FOR R7 TO SEISMIC BASEMENT SEDIMENT THICKNESS

Light stipple thicker than 0.6 s TWT

Heavy stipple thicker than 1.0 s TWT

0.2 s contour interval

and double-channel seismic profiles, and even on some of the multi-channel records, this contour chart is somewhat schematic and subject to fairly heavy interpolation in parts. Nevertheless it highlights the thickest development of the pre-R7 sequence in two broad troughs adjacent to the Rockall and Porcupine Bank margins. Thicknesses in these depocentres occasionally exceed 1.2 s but values of 0.4 to 0.8 s are more usual. On the NA-1 Ext. profile the pre-R7 sequence is masked by younger lavas near the continental margin (Fig. 5.6), though it seems likely that a thick succession is developed comparable to that observed on the GSI-1 profile to the south-west (Fig. 5.5).

Towards the axis of the Trough the pre-R7 succession thins markedly and drapes the basement highs. This is especially noticeable over the Barra volcanic ridge system where the sequence is often absent. Likewise thinning and draping are observed against the steep continental margins but here the reflector correlations across the crests of rotated fault blocks and into the intervening half-grabens are mostly ambiguous. Beneath SP 5470 along profile GSI-1, for example, a thick development of layered sediments appears to belong to the pre-R7 sequence.

The axial thinning, draping and seismic appearance of this sequence are best illustrated on profiles GSI-1, S 79-14 and NA-1/NA-1Ext., and also on CM-04 (Fig. 4.5, section 4.1). Towards the base of the sequence a distinct undulating reflector exhibits proximal onlap onto the continental margin and distal onlap onto the axial basement highs (e.g. GSI-1 SPs 4600, 4880, 5000 to 5100, and 5150). In the overlying main pre-R7 unit numerous persistent strong reflectors are flatter and pinch out against the deeper event or, where that is absent, seismic basement. The reflectors of this main layered series are conformable with each other and the bounding R7 event, except below the continental slope and rise where unconformities arise from slumping and sliding.

Roberts (1975) also recognised the locally developed but distinct unconformity near the bottom of the pre-R7 sequence. He suggested that the thin and pale basal unit was the result of pelagic sedimentation over young oceanic basement, while the overlying layered unit represented fan development at the base of a 'marginal plexus' (his words) which subsequently prograded westward. However, the idea of early pelagic sedimentation is not compatible with the hypothesis of Roberts et al. (1981) that accretion of oceanic crust in Rockall Trough took place in abnormally shallow water depths. Furthermore the seismic profiles and pre-R7 isochron chart (Fig. 5.16) indicate progradation of the pre-R7 layered unit towards the south-east; that is Rockall Plateau and Greenland, which presumably were emergent during the early rifting (and spreading?) stage, acted as westerly provenances for the deep sediments in the Trough.

The disposition of R7, and the reflectors above and below, beneath the continental rise and slope is very variable and rarely unequivocal. As a broad generalisation two types of reflector geometry are observed. In the first and simple case the pre-R7 sequence thins and pinches out against the fault blocks of the continental margin and the subsequent reflectors (R6, R5, R2 and others) exhibit progressive overlap (=onlap) up the margin; this is the pattern observed on profiles GSI-1 (NW end), NA-1, NA-1 Ext. and C 80-1 (Figs 5.4 to 5.6). The second case is more complicated and essentially consists of a moderately thick pre-R7 sequence continuing up the slope, with the younger reflectors pinching out abruptly or smoothly against reflector R7. This appears to be the case at the south-eastern ends of GSI-1 and WI-32 (Fig. 5.17). In the latter example the convergence of the R5 and R7 reflectors leads to uncertainty as to whether there is erosional truncation of R7 by R5, or pinch out of R5 against R7: the second option seems more reasonable since there is no evidence for truncation of the reflectors immediately south-east of the convergence. Notice also the chaotic disposition of the pre-R7 reflectors in the deep basin which sits over the boundary (not imaged) between continental basement and layered (volcanic) basement. The confused reflector pattern is probably a result of gravity-induced slumping and sliding together with tectonic disruption.

The S 79-14 seismic profile (Fig. 5.3, 1400 to 1700/096) shows examples of both of the above two reflector patterns. Here R5 appears to pinch out or merge with R7, while the overlying R5-R2 and R2-seabed sequences overlap the margin to terminate in a chaotic zone in a half-graben. Roberts (1975) described similar relationships between the reflectors at the continental margins.

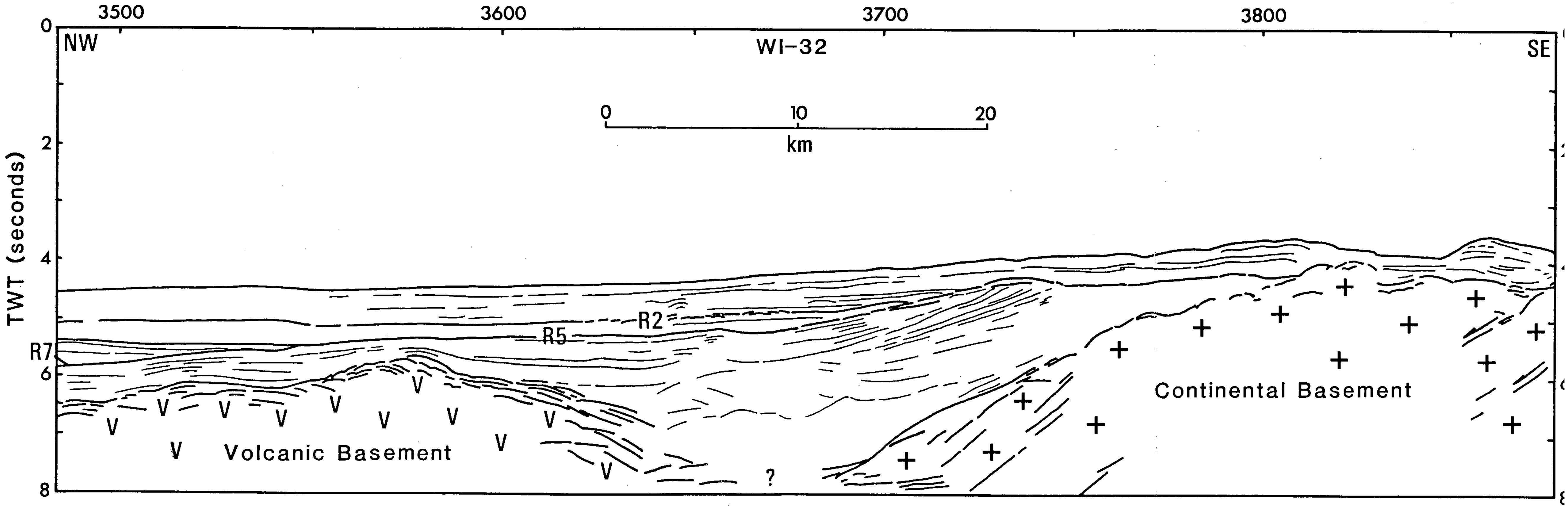


Figure 5.17 Refer to Fig. 5.1 for location

The R5-R7 seismic sequence

As with the pre-R7 sequence the R5-R7 sedimentary interval has been contoured where possible to provide a rough, schematic indication of its distribution and thickness within Rockall Trough (Fig. 5.18). It was not possible to produce similar maps for the R5 to R6 and R6 to R7 intervals because the R6 event is invariably lost in the bubble-pulse reverberations associated with the shallower R5 reflector on the double-channel profiles, which contribute a large proportion of the seismic control. To the south of the research area, around the Charlie-Gibbs Fracture Zone and Clare Lineament, contouring was largely unviable because of the lack of resolution of R7 or, where R7 was absent, of seismic basement. In addition, the correlation of R5 westwards from the Clare Lineament and the CM-04 profile whence it was defined (Chapter 3) is unclear as a result of variations in seismic character.

The isopach chart for the R5-R7 interval emphasises three main features (Fig. 5.18). Firstly, the sequence shows a definite increase in thickness towards the north-eastern area of the chart. Secondly, the sequence thickens, in general, towards the axis of the Trough. This is well marked north of 54°N where the interval expands from less than 0.6 s at the margins to over 1.7 s TWT about the axis of the Trough; the opposite pattern is observed for the underlying sequence - thickening beneath the rise, thinning towards the Trough axis. Thirdly, the pattern of contours in southern Rockall Trough indicates that the BVRS was still a major influence on sedimentation during the time the pre-R5 sequence was deposited.

Apart from the conspicuous and persistent R6 reflector and a number of intrusive sills the R5 to R7 (or basement) sequence has a very characteristic transparent appearance on seismic profiles, as was recognised earlier in Chapter 3. A persistent though weaker reflector is frequently observed below R6 (profiles GSI-1 and NA-1, and CM-04, Fig. 4.5). In places beneath the continental rise and slope this transparent quality gives way to a conformably layered sequence similar to the subjacent pre-R7 succession (e.g. SPs 4500 to 4600, GSI-1, Fig. 5.5). Roberts (1975) also noted the seismic transparency of the R5-R7 sequence and the way it typically thins

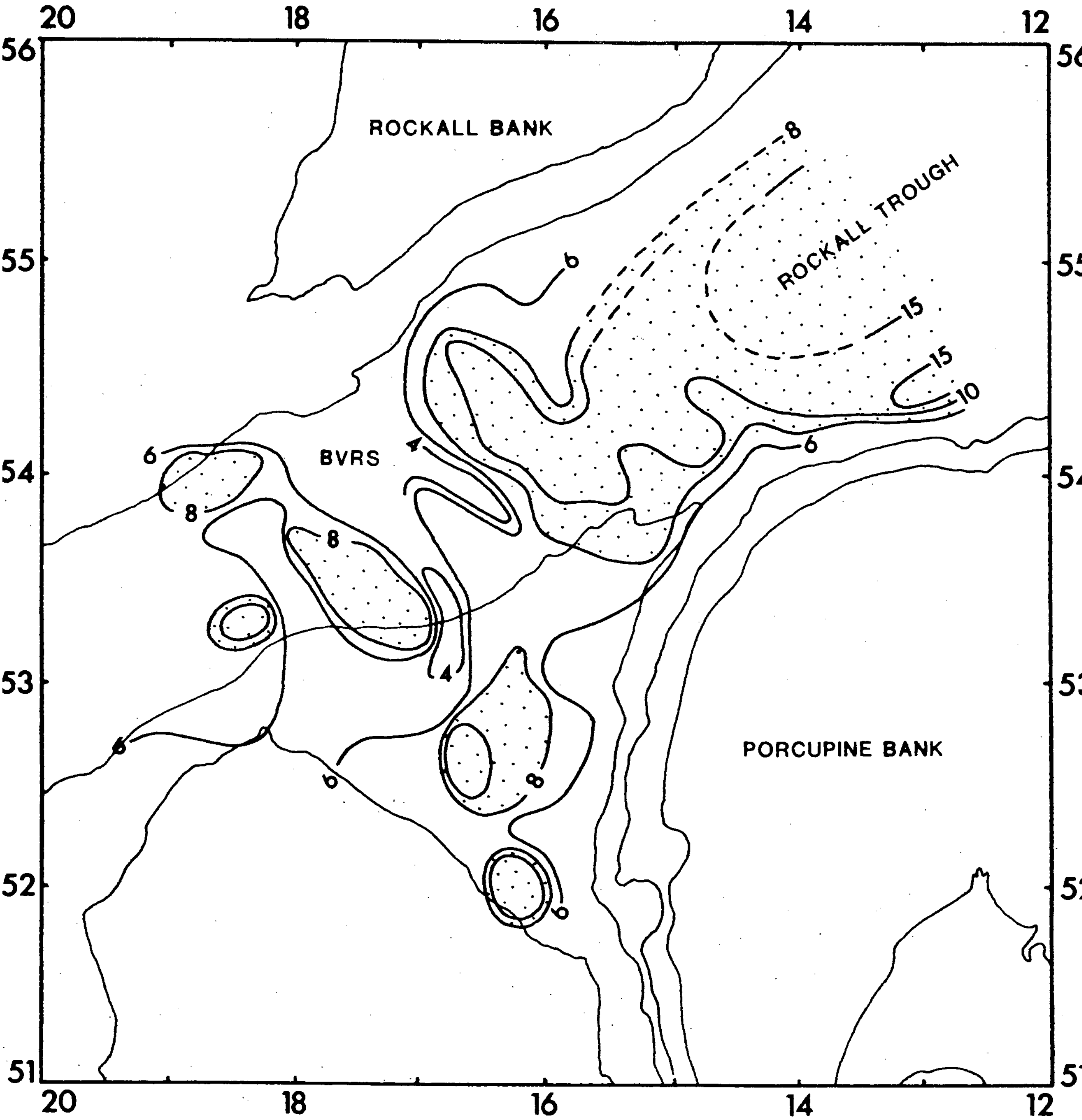


Figure 5.18 Sediment thickness isochron map for R5 - R7 seismic sequence. Contours are labelled in 100 ms of TWT at 200 ms intervals between 0.4 & 1.0 seconds. 1500 ms contour also included. Stippling highlights areas where sequence is thicker than 800 ms TWT. 1000, 2000, 3000 & 4000 m isobaths show general bathymetry.

over and drapes the basement highs within the Trough - an attitude that is well illustrated along profiles C 80-2 and C 80-4 (Figs 5.10 and 5.11).

Roberts did not, however, comment on the possible provenance or lithologies of this sequence other than to remark on the presence of fan development and intense faulting, partly antithetic, adjacent to Lorien Bank. Bearing in mind that R5 is thought to be roughly base Eocene in age, R7 roughly Albian or younger in age (Chapter 3), and that the intervening sediments are so consistently free from layering, it seems reasonable to suppose that they are made up predominantly of upper Cretaceous chalks. At DSDP site 550, approximately 50 km west of Goban Spur, the equivalent interval, which itself has a tendency towards transparency, comprises 9 m of carbonate-free, massive mudstones above 100 m of interbedded bioturbated and finely laminated calcareous mudstones with occasional chalk in the top 10 m (Graciansky, Poag et al. 1985). The drilling information, then, seems to indicate a fairly high calcareous component in the sediments during the Late Cretaceous. The terrigenous component may have been input by turbidity currents flowing down the continental margin, in which case a comparison can be to made the well layered parts of the R5-R7 section at the Trough margins - layering which suggests that interbedding of muddy and marly turbidites with carbonates may be present but passes fairly rapidly, basin-wards, into an unlayered predominantly carbonate section.

The R2 to R5 seismic sequence

The strength and persistence of the R2 and R5 reflectors combined with their regional development make them well suited for defining the upper and lower limits of the next seismic sequence. The late Oligocene R3 reflector of Miller and Tucholke (1983) is not correlated in the Rockall Trough and the upper Eocene-lower Oligocene R4 reflector appears to be only locally developed in proximity to strong bathymetric relief (refer to Chapter 3).

The distribution and thickness of the R2 to R5 seismic sequence show a marked change from the patterns contoured for the previous two intervals (Fig. 5.19). No longer present are the trends governed by the infilling of the main basement depressions in the north-east and south-west (Figs 5.16 and 5.18). Instead a sinuous belt of sediments thicker than 0.5 seconds follows a course similar to that of the present day Feni Ridge. As before the controlling influence of the now-buried BVRS on the sedimentation pattern is seen in the deflection of the contours around 54°N 17°W. A complex region of deposition to the west of this area is caused by the shallow antithetic fault blocks below Lorien Bank (see Fig. 5.7).

Elsewhere in Rockall Trough and beneath the continental margins the R2 to R5 sequence is thin and displays smooth variations in thickness - in contrast to the rapid variations observed in the two underlying sequences. As if to complete the catalogue of changes the area of maximum deposition has shifted away from the north-east of the area and now occupies an irregular tongue of sediments on the western side of the Trough (taken at the 0.75 s thickness contour). If only the sediments thicker than 1.0 s TWT are considered then the depocentre is restricted to the area south of Lorien Bank (Fig. 5.19); this must represent a marked change in oceanic circulation and depositional regime shortly after the formation of reflector R5.

On the seismic reflection profiles the R2-R5 sequence is highly variable, more so than the R5-R7 or R7 to basement intervals. Although the pre-R5 sequence is responsible for infilling and smoothing out much of the basement topography in the Trough there are still places where the R5 reflector exhibits a considerable amount of relief, as above the ridges 1, 2 and 3 of the BVRS (profiles C 80-2 and C 80-4, Figs 5.10 and 5.11). The R2-R5 sequence, in thinning over these highs, masks most of the remaining vestiges of the basement topography. One result of this is that on the flanks of the highs some of the deeper reflectors in the interval pinch out against R5, when elsewhere in the Trough they are normally conformable (e.g. C 80-2).

In general the lower half of the R2 to R5 sequence is fairly clear and unremarkable. The upper half, on the other hand, is strongly reflective and is characterised by a conspicuous crinkled appearance that has been the source of much discussion (Jones et al. 1970; Roberts 1975; Roberts and Kidd 1979). The R2 reflector of this work is taken everywhere as the top of this thick package of reflectors although Masson and Kidd (in press) picked another persistent reflector up to 0.2 s TWT below R2. This second reflector is not resolvable along most of the profiles forming the present research.

The zone of discontinuous crinkled reflectors below R2 may attain a thickness of 0.5 s TWT in places. However, this is exaggerated by the bubble-pulse reverberations and its true depth extent is more like 0.3 seconds or less. Roberts (1975) assigned reflector R4 to the base of this zone of crinkled reflectors, which he believed correlated with the Oligocene cherts beds recovered during DSDP Leg 12 drilling of Hatton-Rockall Basin (Sites 116 and 117, refer to Chapter 3). Drilling over the Feni Ridge at DSDP site 610 has since shown that the R2 reflector of this work corresponds roughly with a zone of velocity and hardness increase arising from compaction and silica diagenesis associated with Middle Miocene siliceous nannofossil chalks (Masson and Kidd, in press). Hence the Eocene-Oligocene R4 horizon must occur deeper in the interval. A variably developed, sometimes strong and persistent, reflector occurs on profiles C 80-1, C 80-2 and C 80-4 usually 0.3-0.4 s below R2. It often marks the boundary between the crinkled upper half of the interval and the clearer lower half (similar relations observed on CM-04 profile, Fig. 4.5). This reflector may correlate with R4 but its lack of marked regional development is at variance with previous stratigraphies.

The crenate aspect of the upper half of the R2 to R5 sequence is caused by the presence of numerous, often overlapping, hyperbolae. These hyperbolae in turn appear to represent diffraction events accompanying the short, discontinuous chert beds at and below the level of R2. This pattern of diffraction hyperbolae should be distinguished from the similar crenate or cuspate appearance of the overlying R2 to seabed interval which seems to result from the rough sea floor topography visible mainly over Feni Ridge. Where this rough topography is absent the pre-R2 section still usually exhibits a typical crenated aspect. Roberts (1975) did not make this distinction between lithology-induced and topography-induced crenations.

From the seismic reflection profiles and the R2 to R5 thickness chart (Fig. 5.19) it is clear that differential deposition and the initial construction of the Feni Ridge sediment drift commenced during the R2 to R5 interval. The absence of persistent reflectors in the lower half of the sequence does not permit the beginning of differential deposition to be defined with any greater accuracy than this. However, the marked unconformity between the flat-lying reflectors of the R2 to R5 sequence and the irregular, undulating R5 and pre-R5 horizons seems to document the change from depositional infilling of the basement topography, to quiet pelagic sedimentation of mainly calcareous and siliceous nannofossil oozes. Thus the available evidence indicates differential deposition operating immediately or soon after the formation of reflector R5, i.e. at the beginning of the Eocene according to Chapter 3, as is well illustrated along profiles S 79-14 and C 80-1 (Figs 5.3 and 5.4). This is slightly earlier than the upper Eocene or lower Oligocene age proposed by Jones et al. (1970), Roberts (1975) and Miller and Tucholke (1983).

The post-R2 seismic sequence

The shallowest seismic sequence, R2 to sea floor, has much the same distribution within Rockall Trough as the underlying R2-R5 interval; that is, thickening beneath Feni Ridge and gradual, smooth thinning elsewhere in the Trough and beneath the continental margins (compare Figs 5.19 and 5.20). The major difference is the shifting of the main depocentre back to the north-east of the region from its previous location south of Lorien Bank. This shift in polarity suggests that the dominant sediment transport path has reverted to along the Trough axis from the north-east. The persistent influence of the BVRS is again indicated by the irregular and constricted isochrons in the vicinity of 54°N 17°W.

Throughout the southern and central Rockall Trough the post-R2 sequence occurs as a highly reflective interval that contrasts strongly with the deeper semi-transparent intervals. In addition there is widespread development of the distinctive crenate seismic appearance noticed by many previous workers. On the crest and flanks of the Feni Ridge drift some proportion of this crenated pattern arises from bubble-pulse reverberations or persistence into the sub-surface of the rough sea bottom caused by the sediment wave fields (e.g. profiles C 80-1 and S 79-14, Figs 5.3 and 5.4). Masson and Kidd (in press) believe that the sediment waves only occur in the top half of the post-R2 sequence (above their R1 horizon). And this now appears to be the case on many of the seismic profiles constituting this research; the crinkled aspect of the lower half of this sequence and the upper half of the R2 to R5 sequence below seems to be a product of irregularly spaced, strong diffraction events rather than the more regular, weaker wavy bedding reflections near the seabed. Support for the diffraction origin of the lower crenated zone comes from regions away from the Feni Ridge where the sediment waves are absent yet the crinkled reflectors are still present directly above and below R2.

A side-scan sonar (GLORIA) survey of the Feni Ridge drift was reported by Roberts and Kidd (1979). They suggested that the sinuousity of the Feni Ridge axis (see Fig. 1.4) parallels changes in the trend and gradient of the adjacent slope of Rockall Bank. But the evidence from this chapter refutes this claim and instead indicates that the south-easterly bulge of Feni Ridge reflects the dominant influence of the pre-existing BVRS. This aside, Roberts and Kidd observed a comparatively smooth western flank to the Ridge but an eastern flank liberally covered with sediment waves, typically 25 to 50 m in height and 1 to 4 km in length. Individual waves could be traced for as much as 26 km. Within the wave fields crest trends were either sub-parallel or markedly transverse to the regional bathymetric contours. The sonar records and high frequency, shallow penetration seismic reflection profiles provided strong evidence for migration of the sediment waves up towards the Ridge crest and downcurrent, that is, to the south-west.

Penetration of the entire post-R2 sequence at DSDP site 610 (report in prep.) revealed a very uniform sequence of nannofossil oozes, marls and charts to total depth (730 m) and a steady above average sedimentation rate, both consistent with drift construction by deposition from long term circulation of bottom waters. The Feni Drift lithologies are essentially pelagic in composition but the sediments also contain evidence of local turbidite activity and minor variations in siliceous, volcanic and mud content (Masson and Kidd, in press; Masson, pers. comm.). These small deviations from the steady state are recorded on the seismic profiles as infrequent stronger, more persistent and flatter reflectors within the post-R2 interval (e.g. SPs 4600 to 4850, GSI-1, Fig. 5.5). They are especially evident below the margins of the Trough where numerous strong reflectors, indicative of turbidity flows, prograde out towards the basin axis and where internal unconformities are frequent (GSI-1 and NA-1 profiles, Figs 5.5 and 5.6). Pinch out of these terrigenous-dominated (turbidite) horizons against R2 is also common.

Away from the Trough margins, in the main body of the post-R2 sequence, the relationship between the reflectors is largely obscured by the pervasive distribution of the crenate events. Internal unconformities are numerous and Roberts (1975) proposed that the sediments in the lens-shaped Feni Ridge are unconformable with both the seabed and underlying siliceous (cherty) horizons. This undoubted chaotic appearance of the post-R2 sequence on the seismic profiles belies the manifestly straightforward stratigraphy and lithology revealed by the DSDP drilling.

5.3 Igneous sills in Rockall Trough

A large number of variously sized, sub-horizontal, strong reflectors are observed within the sedimentary column which are interpreted for the first time as intrusive igneous sills. Similar high-amplitude reflectors have been reported previously in the Rockall Trough (Roberts et al. 1981), in the Faeroe-Shetland Channel (Ridd 1983), and off the mid-Norway continental margin (Price and Rattey 1984): in each of these instances the reflectors bear the same relation to the surrounding sediments as is observed on the seismic profiles of the present study.

In southern and central Rockall Trough Roberts et al. (1981) ascribed these conspicuous abrupt reflectors to abundant lava flows which they dated as earliest Eocene (52 to 55 m.y. B.P.) through correlation with the peak in Early Tertiary igneous activity at that time. The persistent sedimentary reflector Y that they picked at or slightly below the level of the lavas was assigned a similar age accordingly. However, this author believes that the strong flat reflectors in fact represent predominantly, if not wholly, intrusive sills. On the poorer resolution single-channel seismic profiles the reflectors are confused by associated diffraction events, especially around their abrupt lateral terminations, and it is these which give the rough appearance that may render the initial impressions of lava flows (C 84-7 profile, Fig. 5.21; also Fig. 5.9). The strong contrast between the high reflector amplitude of the sill and the weak or transparent nature of the ambient sediments on the C 84-7 profile is typical of many of the other occurrences on single-channel records.

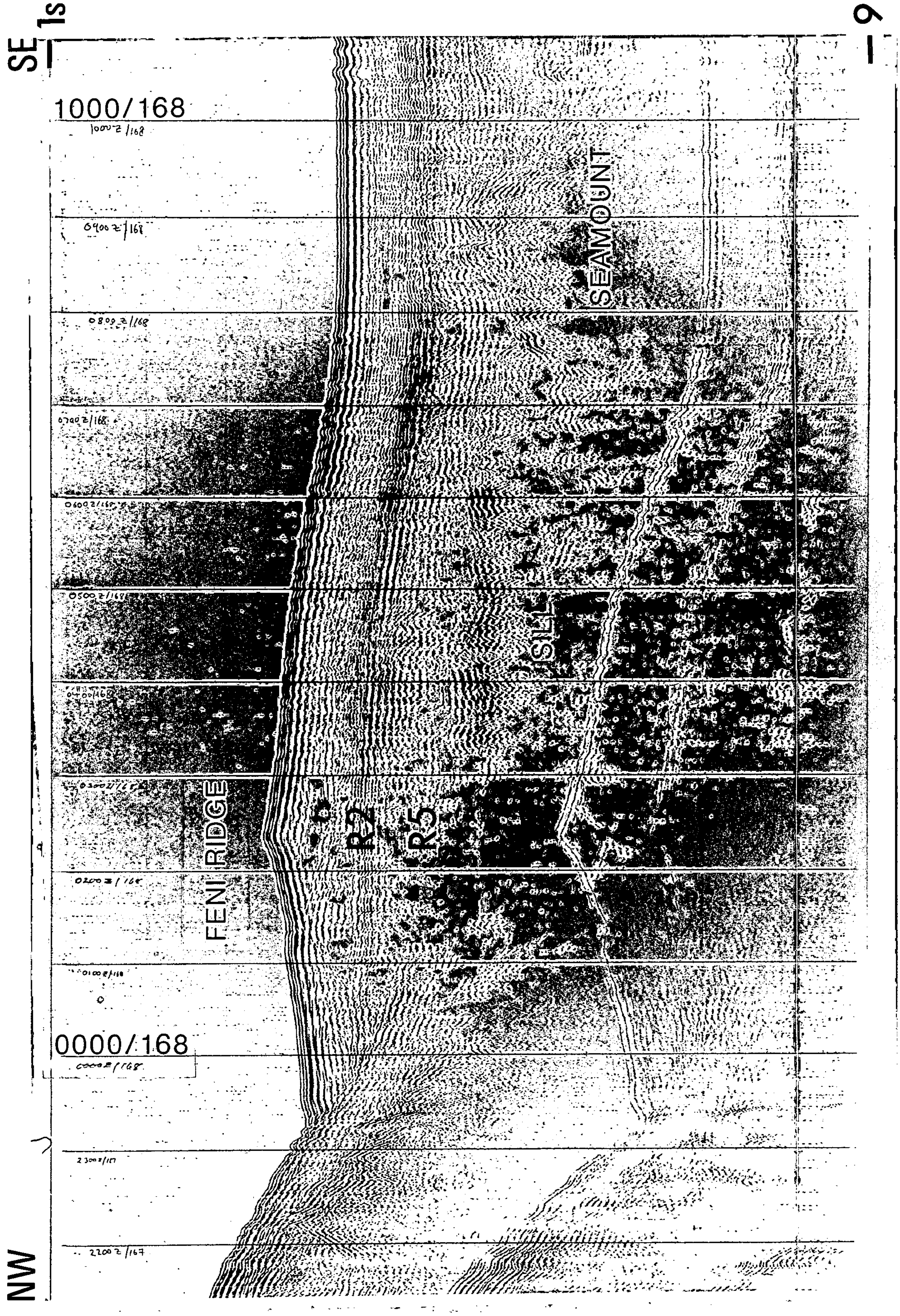


Figure 5.21. Section of C 84-7 seismic profile showing intra-sedimentary sill

On the multichannel seismic profiles the seismic character of the sills and their relationship to the contiguous sedimentary strata is much less ambiguous, largely as a result of the reduced vertical exaggeration; the high exaggeration (about 6 times) on the single-channel profiles aggravates the interpretational difficulties caused by their low resolution. While the multichannel tracks still show the presence of diffraction hyperbolae, it is possible to see that the sills consist of a number of fairly even reflectors varying in length from less than 1 km in some places, in others to as much as 10 km or more without disruption (profile CM-03, Fig. 5.22; profile CM-04, Fig. 4.5). The majority of the sills are concordant with the local sedimentary strata, though transgressions of individual sills to shallower or deeper levels are not uncommon, excellent examples of which are observed on profiles CM-03, SPs 4900-5300 (Fig. 5.22) and CM-04, SPs 3800-4400 (Fig. 4.5). Here is irrefutable proof of the intrusive origin of most of the high-amplitude reflectors in the Trough. It is also noticed that the lateral terminations of the sills regularly end in a short up-turn or down-turn, but in many instances these may be due to diffraction signals or simply artefacts of the seismic record.

The fairly detailed seismic reflection coverage over the southern and, in parts, central Rockall Trough enabled a map of the aerial distribution of the sills to be drawn up for the first time (Chart 6, back pocket). On this map the edges of the sills are marked and interpolated to provide a good indication of their spatial distribution. The representative depths in seconds two-way travel time along most of the controlling seismic sections provide some indication of the down-profile occurrence of the main sill bodies. But no attempt was made either to convert the distribution to true depth or to construct time-depth contours within the sills, owing to the lack of close velocity control and the variable resolution of the sill reflections.

Two main groups or complexes of sills are evident from Chart 6, the first in southern, the second in central Rockall Trough, separated by a 150 km broad zone where no large sills are detected, though small horizontal intrusions are occasionally present but cannot be correlated laterally. Both complexes have the same seismic character and typically cover large, essentially horizontal, areas. The sill reflected on the NA-1/NA-1 Ext. profiles (Fig. 5.6) at between 5.6 and 5.9 s TWT follows an almost unbroken course for over 60 km (55°N, 14°W on Chart 6). Furthermore, in both complexes the component sills often occur at the level of, or a short depth below, the persistent R6 reflector. Few convincing examples exist where the sills intrude above R6, and they seemingly never appear at or above the regional R5 reflector (Fig.s 5.3 to 5.6 and 5.22).

Beneath the level of R6 the sills appear at all horizons from seismic basement upwards. Indeed between SPs 3330 and 3480 on profile WI-32 a strong, persistent and shallow-dipping reflector below the top of seismic basement is interpreted as an intrusive sill by comparison with the adjacent CM-04 profile, which also shows a long irregular sill just above, and dipping into, basement. Elsewhere the sills commonly occur at the level of R7 and a persistent, though not regional, reflector mid-way between R7 and R6. This is noticeably the case north of the Clare Lineament in the southern sill complex (Fig. 4.5), and also below north Porcupine Bank in the central Rockall complex (Fig. 5.22). Interestingly it is rather rare to observe any sills in the poorly layered or transparent intervals between the main regional reflectors, an indication that the latter are important zones of weakness as well as good reflecting horizons.

Apart from the complicated three-tiered style of intrusion immediately north of Porcupine Bank (Fig. 5.22) the north-eastern sill complex appears to show a fairly simple association with the continental margins. In contrast the southern Rockall complex has a more irregular aerial distribution and is completely removed from the continental margins. The part of this complex roughly between latitudes 53°N and 54°N has a strong annular appearance, albeit discontinuous. This impression would probably be enhanced if it could be proved, by further seismic profiling, that the segments of sill to the south-west of this area were also part of the double ring.

It is observed that the conspicuous north-eastern boundary of this annular igneous centre is strongly controlled by the roughly NW-SE trending ridge 2 of the BVRS (Chart 6 and Fig. 5.1). The north to south increase in two-way time to the sill along this boundary closely matches the increase in value of the intersecting basement isochrons of the underlying ridge. The controlling effect is particularly well illustrated near 53°N 16.5°W where the sill, now at a fairly constant 5.7 seconds, curves round the southern lobe of

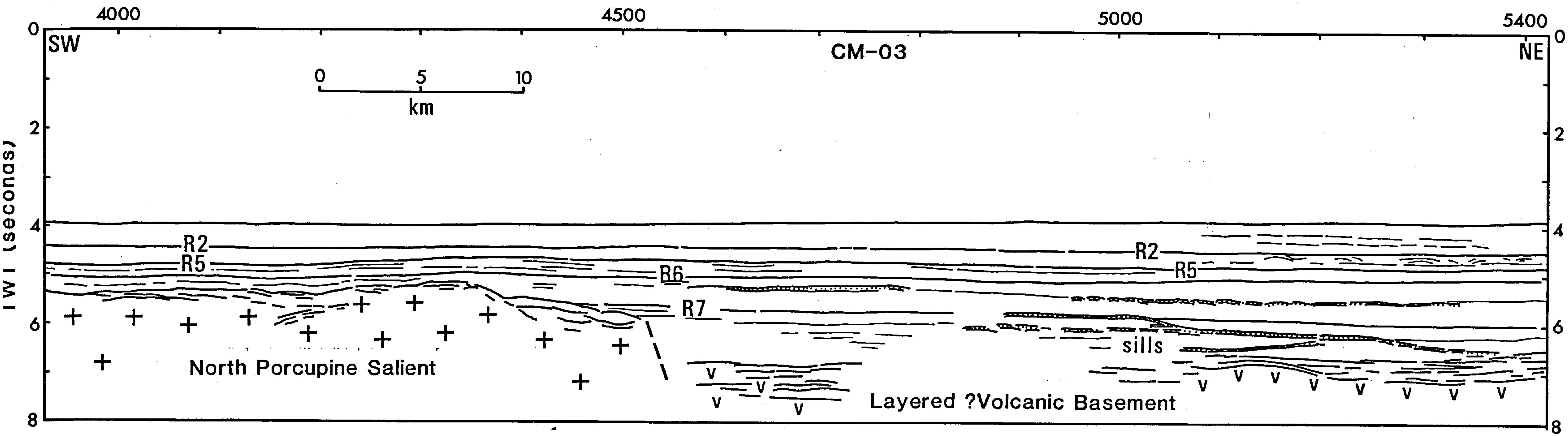


Figure 5.22 Refer to Fig. 5.1 for location

ridge 2. A deeper sill in the same place appears to exhibit the same relationship. This spatial control on the annular centre at least, and possibly also the large sills further to the south, implies that either the sills were intruded into the sediments and had their lateral extension restricted by the impenetrable ridges, or the source and feeders for the sills lay within the volcanic ridges and they then spread out into the surrounding sediment drape. Whichever option is correct, and this author favours the first explanation, it is clear that the igneous episodes which produced the volcanic ridges and the sills were separated by a long period of time - the time it took to deposit the pre-R5 sediments (early Eocene to at least Albian, 40 m.y., according to Chapter 3).

The typical occurrence of the sills at or slightly below reflector R6 but never above R5 has important implications for the age of the sills. In the Gulf of California Einsele (1982) established that sill emplacement was confined mostly to the depth range 50 m to 400 m, with sill thicknesses rarely exceeding 50 m. Despite our lack of knowledge of the physical properties of the sediments and sills in the Trough, if it is assumed that a similar depth range of intrusion was in force here then the sills would be expected to seek their shallowest level 0.05 to 0.3 s TWT below the contemporary sea floor. The sills in the Trough often bear just such a depth relation to the R5 reflector so it can be inferred that they have approximately the same age, that is, earliest Eocene (see Chapter 3). This is the age that Roberts et al. (1981) established for their lavas (my sills), though by a different approach.

On various sections of the WI-32 multichannel profile between shot points 2250 and 2950 the reflections from the sills are more irregular than elsewhere and seem to penetrate very close to the R5 reflector. Whether this is a slightly different interpretation of a highly diffractive sill or whether they actually represent true lava flows is not certain. If they are lavas their apparent close approach to the early Eocene R5 reflector would substantiate the age correlations stated in the previous paragraph. On the other hand it would be difficult to account for the pattern and areal extent of their distribution, especially if they were extruded under water, when one would not expect to see widespread horizontal lava flows. Stepping back, disregarding the finer details, and viewing the whole scene it seems clear that the intrusives and extrusives can comfortably be closely associated together - the latter admittedly being at a slightly higher structural and lithostratigraphical level - as they are in the Gulf of California (Einsele 1982, 1985) without affecting the ages or geological inferences to any great degree.

There is a notable lack of evidence on the seismic profiles to suggest how the magma arrived in the Trough to feed the sills or lavas. In several places sills are seen passing horizontally or at an angle into the seismic basement, in particular the basement ridges (e.g. CM-04 profile, Fig. 4.5). But this neither provides information concerning the direction of intrusion nor does it explain the origin of the remaining sills or flows where such connections are absent. It is more likely, in the present author's opinion, that the sills are fed from beneath by vertical or near-vertical fissure-like feeders. This could explain the long, narrow aspect of many of the sills. In the sill complex off northern Porcupine Bank such feeders appear to have originated beneath the heavily faulted continental margins. In the southern Rockall complex the pattern of sills suggests a series of curvilinear or even ring-shaped vertical feeders at depth, and here analogies can be made with the Tertiary ring complexes of NW Scotland (on a small scale) and, more realistically, again with the Gulf of California where the intra-sedimentary sills and lavas are believed to be fed by vertical dykes (Einsele 1982, 1985).

6. ROCKALL TROUGH: GRAVITY ANOMALY DATA

6.1 Free-air anomaly chart

A free-air anomaly chart of southern and central Rockall Trough is presented in Charts 2 and 7 (both in back pocket). A discussion of the acquisition, reduction and precision of the gravity data was given earlier in Chapter 2. Along-track marine gravity values were available from the three Edinburgh University cruises (Shackleton 3/1979, Challenger 6/1980 and Challenger 1/1984) but only one of the multichannel seismic surveys - the CM survey (tracks 3 and 4; Chart 1). Of a total of 8600 line kilometres of geophysical data approximately 5500 km hold gravity anomaly data, which were used to compile Chart 2. This chart represents the most comprehensive and accurate free-air anomaly map of southern and central Rockall Trough presently available. Certainly in the southern part of the map the control from the gravity tracks is good and cross-over errors are within sensible limits (refer to Chapter 2). To the north, however, the available gravity lines are fewer and more widely separated; consequently the anomaly contours are less well constrained, except in the region around 54°N 14°W, off northern Porcupine Bank, where tight control is provided by the Challenger 1/84 box survey (tracks 8 to 17) and a number of other profiles (Fig. 6.1 and Chart 7). The free-air anomaly field over the Charlie-Gibbs Fracture Zone, Clare Lineament and Southern Porcupine Bank was described and modelled in Chapter 4.

Within the Rockall Trough the free-air anomaly field is dominated by the Barra anomaly and the anomalies caused by two major bathymetric features - the steep continental margins of the Trough and the Feni Ridge sediment drift (Charts 2 and 7). The free-air edge effect over the two margins, as explained in Chapter 4, arises from the rapid increase in water depth across the continental slopes, in turn reflecting attenuation of continental crust at depth (see section 6.2). The free-air edge effect is well marked over the Porcupine Bank where it consists of a steep gradient rising to 60 mgal over the continental slope and a curvilinear anomaly low above the continental rise. The edge effect across the Rockall Bank margin, however, is poorly defined owing to the sparse geophysical coverage and the disrupting effects of the Barra free-air anomaly (Chart 2).

Whereas the continental slopes of Rockall and Porcupine Bank exhibit fairly constant gradients and only smooth variations in trend on the bathymetric maps (Figs 1.1 and 1.4), excepting the re-entrants at Lorien Channel (Fig. 5.2) and east of the Clare Lineament, the free-air edge effect is somewhat more variable. Below west Porcupine Bank, south of 53°N, the edge effect low is regular and narrow. North of 53°N the low broadens and is offset markedly to the west. This offset appears to line up both with a deflection in the anomaly contours beneath the adjacent margin and with the steep, well developed north-east edge of the ovoid Barra anomaly in Rockall Trough (Charts 2 and 7). The disturbance in the FAA contours is coincident with a similarly shaped deflection in the acoustic basement isochrons (Fig. 5.1 and Chart 4, back pocket), presumed to represent a transfer zone in the subjacent continental crust. In addition the offset in the edge effect anomaly overlies the southern limit of ridge 3 of the BVRS (Chart 7 and Fig. 5.1). Such spatial coincidences may be fortuitous but equally they may be indicators of some form of transverse (rifting or spreading?) structure within southern Rockall Trough.

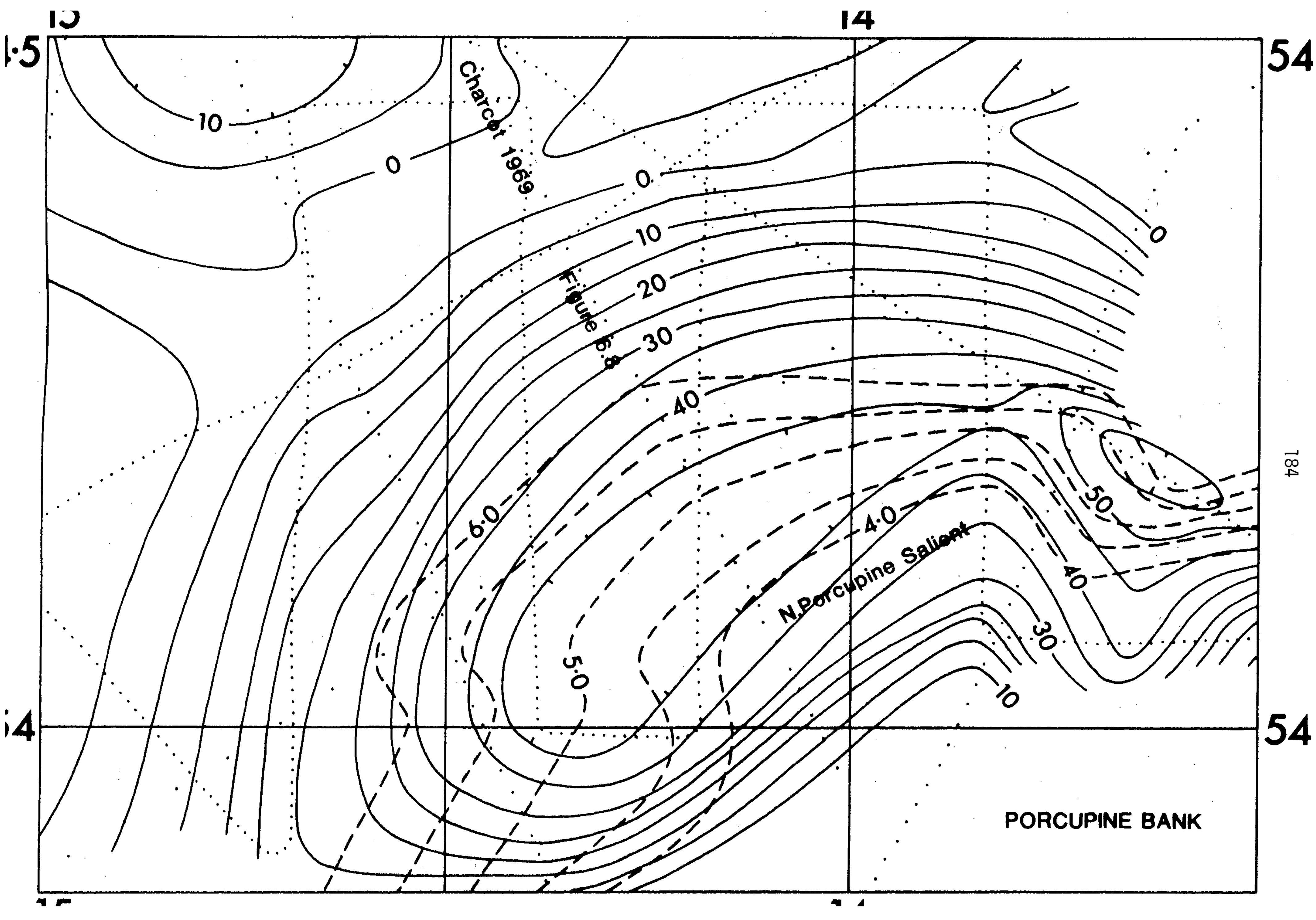
The north-western counterpart of this edge effect, below Rockall Bank, is confused by interference from the Barra and Feni Ridge anomalies. Where it is seen, however, the slope edge effect is less pronounced and ranges from +40 to -20 mgal or less, compared with the steep gradient and range of +60 to 0 mgal for south and west Porcupine Bank (Chart 2). At the north-eastern end of the free-air anomaly map the slope edge effect broadens suddenly and increases its peak-to-trough amplitude dramatically, achieving a nadir of less than -55 mgal below the North Porcupine Salient (Chart 7; Fig. 6.1) at 54.2°N 13.6°W. Does this abrupt change in the gravity field herald a concomitant change in deep crustal structure, or does it merely reflect the diminishing influences of the Barra and Feni Ridge anomalies further to the south, or perhaps a combination of the two? Two dimensional forward modelling (section 6.2) favours the last option.

Megson (1983) also observed this sudden change in the free-air anomaly (FAA) field across a NW-SE trending line joining southern Rockall Bank and northern Porcupine Bank. Her FAA chart showed prominent curvilinear lows at the foot of both margins persisting north-east along most of Rockall Trough. The pronounced semi-circular negative FAA around 54°N 14°W (Chart 7, inset box) was recontoured at a scale of 1:250,000 and at a contour interval of 5 mgal (Fig. 6.1), largely made possible by the close grid survey of the Challenger 1/1984 cruise (see Chapter 2). Superimposing the 4.0 to 6.0 s TWT acoustic basement isochrons of Chart 4 on the detailed FAA map seems to indicate a close correspondence between the centre of the anomaly low and the buried projecting platform area, called here the North Porcupine Salient (Fig. 6.1). The close coincidence in trends of the acoustic basement and FAA field near 54.2°N 13.6°W is especially noteworthy and suggestive of another, north-west oriented, transfer structure within the continental margin.

Just as the rapid variation in water depths across the Trough margins results in the slope edge effect, so the prominent Feni Ridge bathymetric ridge apparently also manifests itself in the FAA chart. Evidence of such is provided by the increase in value of the FAA contours in the north-western half of the Trough and their similar outline to the Feni Ridge isobaths. This is well displayed around 53°N 19°W where the 40 mgal closure follows the local crest of the Ridge and the adjacent contours mirror the shape of the isobaths (Chart 2; see Fig. 6.4). The corrected FAA profile along track C 80-9/10 (Fig. 4.9a) illustrates that the 40 mgal anomaly here largely disappears when the effect of varying water depth is removed. The indentation in the FAA contours at 53°N 18°W appears to correlate with a weakly developed channel in the underlying seabed. But the paucity of seismic reflection profiles here does not allow the writer to identify any deeper source for this irregularity.

The Barra anomaly is a prominent free-air anomaly located on the western side of the Trough and lies roughly over the noticeable south-eastern bulge of the Feni Ridge (53.5°N 17.5°W; Chart 2). The presence of a conspicuous FAA here was first noted by Megson (1983) who referred to it as the "70 mgal anomaly" because of the two small peaks of that value near its centre. Computer processing of the gravity anomaly data from the Challenger 6/80 and 1/84 cruises (see Chapter 2), which largely define the Barra anomaly, show that a value

Figure 6.1 (opposite). Detailed free-air anomaly map over part of north Porcupine Bank and central Rockall Trough. See Chart 7 (back pocket) for location of diagram. Main contours are in mgal at 5 mgal intervals. Labelled contours are negative mgal, reaching a low of -55 mgal at the far right (east) of the area. Dotted lines indicate gravity profile control used in contouring. Bold dashed lines are 4.0 to 6.0 s TWT isochrons to seismic basement forming the North Poecupine Salient, isochrons at 0.5 s TWT intervals. Note close correlation between free-air anomaly and basement contours in the region of 54.2 N, 13.7 W.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N.Porcupine Salient
PORCUPINE BANK

of 70 mgal was recorded at only one control point. However, an irregular area enclosing FAA values between 65 and 69 mgal occurs to the north-west of the localised oval anomaly low within the 60 mgal contour.

The roughly oval outline of the Barra anomaly as a whole is hinted at on the free-air anomaly map of Megson (1983) and confirmed and better defined on Charts 2 and 7 of the present work. Strictly it is more triangular than oval in shape, strong linear gradients forming the western, southern and north-eastern boundaries. Obviously the gravity effect of the Feni Ridge contributes partly to the amplitude and gradients of the Barra anomaly. This is particularly true for the southern gradient which overlies the eastern slope of Feni Ridge; but with the remaining two boundaries, being roughly perpendicular to the underlying bathymetric trends, it is harder to discern the influence of the sea floor relief on the Barra anomaly, though there must be some. The implications for the two-dimensional gravity modelling are considered in the next section.

Thus, even accounting for the variations in water depth over the Feni Ridge, the Barra anomaly remains an important feature of the FAA field of southern Rockall Trough. Comparison of Charts 2, 4 and 6 (all in back pocket) reveals a close spatial relationship between the centre and NW-SE axis of the Barra anomaly, the position of the deep narrow trough between ridges 1 and 2 of the BVRS (Fig. 5.1), and also that of the annular intra-sedimentary sill complex (Chart 7). In the same context two observations are of particular note. Firstly a strong NW-SE trend is developed in the Barra anomaly, the sill complex and the buried ridges of the BVRS. Secondly there is a distinct lack of correlation between the individual Barra volcanic ridges and the free-air anomaly field, despite the local similarities in orientation just mentioned. The long axis of the Barra anomaly appears to be centred over the narrow trough within the BVRS, while the pronounced relief on ridge 3 to the north-east and ridges 1 and 2 to the south (see Chapter 5 for seismic profiles) is notable for the absence of a corresponding gravity anomaly (Chart 7). A weak to moderately developed FAA low sited over the steep north-eastern slope of ridge 3 is the only area where the BVRS seems to be manifested in the gravity field. The conspicuous NNE oriented basement ridges a short way north of the Charlie-Gibbs Fracture Zone, as illustrated excellently on profile S 79-14 (Fig. 5.3), show absolutely no correlation to the approximately east-west trending FAA contours here (Chart 7).

This lack of correspondence suggests that either the basement ridges have a low density contrast with the surrounding sediments or the ridges are considerably thinner than is presently thought, or some combination of the two. The seismic reflection and refraction data do not favour the thesis that the Barra volcanic ridges rest on top of older sediments, as discussed in the previous chapter. If, however, the ridges are assumed to have a density contrast of about 0.25 g/cm^3 with the sediments (2.5 g/cm^3 against 2.25 g/cm^3) then a 20 to 30 mgal anomaly would be expected over the edge of a 2 to 3 km thick basalt pile - thicknesses that are indicated from the reflection and refraction work. Such anomalies are not recorded here so it remains to conclude that perhaps the bulk density of the volcanic ridges is reduced by extensive alteration and interbedding of sediments.

Alternatively, or perhaps additionally, there may be implications for the composition of the supposed lava flows constituting the ridges. If the lavas are predominantly silicic, rather than basic, then the gravity effect would be reduced to some extent. Unfortunately it is difficult to envisage the extrusion of viscous silicic volcanics over such a large (submerged ?) area. Comparisons with present day continental rift terranes may provide some guidelines in this regard. These complications aside, it appears that the Barra free-air anomaly has its source seated deeper in the crust or near the crust-mantle boundary. The following discussion of forward gravity modelling sheds more light on these problems.

6.2 Gravity modelling in two and three dimensions

Forward gravity modelling in two dimensions was performed along a number of informative and representative free-air anomaly profiles within Rockall Trough. These profiles (Figs 6.2, 6.3, 6.6 and 6.8) are located on Chart 7 (back pocket). Rapid computation of gravity anomaly profiles, given a starting crustal model, was made possible using a FORTRAN computer program made available by R.A. Scrutton and mounted on the Edinburgh University ICL 2972 main-frame computer. This program was modified extensively by the present author to permit greater interactive use and to provide a visual appreciation of the model and results by making use of automatic plotting and drawing facilities. Graphical output from the program forms the basis of most of the figures in this chapter.

Since the greatest density contrast present, and therefore that which has the potential for contributing the largest anomaly, is the one between sea water and sediments or indurated basement, it is important that the depth to the seabed along the modelled profiles is defined as precisely as is reasonable given the storage capacity of the program. This is particularly pertinent beneath the steep continental slopes bounding the Trough where much of the free-air edge effect results from the rapid deepening of the sea floor. Consequently the shape of the sea bottom was determined from the along-track corrected bathymetry readings taken during the cruises rather than from the published bathymetric contour charts. The observed free-air anomaly profiles, with which the calculated profiles are compared in the figures, are defined by data points (between 1 and 3 km apart) typed in to an accuracy of 1 mgal in general, or 0.1 mgal in the case of the Challenger 6/80 profiles.

Challenger 80-1 and 84-4 gravity model

The C 80-1 and C 84-4 tracks (Chart 1) together provide a 430 km long free-air anomaly profile across the whole width of Rockall Trough from west Porcupine Bank north-west to southern Rockall Bank (Fig. 6.2 and Chart 7). Despite the slightly different trends of the two sections of the profile and the rather irregular nature of the surrounding gravity field, especially to the north-west (Charts 2 and 7), the length of the profile makes it well suited to modelling in two dimensions. As well as almost traversing the slope edge effects at right angles the model profile passes through the centre of the Barra anomaly, partly along its major axis. The amplitude and gradients of this prominent anomaly are comparable to those of the adjacent slope anomalies (Fig. 6.2).

Below the sea floor, which as stated above is well defined, control on the depth and slope of the crustal and Moho interfaces, and the density contrasts between them, is extremely poor. To the south-east of the profile, beneath Porcupine Bank, the seismic reflection records indicate a very thin sedimentary cover. The small, steep-sided graben at around 370 km is visible on the C 80-1 and WI-10 seismic profiles (see previous chapter). A width of about 20 km and a depth of c. 1 km for this graben is compatible with both the seismic and gravity data - seeming to account nicely for the 15 to 20 mgal anomaly low in the gravity profile (Fig. 6.2).

The seismic refraction work of Whitmarsh et al. (1974) and the gravity anomaly modelling of Buckley and Bailey (1975) indicate crustal thicknesses of 28 to 30 km below Porcupine Bank, typical of continental material. Scrutton (1972) also used a thickness of 30 km for Porcupine Bank in his gravity and seismic refraction crustal models across Rockall Trough. The densities of 2.81 and 3.07 g/cm^3 which he adopted for the upper and lower crust, respectively, also provide a good fit to the gravity observations along the C 80-1/84-4 model (Figs 6.2a and b). Scrutton (1972) placed the mid-crustal discontinuity separating these two densities at around 20 km depth, though in my models a maximum depth of about 16 km gives an optimum fit between the observed and computed gravity profiles and also compares well with the same interface at 16 km below Rockall Plateau (Fig. 6.2).

In an alternative interpretation of their seismic refraction data Whitmarsh et al. (1974) proposed a refracting horizon at 11 km depth below Porcupine Bank, separating crustal velocities of 6.0 km/s above and 6.7 km/s below. These velocities correspond roughly with densities of 2.67 and 2.75 g/cm^3, values rather lower than used here. Likewise the 11 km depth of the interface is somewhat shallower than that modelled for the gravity data (Figs 6.2a and b). It is unlikely that the increase in density of the crust with depth takes place over such a sharp boundary. More realistically a strong density gradient is present in a mid-crustal transitional zone, perhaps thicker than 5 km if the seismic refraction data is to be believed. Disregarding the inevitable complications to the density structure of the crust, particularly at shallow levels, a good approximation to the observed gravity profile is achieved by modelling a 2.75 g/cm^3 upper crust and a 3.00 g/cm^3 lower crust.

Beneath Rockall Bank, at the north-west end of the model, a similar simple crustal structure produces an excellent fit between the observed and calculated anomaly profiles (Figs 6.2a and b). The refraction station A of Scrutton (1972) along Rockall Bank provides evidence for a 7.02 km/s refractor at 16 km depth separating upper and lower crustal densities of 2.81 and 3.07 g/cm^3, as modelled in Figures 6.2a and b. Bunch (1979) reinterpreted the same refraction station and repositioned the 7.1 km/s refractor at a depth of 22 km and also proposed a mantle transition between 30 and 32 km. The C 80-1/C 84-4 gravity model favours the original 16 km depth to the mid-crustal density contrast, an observation that may reflect the location of the model over Lorien Channel, a considerable distance south-west of Scrutton's (1972) refraction station.

The excellent match of the observed and calculated gradients of the free-air anomaly above the Rockall Bank continental margin was achieved with little adjustment to the initial crustal structure (Figs 6.2a and b). Rapid attenuation of the continental crust from about 28 km to 7 km occurs over a horizontal distance of 40-50 km, much of the thinning apparently being taken up in the lower crust thus permitting the marked shallowing of the Mohorivicic discontinuity beneath Rockall Trough. It is the position and slope of the Moho and top crustal surfaces beneath the margin, particularly the latter, which govern the form of the free-air edge effect. Despite the fact that it is not possible to accurately define the block-faulted and rotated geometry of the shallow continental margin basement from the seismic reflection profiles, the linear segment with a gradient of 12.5° used as a substitute in the models appears to be a very good approximation. The deeper Moho slope has a gradient of 21-22° and is presumed to be much more uniform than the faulted, brittle upper crustal slope.

The geometry of the Porcupine Bank continental margin is very similar to that below Rockall Bank, crustal thinning from 30 km to c. 7 km again taking place over a distance of about 50 km (Figs 6.2a and b). The comparatively poor fit of the accompanying steep gravity gradients, a mis-match that is also evident in the models of Megson (1983), is almost certainly due to the lack of control, from the seismic profiles, on the shape of the buried continental slope. The modelled slope of 8.5° is seemingly too shallow; moving the boundary between continental crust and thin crust beneath Rockall Trough some 10 km to the south-east would raise the slope angle to about 16° and improve the fit of the gravity gradients, as well as simplifying the crustal geometry at the extreme thin edge of the margin. The slope of the Moho here is again of the order of 22°, though it is likely that the abrupt intersection with the flat Moho at the base of the continental crust landwards of the margin is in reality replaced by a smooth curved surface.

The steep FAA gradient at the far south-eastern end of the models (Fig. 6.2) results from the sedimentary basin beneath Porcupine Seabight which, while not observed on the seismic profile C 80-1, does influence the gravity field. Allowance for this influence was made by incorporating the crustal structure of the Seabight into the model (though not depicted in Figs 6.2a or b) according to the results of Whitmarsh et al. (1974) and Buckley and Bailey (1975).

Within Rockall Trough proper the two gravity models predict a thin crustal layer, 7 km thick at most, separating the largely undeformed sedimentary basin from the high density (3.33 g/cm^3) elevated upper mantle. This contrast of thin crust beneath Rockall Trough against normal thick continental crust beneath the adjacent banks is a striking feature of all the gravity models in this region. The depths to the top of the crust and Moho below the Trough as predicted by the gravity models conform closely to the depths indicated by the seismic refraction data, for example those of Hill (1952) near the junction of profiles C 80-1 and C 84-4 (Fig. 6.2; see

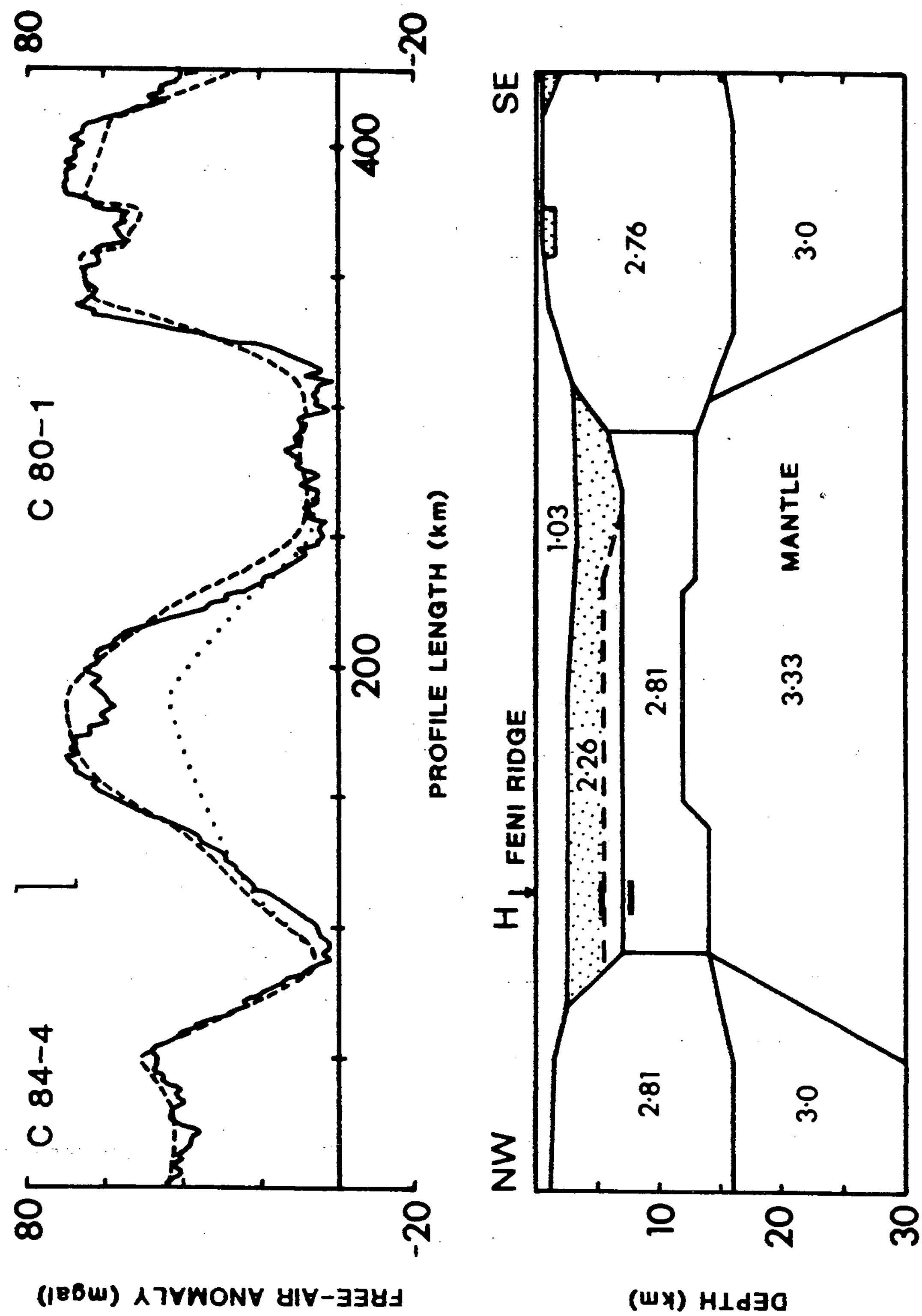


Figure 6.2a. Two-dimensional gravity model along profiles C 80-1 and C 84-4. See text for discussion and Chart 7 for location. Upper figure: solid line = observed anomaly; dashed line = computed anomaly; dotted line = computed anomaly without Moho disturbance. Lower figure: stipple = sediments; H = Hill's (1952) refraction stations; numbers in prisms are densities in g/cm^3.

also Figs 5.12 and 5.14). Hill (1952) commented that an 8.1 km/s layer, i.e. upper mantle, would have been detected by his refraction survey if its top surface lay at 12 km depth or shallower. The 14 km depth for the Moho assumed in Figure 6.2 does not contradict that observation.

The Barra free-air anomaly, reaching nearly 70 mgal in amplitude, dominates the gravity profile over Rockall Trough. The small low within the flattish-topped peak of the Barra anomaly seen along this profile appears to be limited in extent (Charts 2 and 7) and no attempt was made to model it. In Figure 6.2a the effect of the prominent Feni Ridge on the gravity profile is illustrated by removing the geologic/density anomalies in the subjacent crust, from which it is clear that up to 35 mgal of anomaly require accounting for. Two model solutions for the Barra anomaly are presented in Figures 6.2a and b which the present author believes are the most reasonable bearing in mind our current geological knowledge of the region. In both models the poor quality of the associated seismic reflection profiles does not permit the shape and thickness of the sediments and the basement to be quantified with any great accuracy. For instance, along the C 80-1 profile, which covers that section of the model between the 115 and 430 km marks, the relationship between the sediments, the Barra volcanic ridges and the deep basement are hazy at best (see Fig. 5.4): in the central parts of the Trough this ambiguity is worsened by the numerous intrusive sills present within the sediments.

The steep slope in seismic basement at around 1100/096, profile C 80-1 (Fig. 5.4) marks the eastern edge of ridge 2 of the Barra volcanic ridge system (Fig. 5.1). This feature is marked on the gravity models of Figure 6.2 below the 250 km mark and the persistence of the BVRS right across the Trough and onto Rockall bank is shown by the uniform layer, 1.5 km thick, resting on top of thinned crust. However, as remarked in section 6.1 of this chapter, there is a peculiar and unexpected lack of correlation between the ridges of the BVRS and the FAA field. The profile presently under discussion is no exception, for the marked basement gradient at 250 km clearly does not manifest itself in the gravity profile above (Figs 6.2a and b). Thus, in common with other gravity models traversing the BVRS, the volcanic layer is given a much simplified geometry and has no density contrast with the surrounding deeper sedimentary strata.

In the first of the two preferred models (Fig. 6.2a) the Barra anomaly is presumed to arise from a broad (90 km) elevation in the crust-mantle boundary across which there is a density contrast of 0.52 g/cm^3 (2.81 versus 3.33 g/cm^3). The top of the perturbance occurs at 12 km depth and descends to 14 km to the north-west and 13 km to the south-east, resulting in flanking slopes with an angle of 11° to the horizontal -fairly shallow despite their steep appearance on the squashed gravity models (V.E.=c.5). Drawn out at no vertical exaggeration the disturbance in the Moho looks quite plausible and the fit of the observed and computed gradients of the Barra anomaly could be improved perhaps by steepening the flanking slopes of the Moho anomaly. For instance, doubling the steepness of the edges would still only make them the same angle as modelled for the M-Discontinuity beneath the continental margins, 22°. The broad Moho disturbance invoked here produces a flattish looking peak that is in keeping with the appearance of the Barra anomaly on the FAA contour chart (Charts 2 and 7, back pocket).

In the second gravity model (Fig. 6.2b) there is no marked relief on the Moho beneath the Trough, instead the Barra anomaly results from the presence of large block of high density material within the thinned crust. The vertically-sided block, which has a 0.15 g/cm^3 lateral density contrast, is 70 km across and 6-7 km thick, effectively separating the normal crust (2.81 g/cm^3) to either side. There is a noticeable improvement in the fit of the observed and computed gradients of the Barra anomaly, compared with the previous model, but this gain is partly countered by the more sharply peaked aspect of the calculated anomaly which is at variance with both the observed profiles and the FAA chart.

Interpreting the high density mass from a geological point of view leads to two likely possibilities: either it represents an intrusion, probably of gabbroic composition, into the thinned Rockall Trough crust, or it is the focus of some kind of limited oceanic accretionary phase. Although the roughly oval shape of the Barra anomaly in plan might initially favour its interpretation as a single, discrete intrusion there are a number of factors that count against such a proposal. If the intrusion was composed mostly of gabbroic or even ultramafic rocks, as suggested by the density of 2.96 g/cm^3, then one would expect to observe a corresponding magnetic anomaly over the body. It seems unlikely that the body (intrusion)

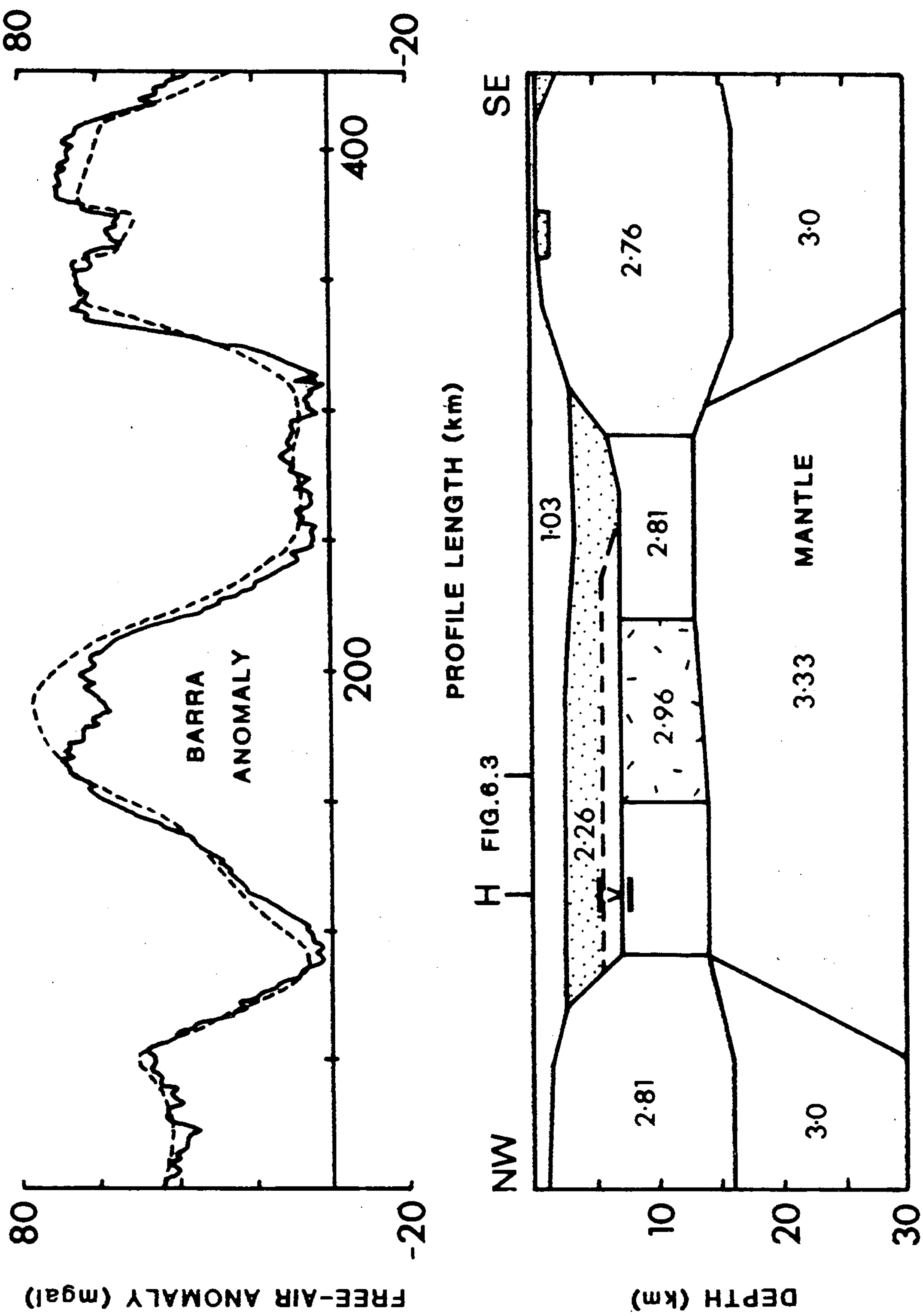


Figure 6.2b. Two-dimensional gravity model along profiles C 80-1 and C 84-4. Ornament, etc. as for Fig. 6.2a. V = volcanic basement; dashes = intrusive body. Intersection of Fig. 6.3 shown in lower figure.

lies below the Curie point isotherm, in which event no magnetic signature would be recorded, since its base only reaches a depth of 14 km.

In addition, it is difficult to envisage a situation whereby a very large intrusion is emplaced into a deep, oceanic environment as seems to be the case here. In the British Isles the Hercynian granites of South-West England and the Caledonian granites of north-east Scotland have exposed diameters measuring up to 35-50 km, and their buried continuations are certain to be much larger than these dimensions; but these bodies were emplaced into thickened continental crust not the much thinner equivocal crust beneath Rockall Trough. Also there would be vertical and horizontal space problems associated with a single, large intrusion.

The second interpretation involving some form of nascent oceanic accretionary process is considerably more attractive. This may take the form of pervasive dyking and minor intrusion of the otherwise rather uniform thin crust in the Trough, the combined effect being to locally raise the density of the crust yet fail to result in any coherent magnetic anomaly pattern. This hypothesis is fine if the host rock comprises continental crust, as is suggested by the densities of 2.81 g/cm^3 in Figures 6.2a and 6.2b. If, however, there is well developed oceanic crust beneath Rockall Trough, which many present workers still believe and which will be debated later in this section and at length in the final chapter, then the 0.15 g/cm^3 density contrast invoked in Figure 6.2b will vanish or be greatly reduced and the gravity anomaly will remain unaccounted for. The same arguments apply to the earlier single intrusion model. For this reason mainly the present author considers the first model of the three, where a broad disturbance - a shallowing - in the M-discontinuity is the source of the anomaly, to be the most realistic from both the geophysical and geological standpoints. It is recognised, though, that some combination of mantle up-warping plus crustal intrusion and dilation may have been in operation simultaneously, indeed may have been coeval (or cause and effect). As is often the case with such problems of ambiguity (or lack of uniqueness) a compromise is the best solution!

The two gravity anomaly models (Figs 6.2a and b) both show an increased depth to the Moho beneath the north-western side of the Trough (14 km) as compared with the south-eastern side (13 km). It is possible that the depth to the Moho to the north-west is in fact 13.5 km with the other 500 m difference being accommodated by an increase in depth to the top of the overlying crust from 7 km to 7.5 km. This would both preserve the close fit of the observed and calculated profiles in this area and also maintain a constant thickness of 7 km for the thinned crust across the entire Trough (excepting, of course, the Moho perturbance), a geometry that is somewhat tidier than those depicted. The 500 m+ downwarping of the crust on this north-western side of the basin may then be related to the subsidence of the Rockall Plateau to which the thin crust is presumably welded. The subsidence of the Plateau, as indicated by its greater water depth relative to Porcupine Bank and the Irish continental shelf, must in turn be strongly influenced still by the long period thermal contraction and subsidence of the oceanic crust in the Iceland Basin (Talwani et al. 1971; Vogt and Avery 1974; Bott et al. 1983).

Challenger 80-4 gravity model

In order to further investigate the source of the Barra anomaly a second crustal model was constructed along the C 80-4 seismic and free-air anomaly profile (Fig. 6.3). The 190 km long C 80-4 profile was chosen for forward two-dimensional gravity modelling in preference to the longer, parallel C 80-2/C 84-3 line because the former traverses through the centre of the Barra anomaly roughly perpendicular to its long axis and also is further removed from the erroneous gravity effects of Rockall Bank (Chart 7).

The previous C 80-1/C 84-4 gravity model has already shown that approximately half of the Barra anomaly is a result of the marked bathymetric relief associated with Feni Ridge (Fig. 6.2a). Since the C 80-4 profile lies roughly along the crest of the Ridge there is negligible relief on the sea floor in this direction (Figs 5.11 and 6.3). Consequently the free-air effect of the bathymetric variations across the Rockall Bank continental slope and Feni Ridge was calculated using a three-dimensional gravity anomaly program written by M. Talwani (made available by R.A. Scrutton and R. Hipkin, University

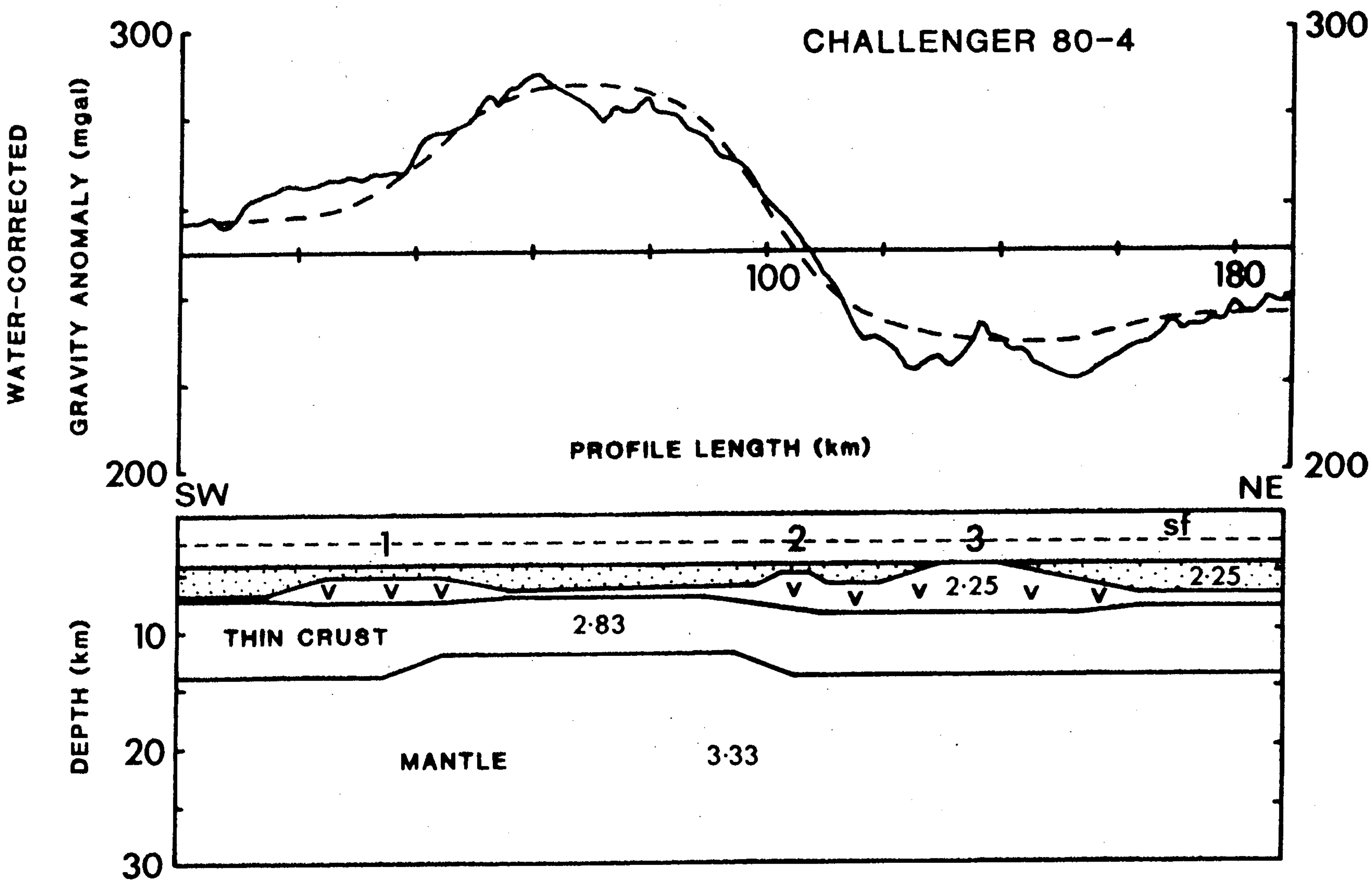


Figure 6.3 Two-dimensional gravity model along profile C 80-4. See text for discussion and Chart 7 for location. Upper figure: solid line = observed anomaly corrected in 3D for regional wter variations according to Fig. 6.4; dashed line = computed anomaly. Lower figure: sf = sea floor; stipple = sediments not removed in 3D correction; V = volcanic basement of Barra volcanic ridges (1,2 & 3); numbers are densities in g/cm^3.

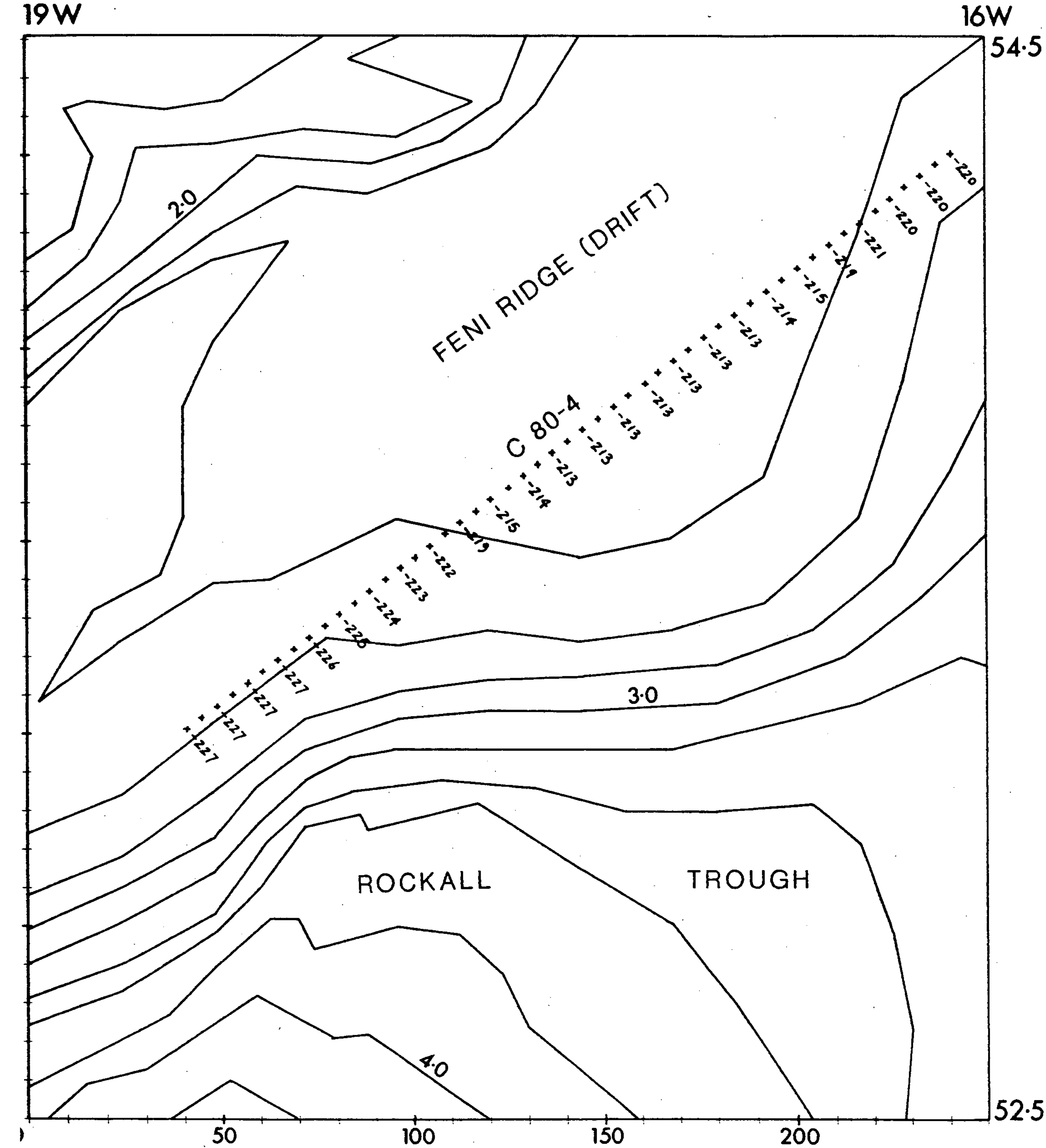


Figure 6.4 Bathymetric contours used to correct in 3D the free-air anomaly profile along C 80-4 for regional water depth variations. Isobaths every 200m between 1600 and 4200 m. 2, 3 & 4 km isobaths labelled. Corrections calculated every 5 km (crosses), annotated every 10 km. Scales on ordinate and abscissa are in km.

of Edinburgh) and based on the gravity attraction of horizontal slabs as developed by Talwani and Ewing (1960).

The attraction of the sediments between 1.6 km and 4.2 km depth was also computed. The linear segment approximation tothe bathymetry was taken from the 1:2400000 bathymetric chart of Roberts, Hunter and Laughton (1979) and the isobaths, at 200 m intervals, were extended laterally to remove erroneous edge effects (Fig. 6.4). In the same figure the combined water and sediment body corrections (with respect to 2.8 g/cm^3) down to a depth of 4.2 km are plotted at 5 km intervals along the approximate position of the C 80-4 profile. Significant errors occur in the positions of these field points and the bathymetric contours because of the different reference scales of the Mercator projections of the published bathymetric chart and the track chart of this work. Nevertheless it is felt that the rough correction arrived at for the C 80-4 gravity profile still merits incorporation into the model

Thus in the top half of Figure 6.3 the observed gravity profile (solid line) is a Bouguer-type anomaly profile, as described previously in Chapter 4, and not a free-air anomaly profile - hence the high values on the ordinates. Accordingly the crust and mantle have only been modelled below 4.2 km. The simple shape of the top surface of the BVRS was obtained by straightforward depth conversion of the seismic basement on the C 80-4 seismic profile (Fig. 5.11) using velocities of 2.65 and 3.22 km/s for the post-R5 and pre-R5 sediments (see refraction station E10 of Fig. 5.14). Obviously the exact spatial relationships and flanking gradients cannot be found without accurate seismic migration but the simple geometry depicted serves to illustrate the three ridges of the BVRS and their lack of correlation with any disturbances in the anomaly profile (Fig. 6.3). The 10 mgal anomaly near the 140 km mark is the only slight evidence for any such correspondence.

A number of relevant observations and inferences can be made from the model. The uniformly thin crust (less than 8 km) and the shallow M-discontinuity beneath Rockall Trough as predicted by previous gravity and refraction information are confirmed again here. Similarly the apparent lack of any density contrast between the volcanic layer and the deep older sediments has been mentioned earlier. The density of 2.25 g/cm^3 advocated for both these layers seems rather low compared with, say, the density of fresh basalt or upper Cretaceous sediments which is nearer to 2.5 g/cm^3. If such a density were to be applied here then a concomitant lowering of the M-discontinuity and thickening of the crust would need to be invoked to balance the model.

The upwarping of the Moho favoured in the earlier C 80-1/C 84-4 2D gravity models (Fig. 6.2a) is also the preferred solution in this instance. Here the perturbance to the Moho is c. 70 km across, 20 km shorter than modelled in Figure 6.2a but this is not unexpected in view of the aerial distribution of the Barra anomaly. The amplitude of the perturbance, 2 km, and the dip of the flanking gradients, 11°, are the same as before and appear to produce a good fit between the observed and computed anomaly profiles (Fig. 6.3). The sharp changes in dip of the Moho around the edge of the supposed disturbance, depicted in the crude gravity model, are in reality likely to be rounded off to create a gentle, smooth up-domed region.

Lastly, a better indication of the overall shape of the presumed volcanic layer, especially its basal surface, is attained with the current model which crosses the Barra Volcanic ridges at nearly right angles (Chart 7) and which benefits from a less ambiguous seismic reflection profile; this comparedwith the prior margin-to-margin gravity model which traverses the ridges obliquely and where the seismic control is poorer. The thickening of the three ridges of the BVRS is in strong support of the loading and subsequent flexural bending of the thin Rockall crust. In excess of 4 km of volcanics, and quite possibly interbedded sediments, are modelled beneath Ridge 3 (Fig. 6.3) to account for the overlying gravity low, though it is still uncertain whether some contribution to the low is achieved by lowering the M-discontinuity by a small amount in addition. This author is in favour of a geometry whereby the highs and lows in the gravity field arise from concordant, gentle variations in the top of the crust <u>and</u> the Moho, rather than larger amplitude perturbances in just one of these density boundaries. The up-domed Moho beneath the Barra anomaly is, of course, the exception here.

The simple crustal and mantle geometry depicted in this model (Fig. 6.3) is remarkably similar to that proposed for the region around and beneath the Wyville-Thomson Ridge (Roberts et al. 1983; Bott and Smith 1984). In this area at the north-eastern termination of Rockall Trough a huge pile of volcanics at least 12 km thick, constituting Wyville-Thomson Ridge, causes a similar flexure in the

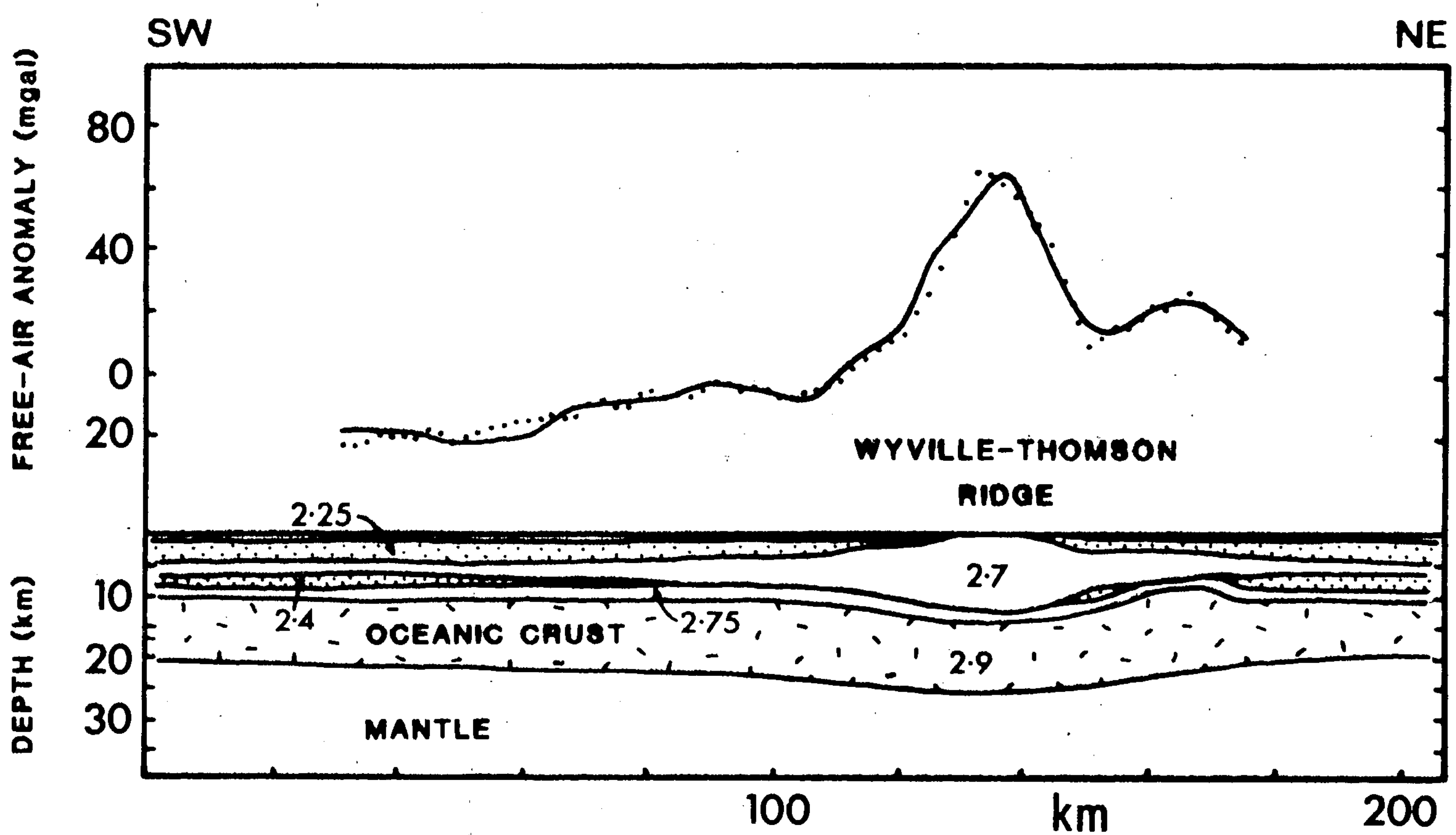


Figure 6.5 Two-dimensional gravity model across the Wyville - Thomson Ridge. Redrawn from Roberts et al. (1983). Solid line = computed anomaly; dotted line = observed anomaly; stipple = sediments; unornamented = volcanics; numbers are densities in g/cm^3.

subjacent (postulated oceanic) crust (Fig. 6.5). In both areas the wavelength of the flexure is not much greater than the width of the overlying volcanic pile. In the southern Rockall example, however, the model is not sufficiently refined and controlled to permit any confident predictions concerning the lateral extent of the flexure, or flexures. Roberts et al. (1983) suggested that the volcanic Wyville-Thomson Ridge sits on a thin layer of compacted sediments (Fig. 6.5). While this may well also be applicable in the case of the Barra volcanic ridges the seismic reflection and refraction surveys to-date do not provide any evidence in support, so the volcanic layer is assumed to rest directly on crustal basement. Note also the extension of a thin layer of volcanic material away from the ridges - a layer which probably corresponds to the deep layered seismic basement on the reflection profiles (see Chapter 5, section 5.1).

Challenger 84-7 gravity model

The C 84-7 track was observed during the Challenger 1/1984 cruise to provide a margin-to-margin gravity profile suitable for two-dimensional gravity modelling and to complement the long C 80-1/C 84-4 model profile discussed earlier in this section (Chart 7). As with the other models elsewhere in the Trough the approach to a two-dimensional condition here is not terribly good, though this author is satisfied that the models illustrated highlight the main geometries and thicknesses of the sediments and crust beneath the Trough and its margins. Three-dimensional gravity modelling across such a large, variable area is certainly impracticable.

The three C 84-7 gravity models presented (Figs 6.6a, b and c) each show the characteristic rapid distension of continental crust beneath the Trough margins and the abrupt lateral change into uniformly thin crust of enigmatic origin below the basin and its thick sediments. The structure and densities of the two continental margins are, unsurprisingly, very similar to those modelled earlier (Figs 6.2a and b). At the north-western end of the present model the stepped appearance of the free-air anomaly profile can be satisfactorily emulated in the calculated profile by modelling the two buried half-grabens visible on the C 84-7 seismic profile (partly on Fig. 5.21). At both this and the south-east margin the gradients of the buried continental slopes and the deep Moho appear to be markedly steeper than those modelled to the south-west. The dip angles of the continental slope and Moho below the north-west and south east margins are 16° and 33°, and 38° and 36°, respectively. Clearly it is these steeper inclinations which are responsible for the more pronounced free-air edge effects observed in the north-eastern part of the research area (Charts 2 and 7, back pocket), in particular the deep negative anomaly (-59 mgal) just north of the N. Porcupine Salient (Figs 6.1 and 6.6). The rather sudden appearance of these prominent gravity lows across a NW-SE trending line north of Porcupine Bank (Chart 2) seems to indicate a corresponding sudden variation in crustal structure to the north-east, that is, a more rapidly attenuated continental margin.

Beneath Rockall Trough the gravity anomaly field can be modelled closely by assuming a comparatively straightforward crustal layer with a thickness between 6 and 7 km (Fig. 6.6). The total width of thin crust here of 160 km should be compared with the 200 km width predicted in Figure 6.2, a value that is more representative of the true width of rifted or accreted crust within the Trough since the orientation of the C 80-1/C 84-4 profile is closer to the trend of roughly 105° thought to define the opening direction of Rockall Trough (refer to Chapter 8). One similarity between the models along these two profiles, however, is found in the occurrence of the top of the crust at a slightly deeper level on the north-west side of the Trough than on the south-east side - 8 km against 7.5 km in Figures 6.6a, b and c. As suggested earlier this disparity may be related to subsidence of the Rockall Plateau microcontinent since the early Tertiary.

The depression of the thin crust below the north-west continental rise, and to a lesser degree below the south-east rise, results in the development of thick sedimentary basins in front of the continental margins and also gives the impression of elevated crust in the middle of the Trough. Below the north-west flank of the Trough almost 6 km of sediments are indicated between the 65 and 95 km marks (Fig. 6.6); a measure of support for such thicknesses is provided by the C 84-7 seismic profile which shows an impressive deep sedimentary section across the whole Trough below which the seismic basement is rarely imaged (e.g. Fig. 5.21).

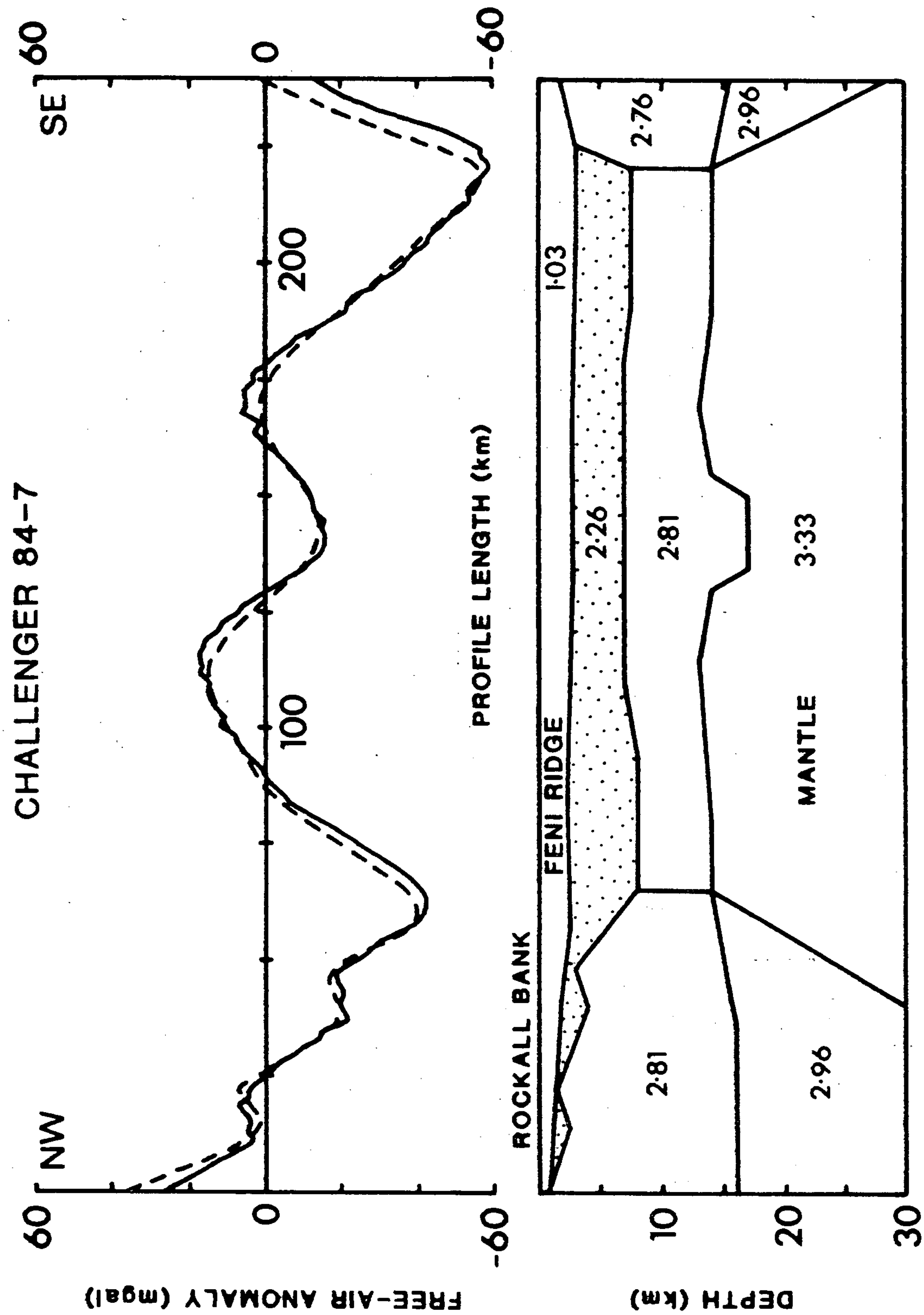


Figure 6.6a. Two-dimensional gravity model along profile C 84-7 with axial crustal thickening. See text for discussion and Chart 7 for location. Upper figure: solid line = observed anomaly; dashed line = computed anomaly. Lower figure: stipple = sediments; numbers are densities in g/cm^3.

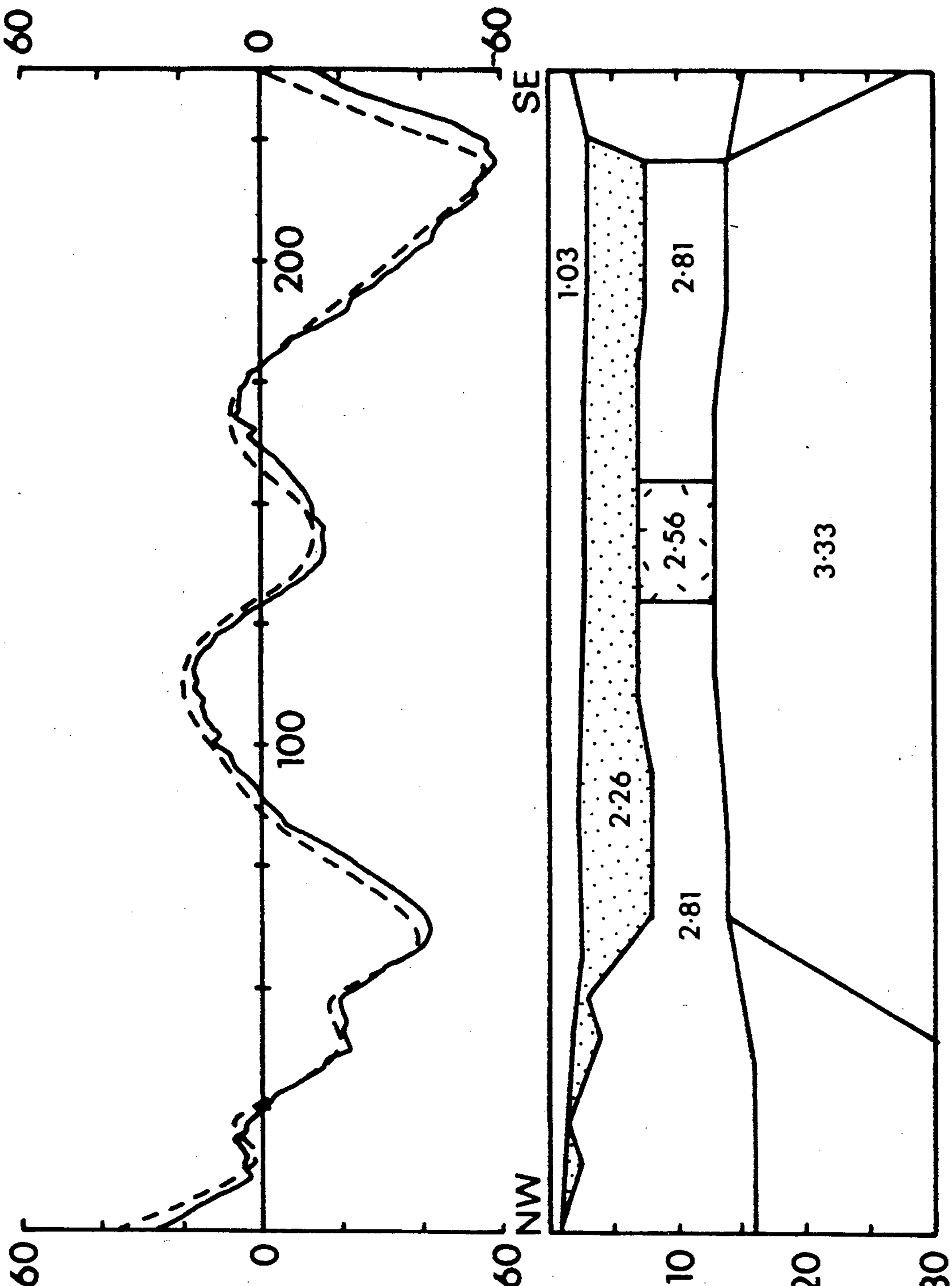


Figure 6.6b. Two-dimensional gravity model along profile C 84-7 with axial low density mass or intrusion. Ornament, etc. as for Fig. 6.6a.

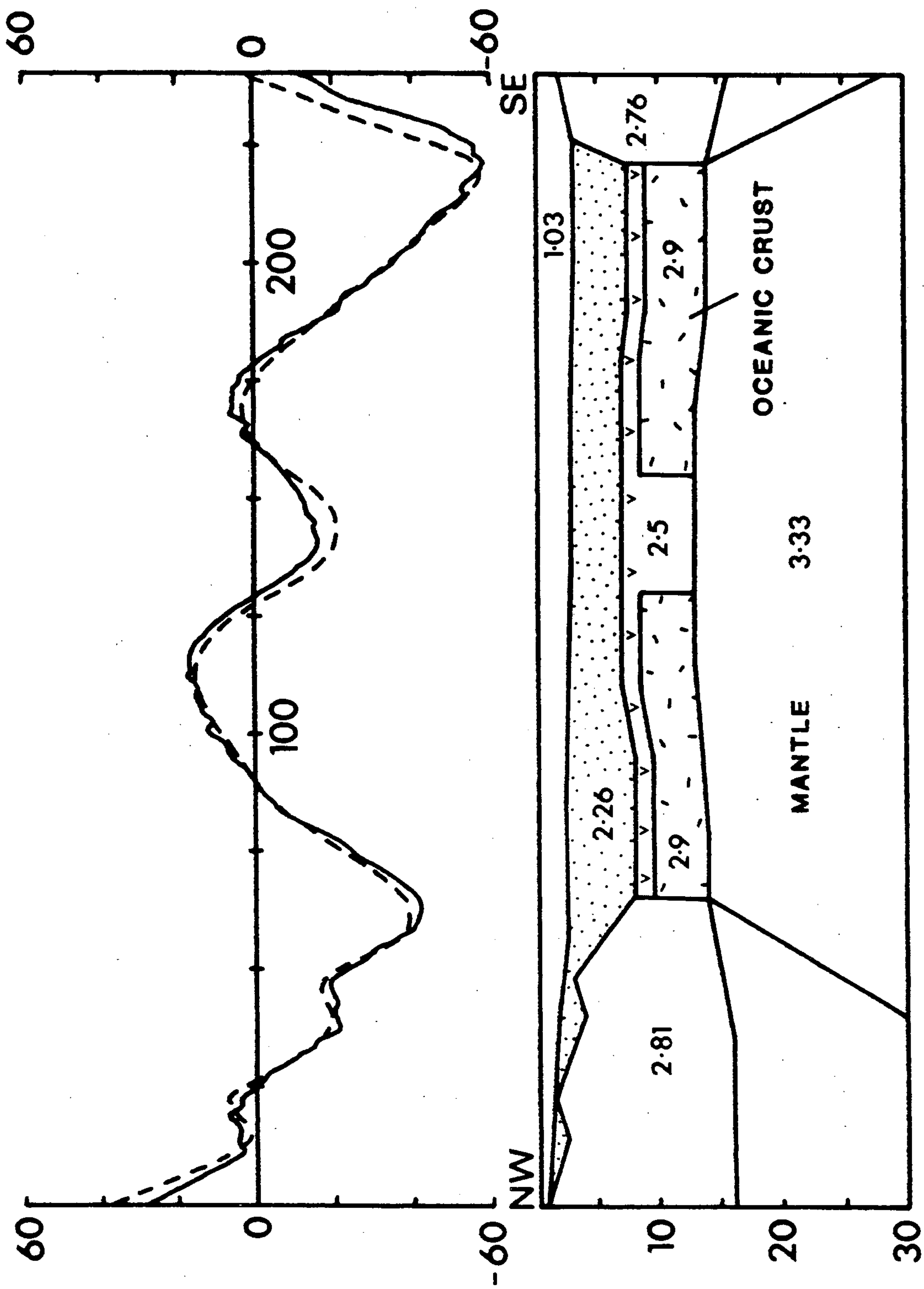


Figure 6.6c. Two-dimensional gravity model along profile C 84-7 with oceanic type crustal structure beneath Rockall Trough. Ornament, etc. as for Fig. 6.6a. V = oceanic layer 2.

The apparent elevation of the axial section of the thin Rockall crust is thought to be merely a corollary of the relative deepening of the adjacent crust, rather than actual shallowing of the central part which would be difficult to explain. The nearly symmetrical shape of both the thin crustal layer and its associated free-air anomaly is remarkable (Fig. 6.6) and is reminiscent of the oceanic crustal structure across the mid-ocean ridges. This analogy is further enhanced by the location of a supposed buried seamount (54.6°N, 14.7°W) in the exact centre of the Trough over the shallower crust (not illustrated here, but see Chart 4 and Figs 5.15 and 5.21). However, the 30-40 mgal relative gravity low in the centre of the Trough confuses what is otherwise a reasonably straightforward FAA profile and crustal structure. This curious gravity low is not observed to the south-west of the profile, in Southern Rockall Trough, but according to the free-air anomaly chart of Megson (1983) it does persist as a gently sinuous anomaly along the axis of central and northern Rockall Trough, suggesting that its source is regionally, not locally, developed and is probably related to the opening of the Trough.

Each of the three models developed in Figure 6.6 produces an excellent match between the observed and calculated anomaly profiles. But it is realised that any number of additional models could be constructed which satisfy the geophysical, though not necessarily the geological, requirements. In the first model (Fig. 6.6a) the gravity low is produced by thickening the axial 20 km or so of crust which has a continent-type density of 2.81 g/cm^3. No sensible geological explanation can be found for such an unusual crustal geometry. If it represented a block of less thinned continental crust then its persistent location at the axis of the Trough would be fortuitous indeed. The alternative explanation of the feature as a low density (high temperature?) cushion near the top of the mantle seems equally implausible by the same argument.

In the second model the gravity low is attributable to a block-like mass of low density within presumed stretched continental crust. If the mass represented some manner of large scale intrusion then the density of 2.56 g/cm^3 would favour a granitic composition. While an intrusion model is attractive in providing a source, at depth, for the volcanic seamount observed on the seismic profiles here (Fig. 5.21) there are difficulties again in explaining the lateral persistance of the anomaly. The third and final model (Fig. 6.6c) represents a variant of the previous model. Here an oceanic crustal structure with a 1.5 km thick Layer 2 (2.5 g/cm^3) and a 4.5-5.5 km Layer 3 (2.9 g/cm^3) has replaced the earlier continental crust. The central gravity low is caused by what is essentially an anomalously thickened Layer 2, a structure that is difficult to justify from actual geological examples from the oceans. Nevertheless the fit of the computed to the observed anomaly profiles is particularly good in this model.

Free-air gravity anomaly lows with amplitudes up to and in excess of 50 mgal have been reported across many active and inactive mid-ocean rifts (Watts 1982) and, as in the case of the C 84-7 model in Rockall Trough, they are frequently accompanied by flanking gravity highs. Watts (1982) also reported the 50 mgal gravity low over the long inactive Labrador Sea median ridge, which is presently buried completely by sediments and as such may be compared favourably to Rockall Trough. Tapponier and Francheteau (1978) and Watts (1982) proposed that the large free-air anomalies recorded over median ridges result from the mass deficit within the rift caused by the replacement of basalt with water (Fig. 6.7). This mass deficit is thought to be compensated isostatically by upward flexure of the lithosphere brought about by bouyancy in the underlying asthenosphere. But owing to the elastic thickness of the oceanic lithosphere, as little as 3-4 km below the actual rift according to Tapponier and Francheteau (1978), the compensation is spread over a broader region than the rift with the consequence that the shorter wavelength anomaly low remains.

It is interesting to speculate whether such an explanation can be applied to the C 84-7 gravity model, where the amplitude of the axial anomaly probably does not exceed 40 mgal (Fig. 6.6), and from there extended to the remainder of the Trough. The poor quality C 84-7 seismic profile fails to show up any seismic basement geometry that may represent the type of median ridge structure - rift shoulders, steep inward-facing escarpments and so on - that are required by the aforementioned hypothesis (Fig. 6.7). However, the multichannel seismic profiles GSI-1 and NA-1 to the south-west and north-east of the model, respectively (Chart 1), do image basement ridges with marked relief near the axis of, and sub-parallel to, Rockall Trough (Figs 5.5 and 5.6). Hence a remnant median rift may

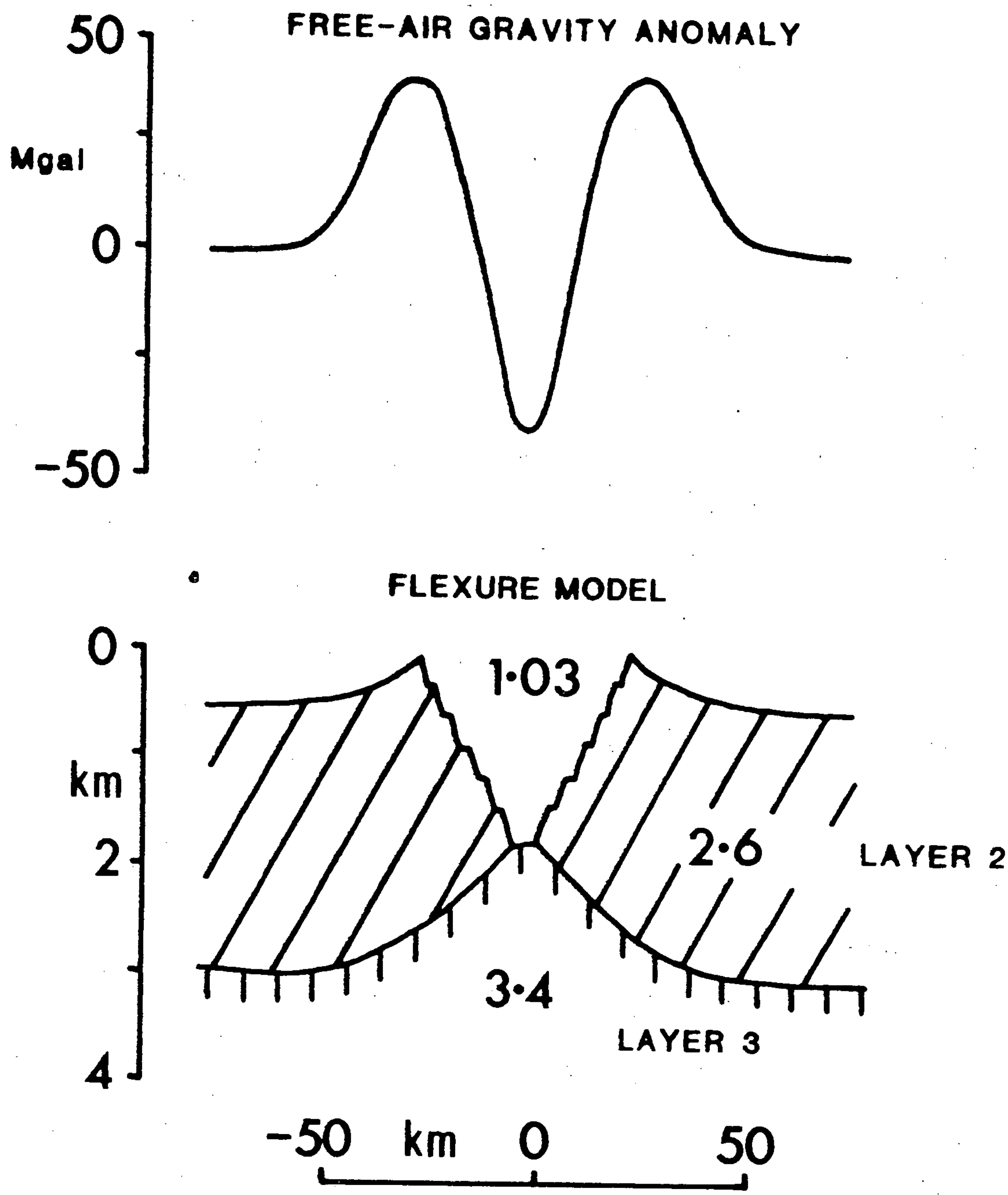


Figure 6.7 A model for free-air gravity anomaly lows over active and inactive spreading ridges. Redrawn from Watts (1982). Numbers in lower figure are densities in g/cm^3. See text for discussion.

persist under the C 84-7 gravity model which is not resolved on the seismic profile beneath the thick pile of younger sediments and the proposed seamount.

It is possible then that the free-air anomaly low occupying the axial portion of the Trough (Charts 2 and 7) may arise from an old deeply buried oceanic ridge system, and the peculiar density distributions illustrated here (Figs 6.6a, b and c) can be dismissed. The reduced amplitude of the axial Rockall Trough low compared with those observed over present-day active rifts (up to 100 mgal, Watts 1982) is then a result of the later infilling of the rift with sediment, bringing about a lowering of the lateral density contrast. Considering the gravity, seismic reflection and seismic refraction data together this simplified oceanic model is the one most favoured by the present author.

Charcot 1969 gravity model

A final two-dimensional forward gravity model was constructed along a track occupied by N.O. Jean Charcot in 1969 that passes through the centre of the pronounced negative anomaly north of Porcupine Bank (54°N, 14°W; Fig. 6.1 and Chart 7). Along-track free-air anomaly and bathymetric values were taken from the J. Charcot scientific cruise report (Scientific Group, 1971) but the accompanying reduced single-channel seismic records were not of use in defining any of the crustal interfaces. The initial sediment, crustal and mantle geometry was established through a consideration of the nearby seismic reflection and refraction data and also the C 84-7 gravity model a short way to the north-east. Adjustments were then made to this basic model to improve the fit of the observed and computed anomaly profiles (Fig. 6.8).

Despite the conspicuous size and amplitude of the negative anomaly on the contoured chart (Charts 2 and 7) it appears from the modelling that no special density distribution is required to account for the anomaly. This well developed free-air edge effect seems to be caused solely by the abrupt termination to the north-west of a plateau-like region in the continental margin which is manifested in the acoustic basement chart (Chart 4) and the FAA contour map (Chart 2) and which is called the North Porcupine Salient (Fig. 6.1). This drawing out of the continental margin has the effect of flattening

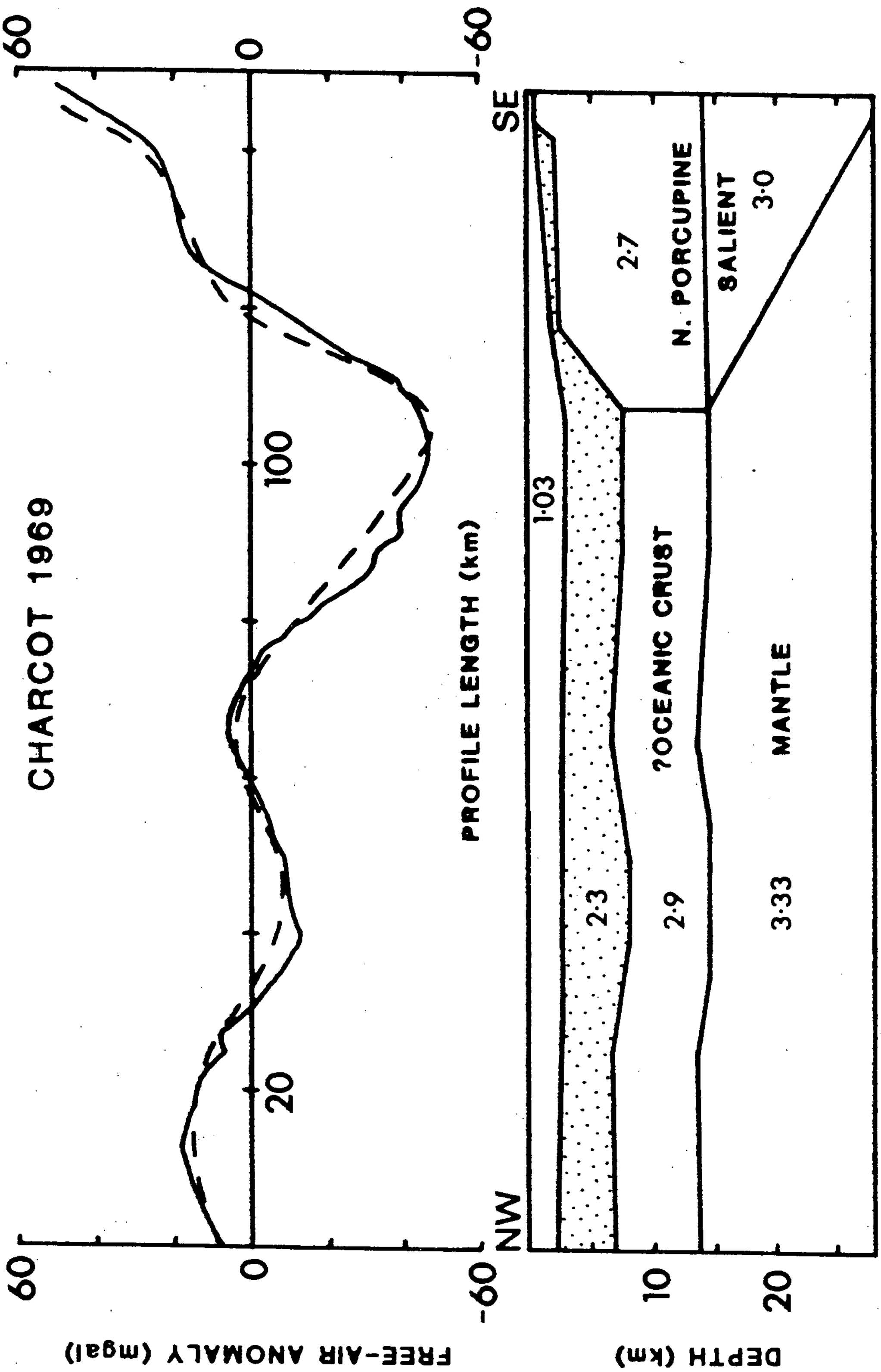


Figure 6.8 Two-dimensional gravity model along the Charcot 1969 profile across north Porcupine Bank. See Fig. 6.1 for location and text for discussion. Ornament, etc. as for Figs 6.2 and 6.6.

the dips of the continental slope (25°) and the Moho (16°) compared with the nearby C 84-7 gravity model (38° and 36°, Fig. 6.6). This marked change in the gradients of the continental margin over such a short distance supports the inference, drawn from the seismic reflection profiles, that a large transfer structure is present below the north-eastern boundary of the N. Porcupine Salient (Fig. 6.1).

Beneath the deeper Rockall Trough a typical thin crust is modelled of the order of 7 km and with an uncomplicated internal structure and oceanic-type density of 2.90 g/cm^3. The simple axial depression in the crust from 25 km to 60 km (Fig. 6.8), invoked to account for the shallow gravity low here, may represent the median ridge-type structure discussed and favoured for the previous gravity model.

7. ROCKALL TROUGH: TOTAL INTENSITY MAGNETIC ANOMALY DATA

7.1 Regional magnetic anomaly charts

Small scale magnetic anomaly maps of the Rockall province have been produced by Vogt and Avery (1974), Roberts and Jones (1975) and Max et al. (1982). Chart 3 (back pocket) of the present work was reproduced from the total intensity magnetic anomaly map of Roberts and Jones (1975) which in this area, southern and central Rockall Trough, is based largely on the closely-spaced (c. 11 n.mi.) sea-towed magnetic observations of Vogt and Avery (1974). The sparse and irregular spacing of the Edinburgh University magnetic anomaly data, together with their variable, often large cross-over errors, do not permit this author to construct an up-to-date magnetic anomaly map. Rough contouring and an accompanying discussion of the magnetic anomaly field around and south of the Charlie-Gibbs Fracture Zone and Clare Lineament were presented in Chapter 4 (see Fig. 4.12).

A noticeable contrast in magnetic character exists between the highly variable, short wavelength, large amplitude signature over the continental margin banks, and the more subdued, longer wavelength, low amplitude signature over the majority of Rockall Trough (Chart 3). To a large degree this must reflect the increased depth to the basement beneath Rockall Trough (Chart 4, back pocket). But it is unclear what effect the locally thick sediments in the basin might have in attenuating the magnetic field here. The above-mentioned magnetic contrast is well illustrated across the Rockall Bank margin (Chart 3) where detailed surveys have been undertaken (Roberts and Jones 1978). The magnetic geology of Porcupine Bank and Seabight and the shelf west of Ireland has been discussed by Gray and Stacey (1970), Riddihough (1975), Riddihough and Max (1976) and Lefort and Max (1984); little can be added to their comments from the present work, which is mainly peripheral to these areas.

The roughly circular or oval positive, large amplitude (up to 900 nT) anomalies over Porcupine Bank (52°N-53°N, 14°W) have been attributed to central-type igneous centres, possibly of Tertiary age, within the upper crust (Riddihough and Max 1976, Megson 1983). Two multichannel seismic profiles, WI-8 and WI-10 (Chart 1, back pocket), that pass over the ends of two of these positive anomalies show no direct evidence of any intrusive bodies. Instead there appears to be a correlation with shallow, faulted anticlines, which themselves could be an expression of deeper large-scale intrusion. Megson (1983) proposed palimpsest control on the location of the igneous centres by SW-NE trending Caledonide lines of weakness, some measure of support for which is provided in the orientation of the prominent SW-NE basement graben between the anomalies (refer to Chapter 5 and Fig. 5.2).

Within Rockall Trough the quiet magnetic field previously mentioned is disturbed by a conspicuous arcuate zone of large positive anomalies - the Barra magnetic anomalies (Chart 3). On Chart 3 a maximum peak-to-trough amplitude exceeding 800 nT is associated with the pronounced elongate anomaly at 54°N 17.5°W. The magnetic anomaly data for the present study and the magnetic compilation map of Max et al. (1982) both indicate a maximum amplitude of 1000 nT or more at the same location (e.g. Fig. 7.3), this disparity presumably reflecting heading and navigation errors, etc. within and between the different surveys.

A very close spatial relationship between the Barra magnetic anomalies and the underlying Barra volcanic ridges is evident when the seismic basement distribution around the BVRS (see Chart 4) is superimposed on the magnetic anomaly map (Fig. 7.1). This close correlation is particularly well brought out by the distribution of seismic basement shallower than 5.0 s TWT at the northern end of ridges 1, 2 and 3 of the BVRS. There appears to be a strong correspondence between areas of marked basement relief and the location of the larger magnetic anomalies, as strikingly illustrated on the S 79-14 (Fig. 5.3), C 80-2 and C 80-4 (Figs 5.10 and 11) seismic profiles.

One possible exception to this observation may be seen in the c. 400 nT anomaly at 53.6°N 19°W (Chart 3, Fig. 7.3) where no corresponding basement ridge is visible on the C 84-3 seismic profile which passes approximately over its centre. Unfortunately the poor quality of the profile and the thick sequence of sediments here preclude any confident seismic interpretation. The absence of an obvious basement ridge here is at first surprising in view of the fact that the lesser (200 nT) anomalies at around 53.5°N on ridges 2 and 3 of the BVRS are underlain by noticeable basement relief. Hence

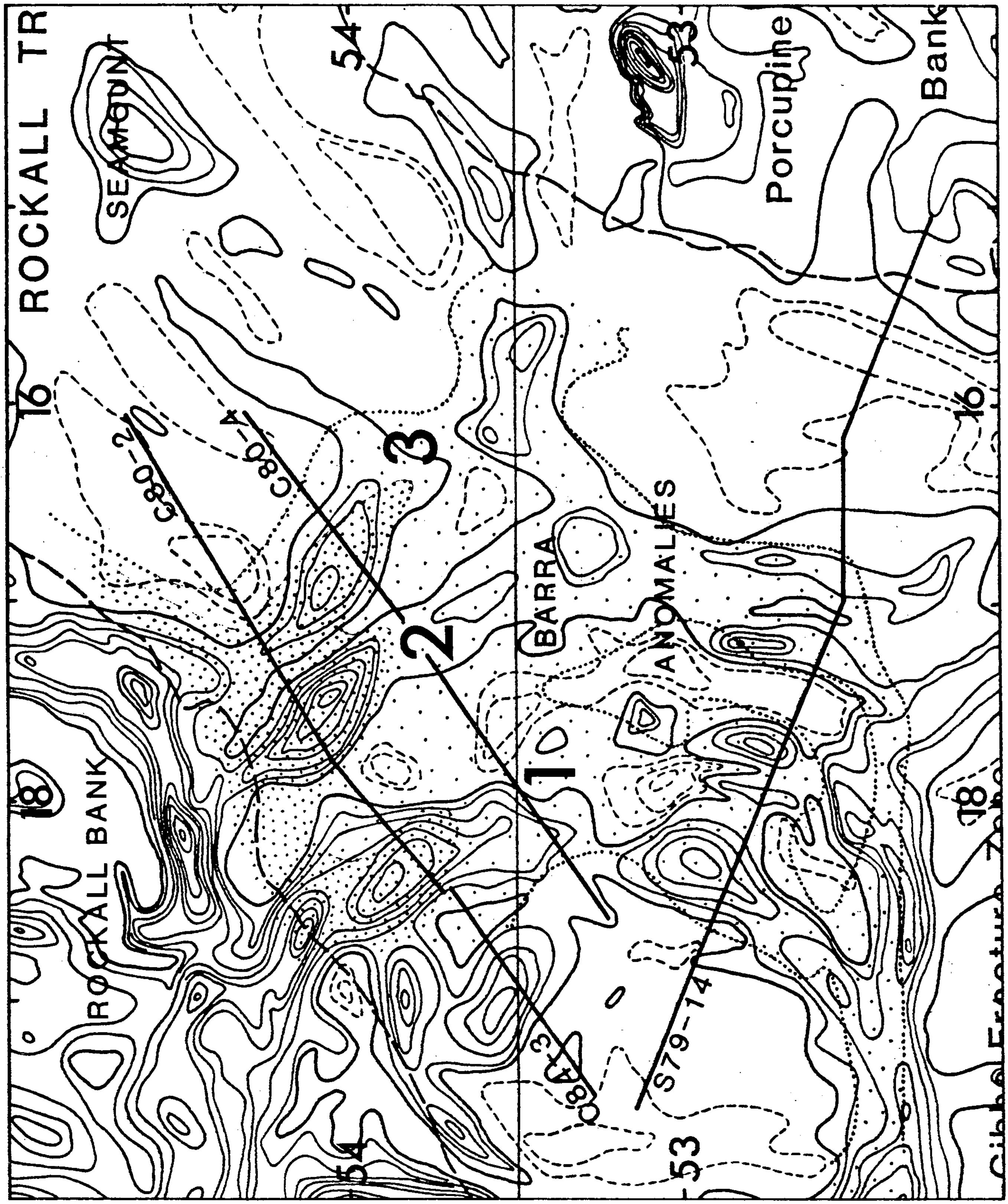


Figure 7.1 Total intensity magnetic anomaly map of southern Rockall Trough. Refer to Chart 3 (back pocket) for location of map area. Contours at 100 nT intervals except over Rockall Bank to NW (200 nT). Bold line = OnT. Dashed line = negative contours. Light stipple = outline of BVRS (ridges 1,2 & 3 marked). Heavy stipple = BVRS shallower than 5.0 s TWT.

the natural conclusion to be drawn here is that some contribution to the magnetic anomalies is made from outwith the volcanic ridges, almost certainly at depth below them. There does not appear to be any resolvable contribution to the magnetic field from the extensive intrusive sills situated over the Barra volcanic province (see Chart 6).

Just as the isochrons on seismic basement showed the apparent continuation of volcanic basement up onto Rockall and Lorien Banks (Chart 4 and Chapter 5), so the magnetic anomaly map seems to predict the persistence of the Barra anomalies into south-western Hatton-Rockall Basin via Lorien Channel (Chart 3, though clearer on original map of Roberts and Jones 1975). Such a continuity of magnetic signature is confirmed by unpublished detailed aeromagnetic data over south-west Rockall Plateau (D.K. Smythe, pers. comm.) which indicate a pattern of smooth S-shaped lineations extending from the Trough north-west into Hatton-Rockall Basin. Smythe (pers. comm.) believes that these proposed curved lineations may represent some form of oceanic-type magnetic striping. As such they may be compared nicely to the S-shaped sea floor spreading lineations of a similar scale reported over Iceland and its insular shelf by Bott et al. (1983, accompanying map). However examples such as this are rare in the ocean tectonic and magnetic environment.

The present author believes that the lineations are not continuous but each consist of a series of irregular oval positive anomalies connected by narrow cols, reflecting the distribution of the underlying volcanic ridges and perhaps some deeper intrusive sources. The broad spacing and variable orientation of the available seismic profiles, and hence the control on basement distribution, could explain why some of the anomalies do not appear to have associated with them similar-shaped relief (in plan view) in the subjacent volcanic basement. The slight variations in shape and amplitude of the positive Barra anomalies between the Trough and the Plateau are thought to reflect the differences in basement depth in the two areas (D.K.Smythe, pers. comm.). D.K.Smythe's upward continuation studies of the Barra magnetic anomalies and their counterparts on Rockall Plateau are in support of such an explanation. Thus there is a strong suggestion that the distinct rift below Lorien Channel, the volcanic ridges, and the associated positive magnetic anomalies are contemporaneous and very likely caused by the same tectonic event.

At their southern end the Barra magnetic anomalies curve abruptly but smoothly into the Charlie-Gibbs F.Z., as was mentioned in Chapter 4 (Chart 3; Fig. 4.12). Further to the west, around 52.5°N 20.5°W, another anomaly exceeding 700 nT and similar in shape though larger than the Barra magnetic highs, curves away from the CGFZ to connect up with a conspicuous magnetic feature considered to mark the buried extension of Lorien Bank (the extreme southern tip of Rockall Plateau at 53°N). The single-channel Discovery 84 seismic profile (see Chart 1, back pocket) that crosses the centre of this broad anomaly roughly at right angles shows that it is underlain by highly reflective, irregular, peaked seismic basement identical to that seen along S 79-14 and other profiles to the east.

Both the magnetic anomaly map and the seismic basement chart strongly suggest that the BVRS, Ridge N of the CGFZ (Fig. 4.6), the presumed volcanic ridge south of Lorien Bank and their over-lying positive magnetic anomalies are all continuous, together forming an almost complete, roughly circular or annular volcanic province. If this is correct then it is quite possible that the whole Barra volcanic and magnetic system is contemporaneous with, and probably a direct result of, the early development of the CGFZ in the Late Cretaceous. This implies the formation of the arcuate Barra system soon after the completion of rifting and any oceanic accretion in southern Rockall Trough and the initiation of the fracture zone. The important consequences for the age of the Trough and the seismic stratigraphy are considered further in the final discussion (Chapter 8).

The east-west trending 200 nT anomaly at 53.5°N 16°W appears to be connected to the curvilinear anomaly high overlying ridge 2 of the BVRS (Fig. 7.1), although the seismic basement chart appears to favour continuity with the pronounced ridge 3 to the north-west. Despite the fairly good seismic reflection coverage here and the congruency of the magnetic and basement contours it seems that the distribution of the basement depicted may be somewhat of a simplification of the true pattern in this area. The rather similar 200 nT anomaly about 50 km to the east, overlying the base of the continental slope, has no discernible expression on the seismic profiles in its vicinity.

Within the Rockall Trough and away from the prominent Barra magnetic anomalies the magnetic field is mainly quiet, rarely exceeding -300 nT and only locally -200 nT. The proposed buried seamount centred on 54.6°N 14.7°W, as evidenced from the seismic profiles here, has a corresponding isolated positive anomaly whose shape, amplitude and gradients are similar to those observed over the BVRS (Chart 3). The small-amplitude, annular positive anomaly at 55°N 13°W is also thought to represent a buried seamount but this has yet to be confirmed by seismic profiling. Whereas the westerly of the two seamounts is clearly normally magnetised the feature to the east is characterised by a central negative component analogous to the magnetic signatures over the axially-placed seamounts further north-east in the Trough, e.g. Hebrides Terrace and Anton Dohrn seamounts. Such magnetic signatures may be indicators of reversely magnetised seamounts. Magnetic modelling of the seamount at 14.7°W was not feasible owing to the lack of constraints on the buried shape and magnetic properties of the body.

There is a tendency towards simple magnetic lineations in the region immediately north-east of ridge 3 of the BVRS (Chart 3; Fig. 7.1). How much this reflects subjectivity in the contouring of the magnetic anomaly map is unknown. Support for their authenticity comes from the total field magnetic map of Max et al. (1982) which picks out SW-NE oriented lineations in precisely the same area. In addition the lineations appear to correlate in a general manner with the underlying basement ridges of similar trend (Chart 4). The symmetrical disposition of magnetic highs and lows about an axial linear is worthy of note (Fig. 7.1) and reminiscent of weakly developed ocean floor striping. However their lack of persistence further north-east along Rockall Trough is perplexing in such a context; it is equally likely that the linear anomalies result solely from the topography on the subjacent basement ridges.

The quiet magnetic field over much of Rockall Trough has popularly been attributed to the formation of oceanic crust beneath the basin during the long period of normal polarity (Anomaly 34) in the Late Cretaceous, hence accounting for the absence of pronounced magnetic reversals. Roberts et al. (1981) extended this argument and suggested that the lack of any coherent magnetic pattern here can be ascribed to diffuse oceanic accretion in a Gulf of California style spreading environment. Such an interpretation may be applicable in the central and northern parts of Rockall Trough where multichannel seismic profiles image a poorly defined seismic basement produced, perhaps, by interbedding of sediments, sills and volcanics. On the basis of magnetic modelling Megson (1983) believed the quiet magnetic field reflects the presence of highly stretched and thinned continental crust beneath the Trough, with the Barra anomalies being produced by large normally magnetised intrusions within the crust. The origin of magnetic quiet zones in the oceans has long been a source of controversy, many potential explanations for which are currently in vogue (refer to Barrett and Keen (1976) for a good review of these). The elegant model expounded by Roots et al. (1985) for the development of quiet magnetic zones adjacent to passive margins may be applicable to some extent in Rockall Trough.

If the basement beneath the Trough comprises oceanic crust then there is no obvious magnetic edge effect anomaly at the boundary with the continental margins as might be expected, such as is recorded off the eastern seaboard of North America (the East Coast Magnetic Anomaly). In fact the steep continental margin of Porcupine Bank is not well marked on the magnetic anomaly maps. The N. Porcupine Salient, prominent on the seismic basement and free-air anomaly charts (Fig. 6.1), is surprisingly not manifested in the magnetic field other than as a gentle northerly bulge of the contours on the maps of Roberts and Jones (1975; Chart 3) and Max et al. (1982). However, rough contouring of the Edinburgh magnetic data in this area, despite the large cross-over errors between the profiles, highlights a magnetic gradient from 50 to -150 nT paralleling the steep basement slope at the front of the salient.

The edge of the continental margin below Rockall Plateau is comparatively well defined, especially on the south-eastern side of Lorien Bank (Fig. 7.1), perhaps as a consequence of the shallow Tertiary volcanics thought to be present here (Roberts 1975). The broad area of low amplitude positive anomaly projecting beyond the 2000 m isobath between 14°W and 16°W correlates closely with the proposed intra-sedimentary lavas interpreted from the seismic profiles across this margin (Figs 5.5 and 5.6) - hence refuting the existence of the Jean Charcot Fault Zone and its associated buried continental platform (see Chapter 5).

7.2 Forward and inverse two-dimensional magnetic modelling

Forward magnetic anomaly modelling

Fairly rudimentary forward magnetic modelling was performed along two geophysical profiles traversing the northern BVRS in a SW-NE orientation - the C 80-4 and combined C 84-3/C 80-2 profiles (Fig. 7.1). Clearly a two-dimensional magnetic field does not prevail on either side of these profiles but the absence of close control on the shape and magnetic properties of the source bodies precludes any meaningful three-dimensional modelling. Nevertheless the two profile models presented provide some indication of the general geometry of the magnetic sources beneath the BVRS. None of the remaining magnetic profiles constituting the data base of this research were considered suitable for modelling the anomaly field elsewhere in the Trough or across its margins.

In the magnetic model along profiles C 80-4 (Fig. 7.2) the geometry of the presumed volcanic basement and also the underlying thin crust was copied exactly from the 2D gravity model along the same track (Fig. 6.3). A surprisingly good match to the positions and inflections of the Barra anomalies was achieved by simply assigning intensities of magnetisation of 5 A/m and 2 A/m to the volcanic basement and thin crust respectively (remanant inclination = 50°), before any deeper magnetic sources were introduced into the crust. In common with the magnetic models across the Charlie-Gibbs F.Z. and Clare Lineament the sediments and upper mantle were assumed to have a negligible induced magnetic intensity.

The irregular volcanic basement (Fig. 7.2) is considered to have a high Koenigsberger ratio, with the remanant magnetic vector having an inclination of 50° and declination 0°. This angle is compatible with its "freezing in" sometime during the Late Cretaceous, as discussed in Chapter 4, although varying the angle by $\pm$ 10° does not alter the calculated anomaly significantly. The magnetic intensity of 5 A/m used in both models (Figs 7.2 and 7.3) is a value commonly assigned to the volcanic oceanic layer 2 in the literature (e.g. Harrison 1981; Roots et al. 1985; and Scrutton 1985). The thin crust beneath Rockall Trough is modelled as also having a remanant magnetic inclination of 50° but an intensity of 2 A/m. The latter value was introduced to represent what this author believes to be oceanic crust beneath the Trough, though such a value is necessarily very much a simplification of the true magnetic structure and distribution.

In the C 80-4 magnetic model (Fig. 7.2) the piles of volcanics constituting ridges 1 and 3 of the BVRS account for all but about 200 nT of the positive anomalies at the 30 and 135 km marks. Ridge 2 of the BVRS appears to satisfy most of the magnetic peak at 105 km but not the smaller peak to the south-west at 90 km. To achieve a good fit of the observed and computed magnetic anomaly profiles at these three main peaks vertical-sided bodies were introduced into the thin crust of the Trough below the volcanic ridges. Their remanant magnetic intensity and inclination of 5 A/m and 50°, respectively, in broadly satisfying the observed anomaly profile, suggest consanguinity with the shallow volcanic basement.

The shape and size of the strongly magnetic sources within the thin crust and their location directly beneath the thickest development of shallow volcanic basement (Figs 7.2 and 7.3) favours the interpretation that they are intrusions which acted as the sources for the overlying Barra volcanic ridge system. The same conclusion was drawn from the magnetic anomaly modelling across the CGFZ-Clare Lineament (Figs 4.13 and 4.14). The high magnetisation of these supposed intrusive bodies implies a mafic or ultramafic composition, an inference that may provide an explanation for the absence of gravity anomalies over the same features - since there would then be little or no lateral density contrast with the postulated contiguous oceanic crust of the same composition. But it is then necessary to invoke differing magnetic source properties to account for the magnetic anomalies described here.

The alternative interpretation, as described earlier in Chapter 4, assumes that the deeper magnetic sources are the result of pervasive dyking and minor intrusion, rather than single large bodies. The linear outline of some of the Barra anomalies, around 52.5°N 17°W for instance, favours the latter interpretation. But

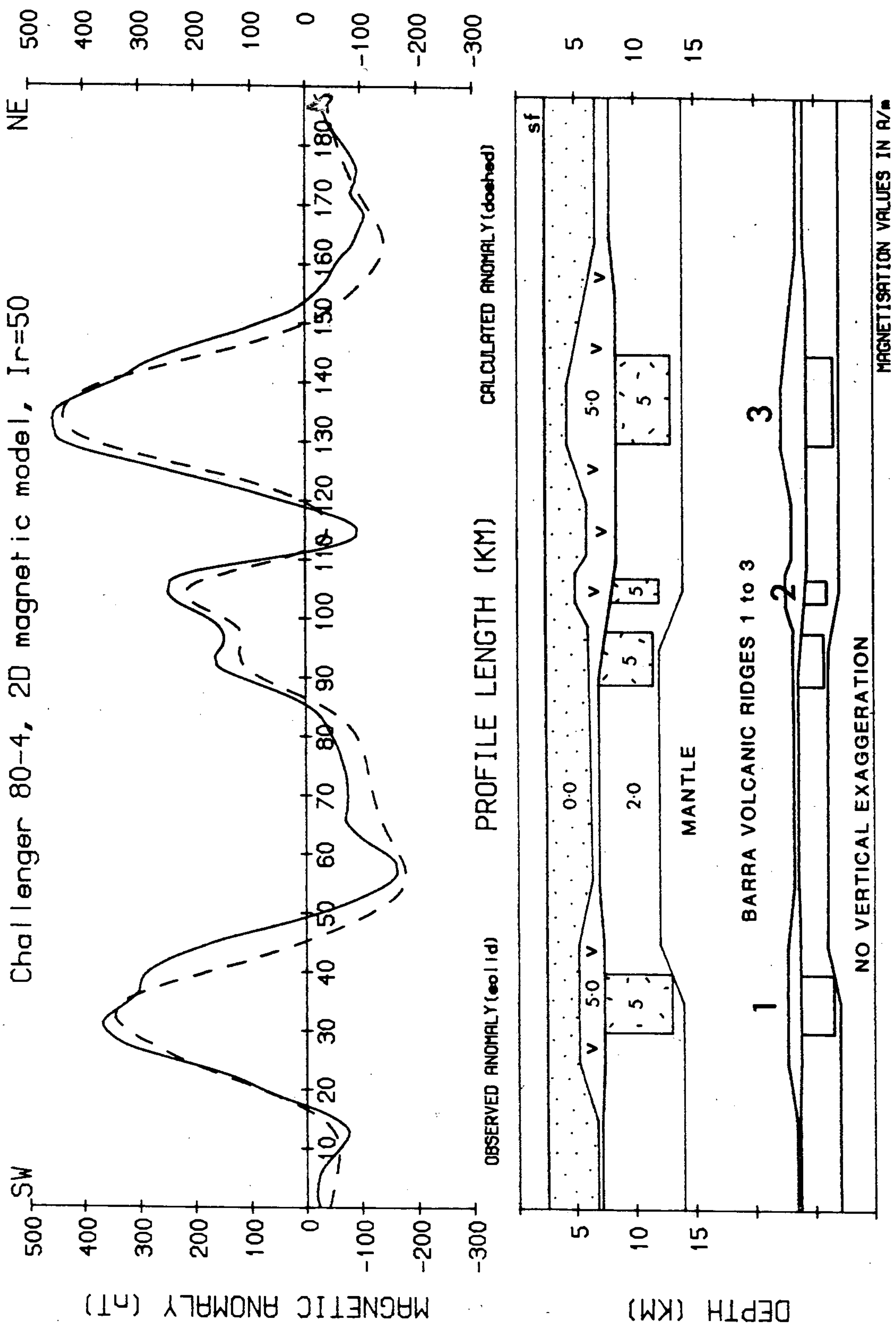


Figure 7.2 Two-dimensional magnetic anomaly model along profile C 80-4. See Fig. 7.1 for location. Lower figure: stipple = sediments; V = volcanic basement; dashed prisms = intrusive bodies; numbers in prisms are magnetic intensities in Amp/metre. Model redrawn at no vertical exaggeration at bottom of figure.

elsewhere the anomalies take on a circular or elliptical outline, suggesting that the deep crustal magnetic sources could well be single intrusions with a similar shape in plan view (Fig. 7.1).The occasional mis-matches between the gradients of the observed and computed anomaly profiles (Fig. 7.2) probably reflects the poor seismic control on the exact geometry of the volcanic basement. Likewise the deeper intrusive zones are certainly not regularly shaped as depicted, more likely they may broaden with depth.

In the second forward magnetic model along C 84-3/C 80-2 (Figs 7.1 and 7.3) no control on the general shape and thickness of the volcanic and crustal layers was obtained from gravity modelling, owing to the proximity of the track to the free-air edge effect over Rockall Bank continental margin. As in the prior C 80-4 model the top surface of the volcanic basement was approximated by performing a simple depth conversion of the two associated seismic profiles according to the refraction data of Ewing and Ewing (1959; station E10, Fig. 5.14). Obviously the basement reflectors would need to be migrated to establish the precise location of the Barra volcanic ridges. Despite this simplified basement topography a good fit between the observed and computed anomalies is obtained by applying a similar geometric pattern of volcanic ridges with underlying zones of intrusion as developed for the first model (Figs 7.2 and 7.3). The H1-3 refraction stations of Hill (1952) provide a small element of control at the beginning of profile C 80-2 (64 km mark in Fig. 7.3).

The subtle inflections and gradients (skewness) in the higher parts of the Barra anomalies, noticeable over the 1 and 2 ridges of the BVRS (Fig. 7.3), are imitated reasonably well in the calculated anomaly profile when just the volcanic ridges are modelled. However, when the underlying intrusive zones extending down to the M-discontinuity, are included in order to improve the fit of the anomaly amplitudes these subtle changes become lost in the large peaks. This may indicate that the magnetic contribution from the shallow volcanics is greater than depicted in the two models (Figs 7.2 and 7.3), and that the obscuring effects of the deeper sources is correspondingly less. A higher magnetic intensity of 6 A/m seems to be demanded in this second model to satisfy the observed profile, compared with the value of 5 A/m employed in the first. It seems unlikely that any such variations in intensity prevail in the BVRS igneous province so some minor adjustments may be necessary to one or

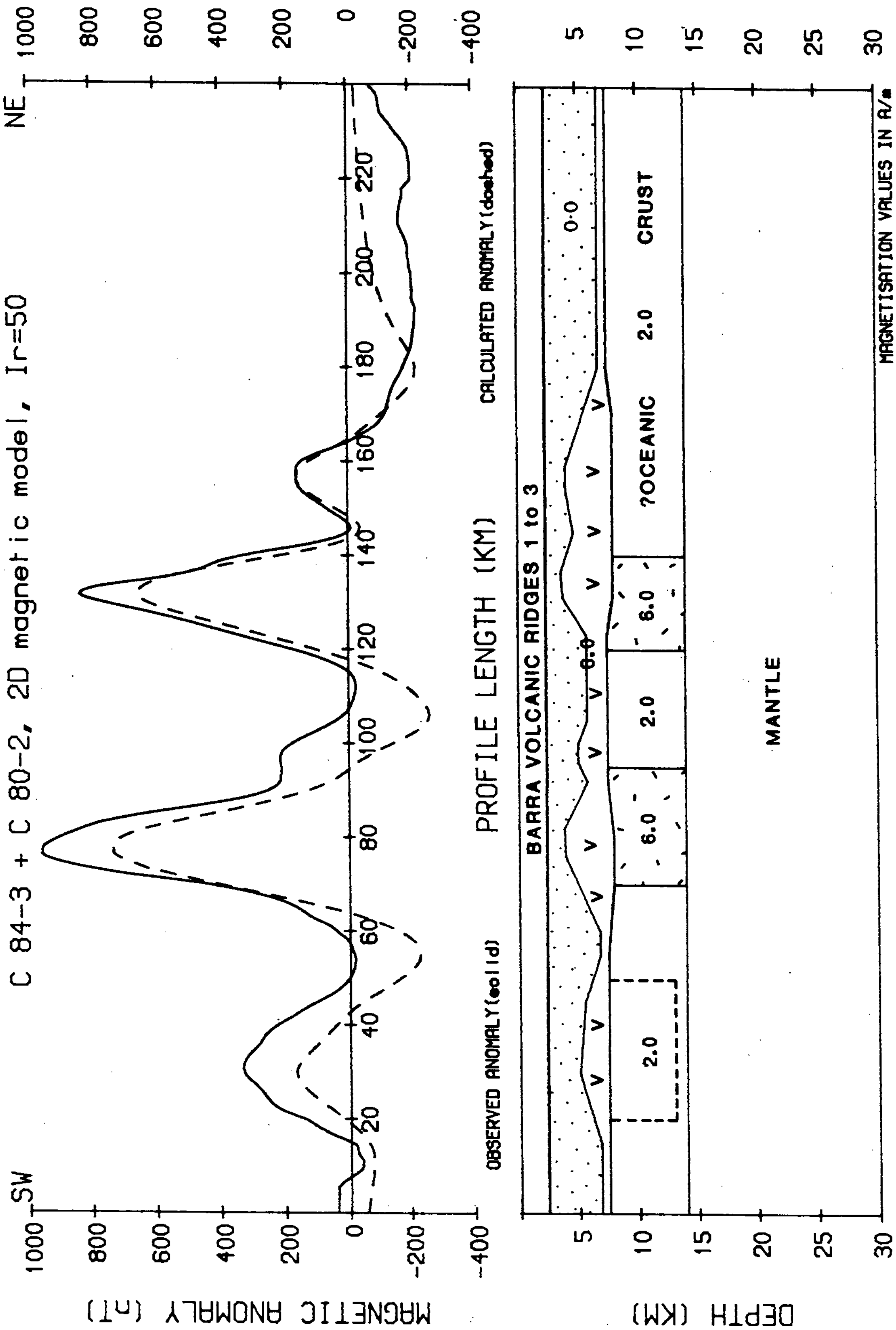


Figure 7.3 Two-dimensional magnetic anomaly model along profiles C 84-3 and C 80-2. See Fig. 7.1 for location of section. Ornament, etc. as for Fig. 7.2. Refer to text for discussion.

other of the models. Whichever of the two models (Fig. 7.2 or 7.3) is correct, or whether there is a compromise somewhere between the two, is uncertain. It is clear, though, that the Barra anomalies can be modelled satisfactorily by giving the volcanic ridges, as defined by the seismic reflection and gravity data, a normal magnetisation of 5-6 A/m, and by introducing cylindrical or prolate zones of intrusion into the underlying crust which have a similar remanant magnetisation.

In the C 84-3/C 80-2 magnetic model (Fig. 7.3) it was found impossible to precisely match the observed and calculated anomaly profiles. However, since the observed profile is reduced relative to a time-varying magnetic reference field it is only important to achieve a good fit of the relative amplitudes and gradients of the two profiles, and not their absolute levels. The c. 350 nT positive anomaly above the 30 km position can be modelled reasonably closely by thickening the shallow volcanic basement as depicted, or by including a further large intrusive zone at depth within the thin crust (dashed lines in crustal model). Although the C 84-3 seismic profile along this section of the model is of poor quality, it is unlikely that an additional volcanic ridge is present here with the relief illustrated (almost 2 km); such a ridge would certainly be resolved on the reflection profile. It is more probable that a low relief volcanic ridge or oval mound is succeeded below by a broad zone of intrusion in the same fashion as further north-east. Otherwise it is possible that the anomaly here arises solely from intrusive rocks within the crust; but then one would expect to see extrusive rocks associated with the sizeable intrusive zone, such as predicted by the Figure 7.3 model.

Inverse magnetic anomaly modelling

Following the detailed sea-towed magnetic surveying in the Northeast Atlantic reported by Vogt and Avery (1974) a number of workers, including those just mentioned, have identified the Barra anomalies of this work as potential oceanic magnetic reversal lineations and appropriately discussed the existence of oceanic crust beneath Rockall Trough and its evolutionary place in the early drifting and fragmentation history of the North Atlantic continents (Le Pichon et al. 1977; Kristoffersen 1978; Srivastava 1978). A more detailed discussion of the magnetic signature of Rockall Trough was presented by Roberts et al. (1981) who postulated that the SSW-NNE trending lineations constituting the southernmost Barra anomalies of this work (Fig. 7.1) represented pre-Anomaly 32 or 34 (70 or 83 m.y. B.P.) oceanic reversals. The NW-SE trending anomalies at the northern end of the Barra province were thought to indicate fracture zone trends within the Rockall Trough oceanic crust. The age and geometry of rifting and spreading that Roberts et al. (1981) developed has since been adopted and extended on a more regional basis by a number of workers without the benefit of additional geophysical data (Hanisch 1984; Price and Rattey 1984).

To test these theories of oceanic accretion and formation of magnetic reversal lineations this author performed inverse modelling of two long, representative magnetic profiles across the main parts of the BVRS, tracks S 79-14 and C 84-3/C 80-2 (Fig. 7.1). The Fortran computer program used to calculate the inverse models is fully listed and described by Nunns (1980), based on the inversion procedure developed by Parker and Huestis (1974). Essentially the program calculates the distribution of magnetisation along the observed profile given the following simplifying assumptions:

i) the source layer for the magnetic anomalies, Layer 2 of oceanic crust in this instance, is of uniform thickness below the input basement surface topography.
ii) the crustal structure and magnetic anomaly lineations extend to infinity on either side of the profile, i.e. a two-dimensional condition prevails.
iii) the remanant magnetic inclination and declination remain constant except for exact 180° reversals.
iv) the intensity of magnetisation within the source layer varies horizontally but not vertically.

The implications of these assumptions for the models, and the use of numerical filtering to remove the effects of diurnal variations, etc. on the observed anomaly profile, and so on, are considered at length by Nunns (1980) and need not concern us unduly in these comparatively simple models.

In both the S 79-14 inverse models (Figs 7.4 and 7.5) the presumed volcanic basement observed on the seismic profiles is considered to represent the oceanic magnetic source layer, as predicted by Roberts et al. (1981) and other workers. The top surface of this basement was approximated to by a rough conversion of two-way travel times along the S 79-14 profile. A remanant magnetic inclination of 50° was used to be consistent with the earlier forward magnetic models. No declination information is available so a value of 0° was used.

In the three C 84-3/C 80-2 inverse models (Figs 7.6 to 7.8) the same remanant magnetic vector was used and the magnetic source layer topography was found in the same manner as for the forward model of the same profile (Fig. 7.3). For the inverse models of both profiles, S 79-14 and C 84-3/C 80-2, the observed and recomputed anomalies (panel b of Figs 7.4 to 7.8) are replotted about their mean observed values, such that there is an equal distribution of positive and negative anomalies. The magnetic inversion program automatically recomputes a calculated anomaly profile (in the forward sense) for comparison with the observed profile in order to test the plausibility of the inverted magnetisation distribution.

Both the Shackleton and Challenger magnetic anomaly profiles were initially inverted assuming a 500 m thick oceanic Layer 2, a value that is currently widely used in magnetic studies of the oceanic crust. With the remanant inclination set at 35°, in imitation of the same value used by Roberts et al. (1981) in their forward modelling, maximum absolute intensities of magnetisation of 27 and 35 A/m were calculated, respectively, for the S 79-14 and C 84-3/C 80-2 models. Additional inverse modelling using inclinations of 50° and 60° reduced these values only slightly and it became clear that magnetic intensities realistic of those measured from oceanic crustal samples could not be attained with a 500 m thick source layer.

Consequently in the four models depicted here (two for each profile; Figs 7.4 to 7.7) the source layer was thickened to 1.5 km and 3.0 km. A remanant inclination of 50° was used in each model as this value is fairly representative of the Late Cretaceous magnetic vector in north-west British Isles (McElhinny and Cowley 1980). The angle of 35° used by Roberts et al. (1981) to represent the upper Cretaceous is certainly too shallow since it corresponds to a

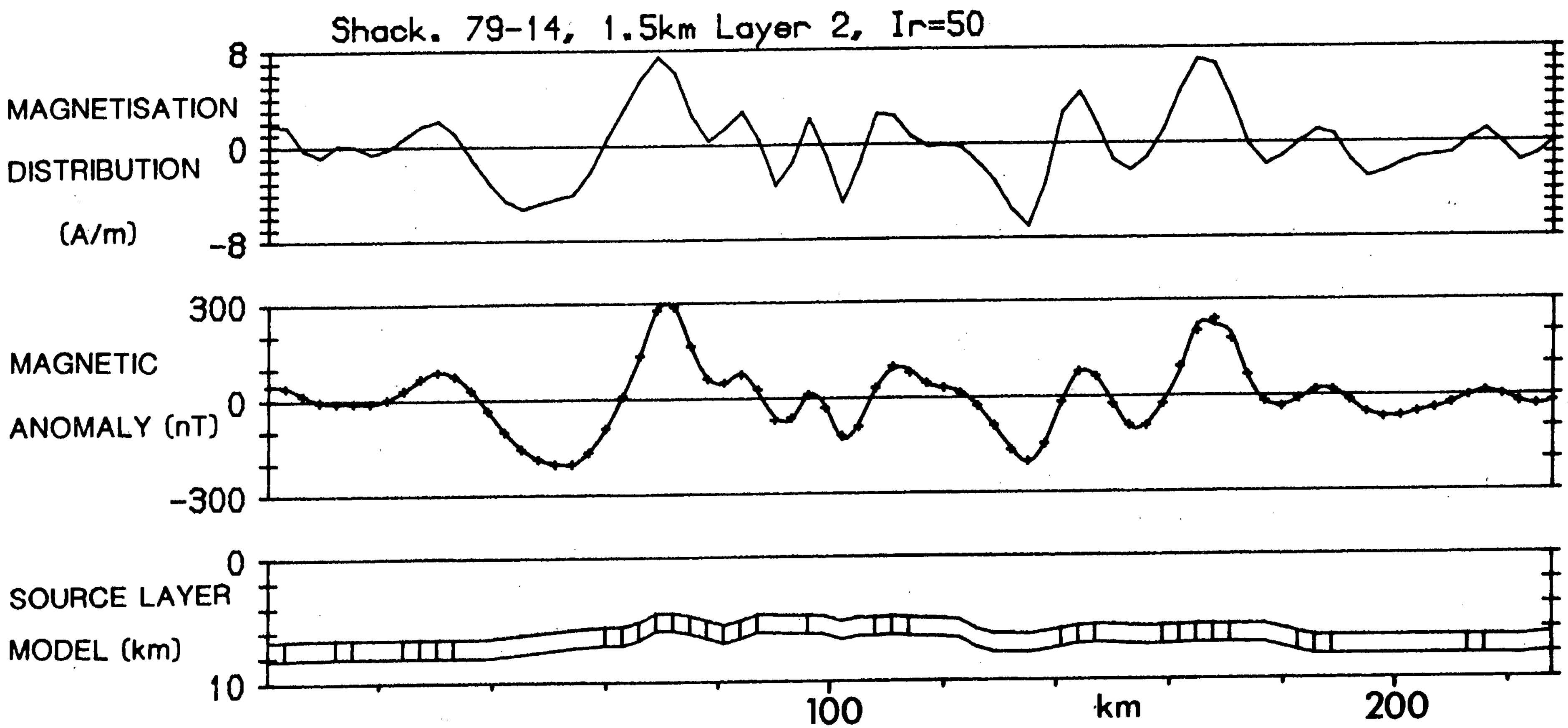
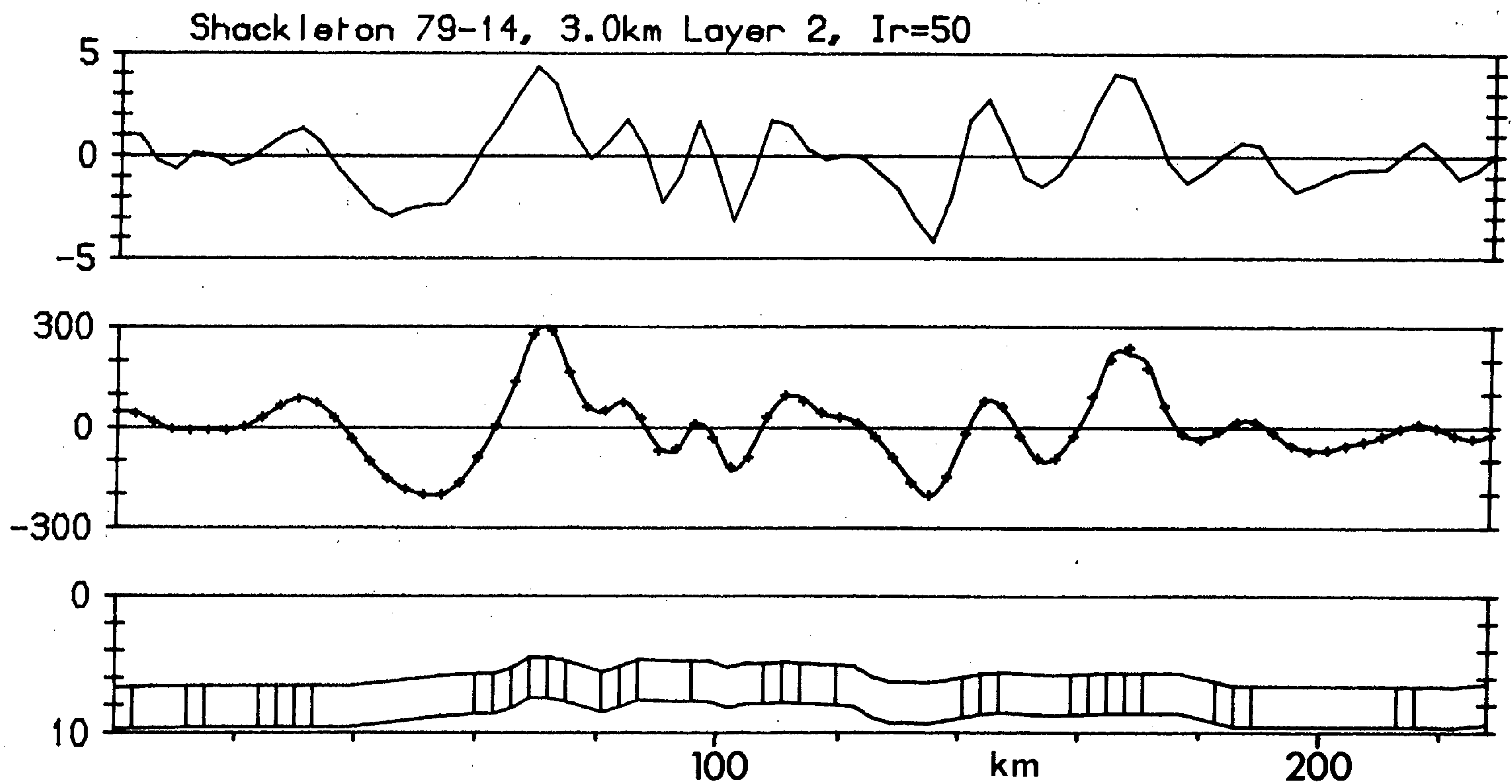


Figure 7.4 Inverse magnetic model along profile S 79-14. See Fig. 7.1 for location of track. Lower panel: 1.5 km thick magnetic source layer; vertical barring indicates normal polarity blocks. Centre panel: solid line = observed magnetic anomaly; crosses = anomaly recomputed from magnetic distribution in upper panel. Remanant magnetic inclination = 50, declination = 0.

Filter values (km^{-1}): F1 = 0.008, F2 = 0.015, F3 = 0.1, F4 = 0.125. Projected bearing = 103 deg.

Figure 7.5 Inverse magnetic model along profile S 79-14 with 3.0 km thick magnetic source layer. Otherwise as for Figure 7.4.



contemporary latitude of roughly 20°N, whereas the palaeogeographic maps of Smith et al. (1981) show the Rockall area as having drifted north to about 40°N by the Early Cretaceous.

In the example of the combined Challenger profile with a 1.5 km Layer 2 (Fig. 7.6) the calculated magnetic intensities of up to 12 A/m are unreasonably high for oceanic crust of presumed middle to upper Cretaceous age. Even the values of ±8 A/m computed for the S 79-14 model (Fig. 7.4) seem to be rather too high when compared with the ranges quoted by Harrison (1981). Yet Roberts et al. (1981) modelled a 500 m thick source layer containing crustal strips with magnetic intensities of up to 20 A/m along three profiles in proximity to the S 79-14 track of this study (Fig. 7.1).

Detailed studies of the magnetism of oceanic crust in the vicinity of the mid-ocean ridges have shown that magnetisation values in excess of 30 A/m can pertain in zero-age oceanic crust (e.g. Talwani et al. 1971). However, it has also been established that the magnetisation decreases rapidly over a period of several million years after initial accretion at a median rift. Away from the median ridge crests intensities of magnetisation generally lie between 2 and 6 A/m, normalized to the equator (Harrison 1981), a marked reduction that is commonly attributed to submarine alteration of the shallow volcanic Layer 2 (Cande and Kent 1976; Lowrie 1977; Kent et al. 1978). Hence the magnetic intensities of between 5 and 7 A/m (2.9 to 4.1 A/m normalized to the equator) computed for a 3.0 km thick source layer along both modelled profiles (Figs 7.5 and 7.7) are more comparable to the measured values reported in the literature.

Unfortunately problems then arise in attempting to explain the unusual thickness of the source layer. Although the source of oceanic magnetic lineations is most often quoted as being confined to the top 500 m or so of Layer 2 basalts, there are some workers who believe that a significant contribution to these characteristic anomalies is made from the underlying gabbroic Layer 3 (e.g. Kent et al. 1978; Harrison 1981). Unhappily drilling evidence from a number of DSDP cruises in the deep oceans has failed to clarify the current controversy. Instead the drilling data have revealed lateral and vertical changes in magnetisation that are considerably more complicated than was realised beforehand (Harrison, op. cit.). Such

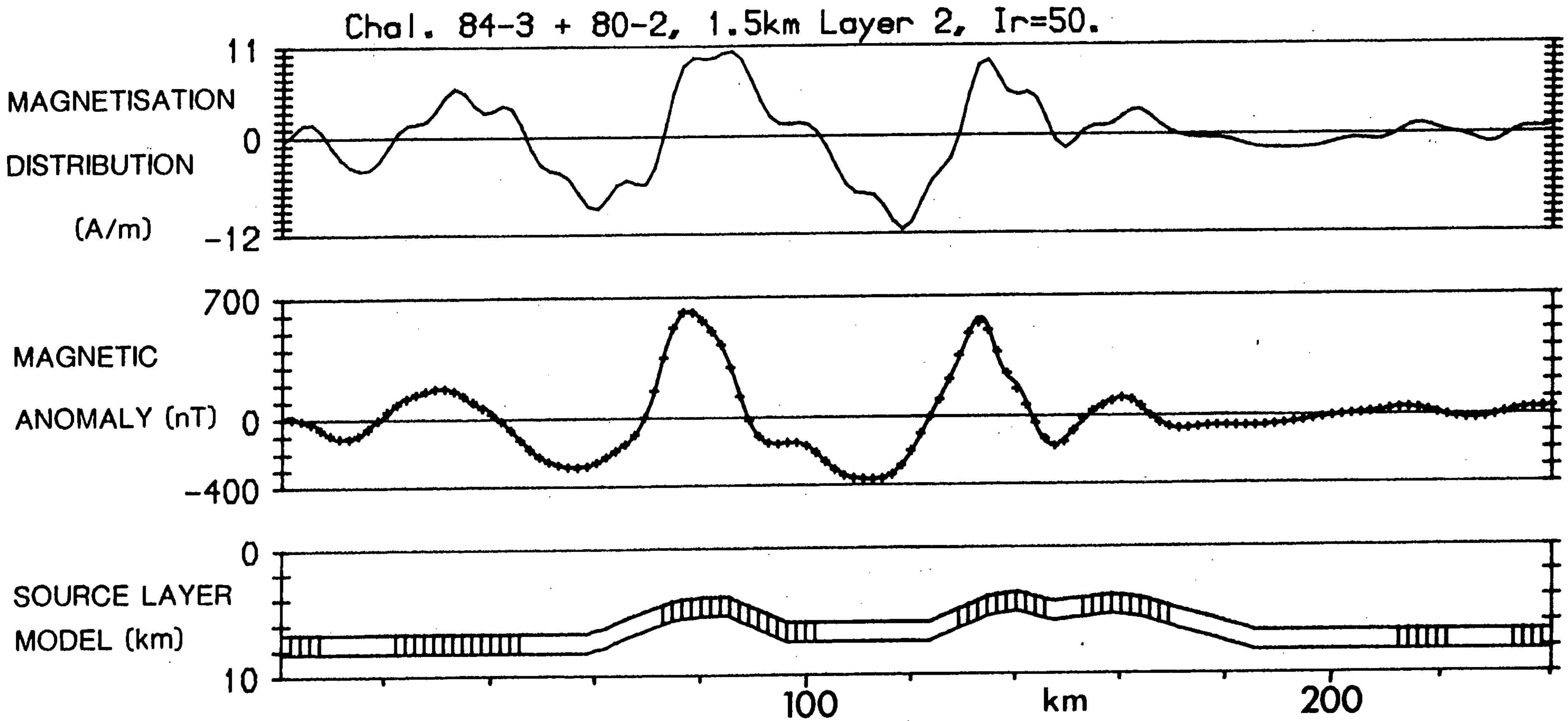


Figure 7.6 Inverse magnetic model along profiles C 84-3 and C 80-2. Refer to Fig. 7.1 for location and text for discussion. Ornament, scales, etc. as for Fig. 7.4.

Filter values (km^{-1}): F1 = 0.008, F2 = 0.015, F3 = 0.1, F4 = 0.133. Projected onto bearing 55 deg.

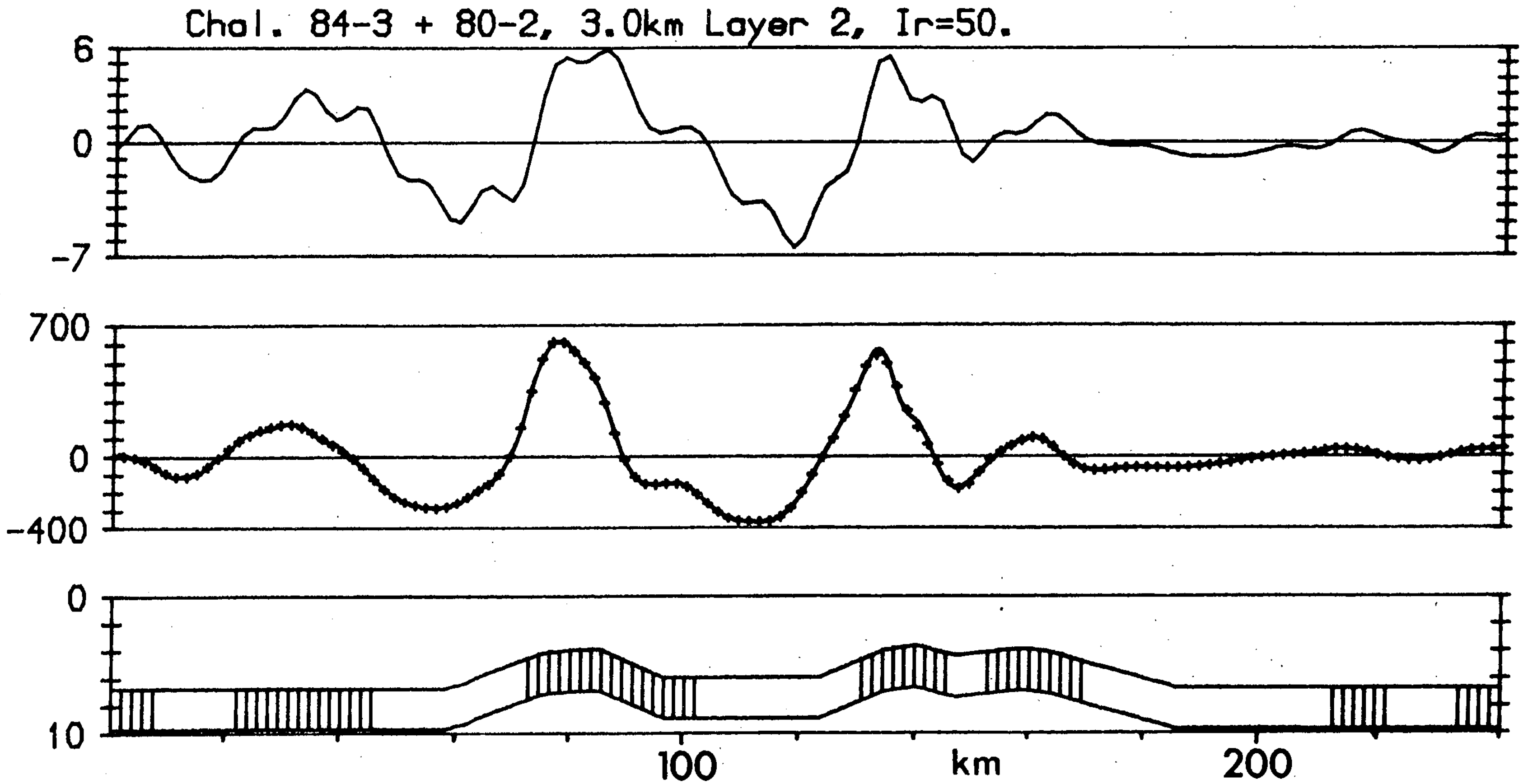


Figure 7.7 Inverse magnetic model along profiles C 84-3 and C 80-2 with 3.0 km thick magnetic source layer. Ornament, etc. as for Fig. 7.4.

revelations, therefore, lead one to be somewhat sceptical as to the geological realism or applicability of the simple models presented here (Figs 7.4 to 7.8).

If the basement modelled in Rockall Trough is truly oceanic then a source layer at least 3 km thick appears to be required below the Barra magnetic anomalies. Even were the remanant inclination to be increased to 60° or more, a source layer just 500 m thick would still demand intensities of magnetisation implausibly high, given the measured ranges of dredged and cored oceanic samples and the comparative antiquity of the Rockall Trough basement - as evidenced by the thick sedimentary cover.

Moreover, while there may be some weak elements of symmetry present in the Barra anomalies (Chart 3; Fig. 7.1) this is not manifested in the alternation of normal and reversed polarity blocks in the magnetic source layer of the models (panel C, Figs 7.4 to 7.7), at least not in the manner in which they are frequently and unambiguously reported from beneath the large ocean basins. In fact the inverse magnetic models succeed in confirming what was made evident from the earlier forward models; that the volcanic ridges have a high normal remanant magnetisation and are associated, accordingly, with one of the overlying Barra anomalies, the possible exception being the magnetic high at the south-western end of the C 84-3/C 80-2 model (Fig. 7.7).

This obvious correspondence between the strong relief of the volcanic basement ridges or oval domes, normal magnetisation, and a large magnetic anomaly is best illustrated in the combined Challenger profile models (Figs 7.6 and 7.7), and is emphasised in the final model (Fig. 7.8). Here all the parameters were kept the same as before, except the average intensity of magnetisation which was requested in the computer program to be 5 A/m as opposed to 0 A/m. The whole of the source layer is now normally magnetised (disregarding the 7 km wide zone of reversed polarity near the centre of the model) and the Barra anomalies are emulated by juxtaposing zones of high and low magnetisation to produce the peaks and troughs, respectively. The only difference with the forward models is that in those the large amplitudes and gradients were induced by establishing a sharp magnetisation contrast at depth in the crust.

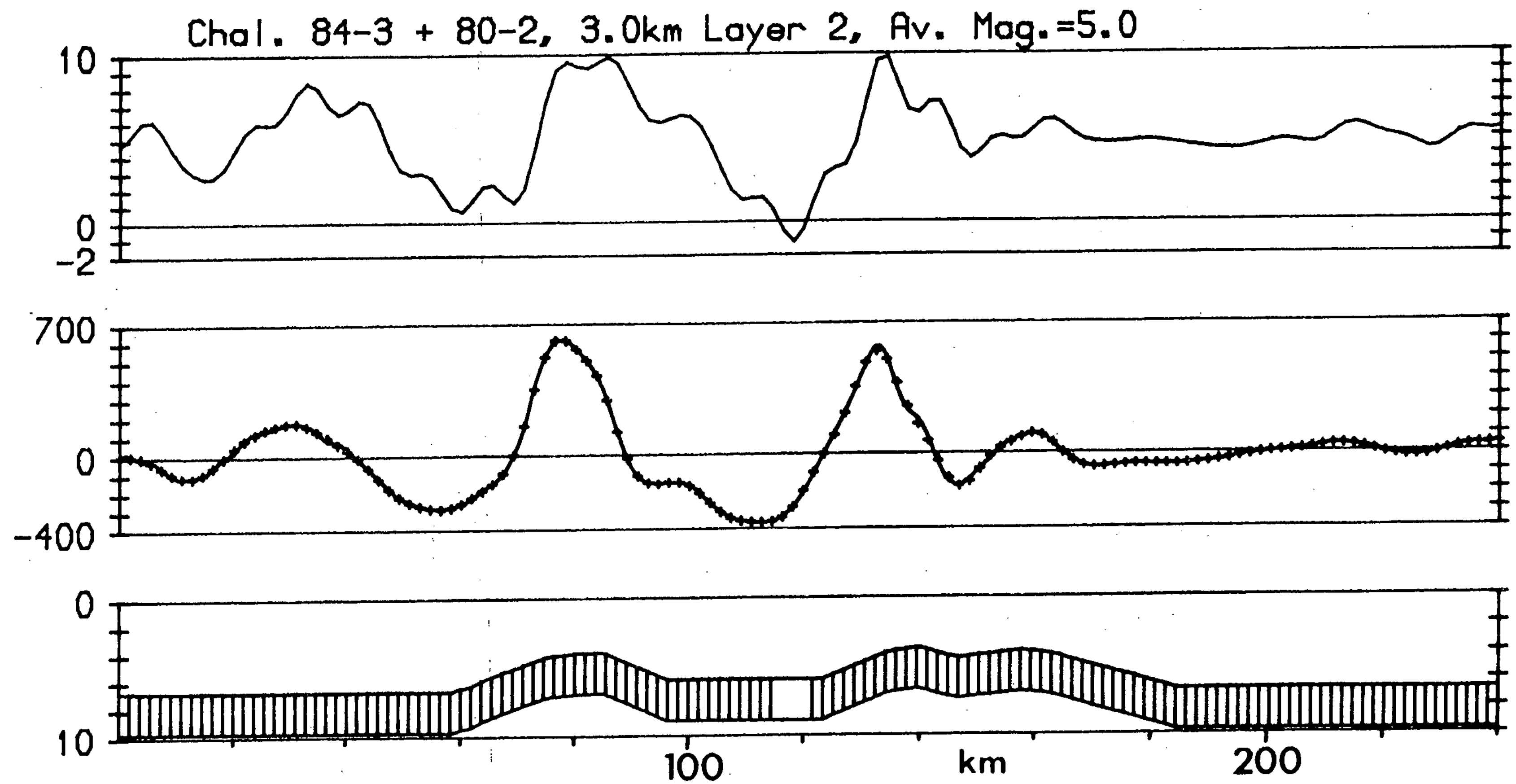


Figure 7.8 Inverse magnetic model along profiles C 84-3 and C 80-2 with 3.0 km thick source layer required to have an average magnetisation of 5.0 A/m. Here the proposed oceanic layer 2 is predominantly normally magnetised and the large observed magnetic anomalies result from high magnetisations associated with regions of elevated crust.

In conclusion then the inverse magnetic modelling appears to agree well with the prior forward modelling but does not support a number of expositions favouring an oceanic magnetic reversal origin for the arcuate Barra anomalies. This does not, however, preclude the formation of Rockall Trough by sea floor spreading. Further research into the geometry, nature, and subsequent history of the magnetic source (or sources) in oceanic crust is necessary before ambiguous regions like southern Rockall Trough can be addressed in greater detail.

8. DISCUSSION: THE NATURE, ORIGIN AND EVOLUTION OF ROCKALL TROUGH

8.1 Prologue

The first seven chapters of this study have dealt in some detail with the known geology of southern and central Rockall Trough and its surrounding regions, with the extent of geophysical coverage thereabouts, and most importantly with the presentation, interpretation and implications of a sizeable portion of that geophysical information. From all of this it is seen that the Rockall Trough is a more or less linear, aseismic bathymetric deep; one which is small as far as the larger oceans are concerned, but nonetheless is one of the largest sedimentary basins on the present Northwest European shelf and margin.

It is the only significant basin at bathyal depths in the region around the British Isles and as such favours comparisons to the Bay of Biscay, Iceland Basin, or Labrador Sea, rather than to similar-sized basins on the U.K. Continental Shelf - the North Sea for instance. The deep water in the Trough and its steep bounding margins serve to isolate the Rockall Plateau-Faeroe Platform micro-continent from the nearby continental shelf west of Scotland and Ireland. It is the considerable depths of and the thick sedimentary infill to the Trough that have so far hindered useful geological sampling or unambiguous remote sensing.

Consequently the age, origin and subsequent evolution of Rockall Trough have remained largely controversial and subject to speculation. In particular a debate persists regarding the possible continental or oceanic composition of the deep Rockall Trough, or whether there is present instead some form of unusual transitional crustal type. The absence of evidence from deep drilling information is probably responsible for prolonging this important argument, though it is possible that even were a drilling programme to be executed, given the current coring and penetration limitations at sea, it still may not resolve this long drawn out problem.

In addition to the main oceanic versus continental crust conundrum there are a number of related questions or ambiguities. What was the manner and age of rifting across the Rockall Trough? If sea floor spreading was responsible for creating the Trough what age is it? What was the direction of spreading and orientation of the spreading ridges? Is the resultant oceanic crust of the classic North Atlantic type or did diffuse accretion, perhaps combined with rapid sedimentation, give rise to a thicker crust with an anomalous structure? Are the rifting and spreading phases manifested in the sedimentary strata overlying the Trough and its margins?

The following discussion attempts not to give dogmatic answers to these questions, and other minor ones, but rather tries to present the various arguments and evidence for and against each, and from these suggests the most likely solution or solutions.

8.2 Continent or Ocean?

In 1965 Bullard, Everett and Smith presented their reconstructions of the North Atlantic continental jigsaw, in which they presumed the Rockall Trough to be floored by oceanic crust and accordingly rotated the Rockall Plateau to the south to accommodate the gap between North America, Greenland and Europe. Thus, essentially, started the debate as to the origin and composition of Rockall Trough. But twenty years later the Earth Science fraternity appears to be not much nearer to providing a correct solution and evidence in support. Nevertheless, what information has been collected and collated in the Rockall region over the past two decades finds a broad measure of support for the conclusions of Bullard et al. (1965).

That there is unusually thin crust beneath Rockall Trough has been repeatedly proven from the available gravity anomaly and seismic refraction data, as described earlier in Chapters 5 and 6, and is no longer under question. In the southern and central parts of Rockall Trough the gravity anomaly modelling of Scrutton (1972) and seismic refraction interpretations of Ewing and Ewing (1959) and Jones et al. (1970; reinterpretation of E10, see Fig. 5.14) are in broad agreement in predicting a crustal thickness of about 7 km and an M-discontinuity at roughly 14 km depth. These observations are further supported by the two-dimensional gravity modelling of this research (e.g. Figs 6.2, 6.3, 6.6 and 6.8) which appear to indicate fairly constant crustal thicknesses right up to the base of the continental slope, across which there is a rapid transition - 50 km or less - to normal continent crustal thicknesses (28-30 km).

This geometry of abrupt transition is, in this author's opinion, an excellent candidate for a classic continent-ocean boundary. It seems unlikely that continental crust could be so highly thinned over such a short distance and thereafter assume a comparatively simple, uniform geometry and thickness across the whole Trough. From surveying and drilling over the Bay of Biscay margins it is known that continental crust can be distended to as little as 5-6 km, i.e. oceanic crustal thicknesses (Montadert, Roberts et al. 1979). But here there is clearly no sudden boundary between normal continental and thinner oceanic (or continental?) crust, merely a smooth, gradual transition between the two types.

The compressional wave velocities observed at the seismic refraction stations within the Trough (Fig. 1.6) have been variously interpreted as being due to either an oceanic or continental structure. The shallow basement refraction velocities of between 4.72 and 5.12 km/s recorded by Hill (1952), Ewing and Ewing (1959) and Scrutton (1972) (refer to Figs 5.12 and 5.14) almost certainly correspond to a layer of basalt lavas, or at least extrusive rocks of similar composition. This layer may be up to 4.0 km thick according to the gravity modelling in southern Rockall Trough (Fig. 6.3). Underlying main crustal velocities of between 6.22 and 6.96 km/s may relate to Layer 3 of oceanic crust or simply stretched continental crust. In the former instance the overlying basement velocities of around 5.0 km/s possibly indicate the presence of oceanic Layer 2, in which case a standard oceanic velocity structure exists by comparison with the crustal profiles of Bott (1982, Ch. 3). Such a normal ocean crustal structure is favoured by this author, since the deeper refraction velocities are possibly too high for what would otherwise be highly stretched and faulted continental crust, and because it is unnecessary to advocate the exception when the standard rule seems to be in operation. Why seek to explain the gravity models and refraction velocities in terms of an unusual continental structure when an oceanic accretionary origin is simpler and more obviously favoured?

Towards the north-eastern end of Rockall Trough a c. 300 km explosion refraction line shot along 58°N, just north of Anton Dohrn Seamount, was reported by Bott et al. (1979; Figs 1.6 and 8.1). Analysis of the refraction data indicated a thickness of crust beneath Rockall Trough of roughly 14 km, twice that predicted for the region covered by this author's study, and a depth to the Moho of approximately 18 km, about 4 km deeper than further south (Fig. 8.1). The rather uniform crust below Rockall Trough was found to have an average P-wave velocity of about 6.4 km/s, and a value of 8.2 km/s was calculated for the top of the underlying mantle. Bott et al. (1979) interpreted this velocity structure as representing anomalously thick ocean crust, though they recognised that a thin continental crustal origin was feasible, albeit unlikely. The same authors, in identifying the northward thickening of probable oceanic crust in Rockall Trough, provided a plausible explanation for the gradual north-easterly reduction in the height of the continental slope, from about 3 km below west Porcupine Bank to about 1 km near the Wyville-Thomson Ridge (Fig. 1.4) - although some part of this disparity is attributable to the greater thickness of sediments to the north-east. However, Bott et al. (1977) did not put forward an explanation as to why the crust should thicken so in this direction.

Two points are worth mentioning here as regards the free-air anomaly data and the model profiles. Firstly, the gravity crustal models across the central parts of Rockall Trough (Figs 6.6 and 6.8) indicate that the axial sections of the thin Rockall crust occur at a shallower level than the marginal zones. This broad central elevation, with a median rift or hollow in the example of the Charcot gravity model (Fig. 6.8) giving a rift shoulder appearance, is strongly reminiscent of the mid-ocean ridge topography. The analogy is strengthened by the apparent symmetry of the topography on the thin crust in Rockall. Bott et al. (1979) model a similar sort of pattern along their free-air anomaly profile trending NW-SE across northern-most Rockall Trough. This strongly favours the contention that oceanic crust persists along the whole length of Rockall Trough.

Secondly, the densities applied to the thin crust below the Trough in the 2D gravity models of Chapter 6 cannot be tightly constrained, at least not sufficiently to allow us to be definitive about the composition of the deep basin. In many of the models a crustal density of roughly 2.80 g/cm^3 was used which can be interpr-

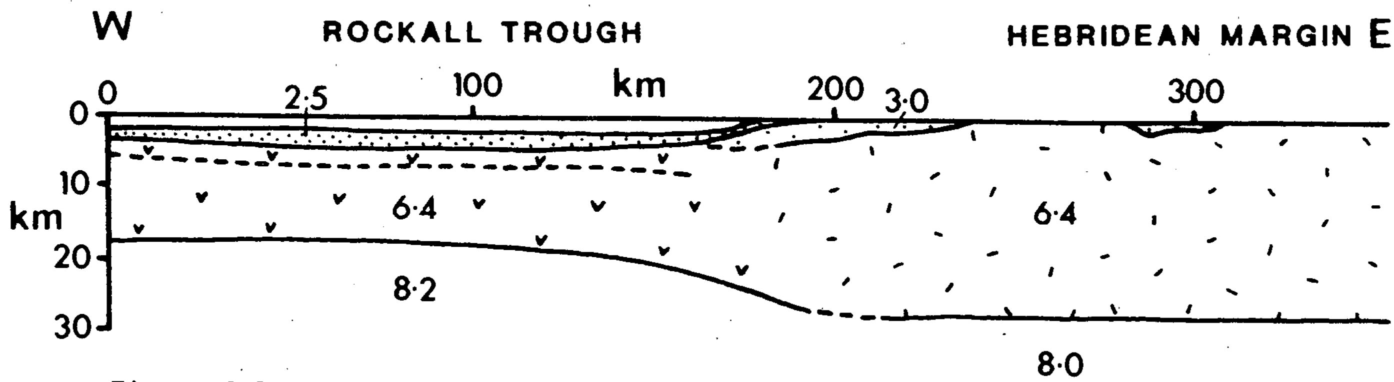


Figure 8.1 Explosion refraction line in northern Rockall Trough observed by Bott et al. (1979). Velocities in km/s. V symbol = oceanic crust. Dash symbol is continental crust. Heavy stipple = younger post-rift sediments in Trough; Light stipple = Mesozoic and older pre-rift sedimentary basins below the Hebridean margin.

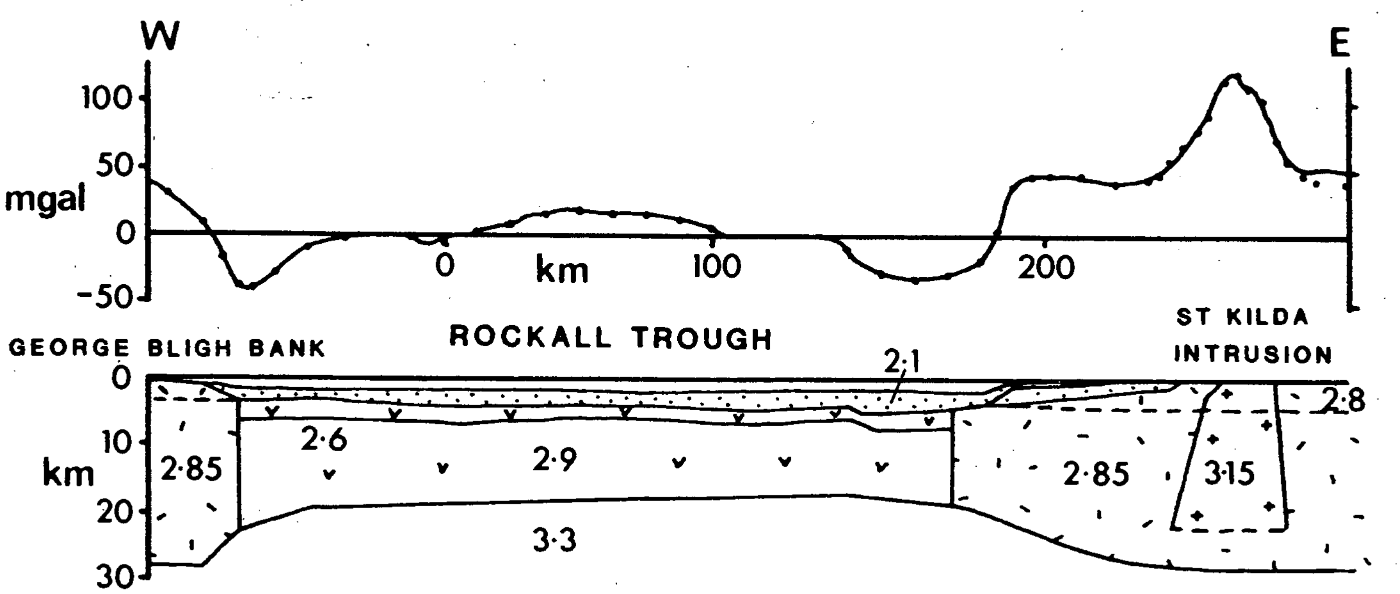


Figure 8.2 Gravity crustal model across northern Rockall Trough, after Bott et al. (1979). Density values shown in g/cm^3. Symbols as for Fig. 8.1.

eted as representing continental or oceanic material. However, in each of the models it is possible to increase the main crustal density to 2.90 g/cm^3 (Layer 3) while introducing some thickness - c. 1 to 2 km - of lighter Layer 2 volcanic rocks of density 2.5-2.7 g/cm^3, hence still maintaining a close fit to the observed anomaly. Such an oceanic density structure was modelled successfully in Figure 6.6c and also by Bott et al. (1979; Fig. 8.1) to the north-east.

Along their main refraction line K Bott et al. (1979) calculated a P-wave velocity of 8.2 km/s for the topmost mantle beneath Rockall Trough but a more usual velocity of 8.0 km/s below the c. 27 km thick Hebridean continental margin (Fig. 8.1). The distinctly higher velocity beneath the deep Rockall Trough may be further evidence, albeit limited, that the crust here is of oceanic composition, since ithas been observed that the mantle beneath the ocean basins exhibits velocity anisotropy (Hess 1964; Bott 1982). It has been shown that the mean sub-Moho velocity is about 8.0 km/s parallel to the ridge crests and 8.3 km/s parallel to the spreading direction. The velocity of 8.2 km/s found by Bott et al. (1979), although not well controlled, is compatible with such a seismic anisotropy since the trend of the refraction line is close to the probable spreading direction in Rockall Trough. Moreover, the uppermost mantle velocity recorded at the Trough-parallel refraction station E10 (Fig. 1.6), as reinterpreted by Jones et al. (1970; Fig. 5.14), is the normal 8.0 km/s. Recent two-ship seismology in the Rockall Trough (White et al. 1985) is consistent with both the oceanic velocity structure of the thin crust below the Trough and also the upper mantle seismic anisotropy just described.

As a last word on the seismic refraction data, Scrutton (1972) recorded a weak refractor with a velocity of 3.89 km/s at a depth of roughly 9 km along his station C in the Rockall Trough (Fig. 1.6). A small number of workers (e.g. D.K.Smythe, R.A.Scrutton, pers. comm.) have suggested that this event may mark a converted S-wave arrival from Layer 3 gabbros. In the absence of any further supportive data this is difficult to confirm; the S velocity is possibly too low for oceanic Layer 3 which normally renders values in the range 4.3 to 4.5 km/s according to Harrison and Bonatti (1981).

Moving on now to consider the seismic reflection profiling data in the context of the continent versus ocean debate, it was shown in Chapters 4 and 5 that beneath Rockall Trough a highly reflective, irregular seismic basement is extensively developed which which does not favour the transmission of coherent reflections from the deeper geological basement - continental or oceanic crust. The reflection profiles available to this study indicate that this distinctive seismic basement persists right across and along the Trough, from the base of one margin to the other. Its strong reflectivity, gently undulating but irregular surface relief, and characteristic variable, impersistent layering can best be interpreted as thickly developed, overlapping lava flows.

It is the view of this author that such deep, layered seismic basement represents the top of Layer 2 of oceanic crust and that the arcuate volcanic piles constituting the Barra volcanic ridge system are effectively lateral equivalents of the same material, though differing in their anomalously thick and shallow development. If this oceanic origin is correct and the crust here is the same age (middle Cretaceous) as that west of Goban Spur and southern Porcupine Bank, it seems curious that the relief on the basement below Rockall Trough does not exhibit the classic Layer 2 topography displayed in the latter areas (profile CM-11 of Graciansky, Poag et al. 1985). Smythe (pers. comm.) agrees independently that the layered seismic basement in Rockall Trough represents the top of old oceanic crust. He has migrated and depth-converted good quality sections of multichannel seismic profiles in central Rockall Trough and from these interpreted scarps in the oceanic volcanic basement which face inwards towards the axis of the basin, in the same fashion as seen in younger crust in the major oceans. Similar small inward-facing scarps can be picked, albeit weakly, on a few of the double-channel profiles making up this study; for example tracks S 79-14 (Fig. 5.3), C 80-1 (Fig. 5.4) and C 84-1. Nevertheless the physical differences between the presumed oceanic crust north of the Charlie-Gibbs F.Z.-Clare Lineament and known oceanic crust to the south of that line require some sort of explanation.

Roberts et al. (1981) recognised the rather unusual appearance of the postulated oceanic crust in the Trough, crust which they assigned the same age as that off Goban Spur, and proposed its formation in a tectonic environment similar to that currently operating in the Gulf of California; that is, shallow accretion by intrusion and extrusion into a rapidly sedimented basin. This remains an attractive and plausible explanation for the rather diffuse appearance of the basement on the seismic profiles, caused by the interlayering of lava flows, sills and sediments; but it is essential, at the same time, to try and understand why the oceanic crust on the southern side of the CGFZ-Clare Lineament is typical and has seemingly undergone a normal evolution. The obvious solution is to infer that the basement in Rockall Trough is older and/or was formed under different conditions.

One possible interpretation of the seismic basement, not previously put forward, is that it represents a superficial volcanic layer covering and obscuring deeper, highly stretched and thinned continental crust. Megson (pers. comm.), on the basis of commercial multichannel seismic profiles, recognises rotated fault blocks of presumed continental material near the base of the Trough margins. These, she believes, persist across the basin such that it is floored wholly by distended continental crust, an option that was favoured by Talwani and Eldholm (1972). The simple seismic profiles of this work and also the migrated, depth-converted profiles of Smythe (pers. comm.) similarly show clear down-faulted and rotated blocks beneath the continental rise (e.g. Figs 5.3, 5.4 and 5.15): indeed those of Smythe seem to indicate that some of these tilted blocks are faulted in an antithetic sense, i.e. they downthrow towards the continent. But these features are almost certainly a normal consequence of the formation and subsidence of passive continental margins; there is no convincing evidence for these or any other continental structures within the Trough proper.

There are a number of other aspects of the seismic reflection data which again favour the Rockall Trough being composed of oceanic rather than continental crust. The relief on seismic basement has a grain that often trends parallel or sub-parallel to the margins of the Trough, at least away from the BVRS, as shown by the seismic basement chart (Chart 4, back pocket). It is difficult to envisage this sometimes symmetrical pattern developing within strongly rifted continental crust. There is little evidence however for any transform or fracture zones that could be used to define the opening direction of Rockall Trough. This apparent absence must partly reflect the erratic coverage and occasionally poor quality of the available seismic data. If Rockall Trough is indeed the product of sea floor spreading then the curvature of its two margins alone demands that transform faults should be present, if one adheres to rigid plate tectonic theory.

The region between north Porcupine Bank and Rockall Bank is the obvious place to suspect the presence of transform features, but no evidence was found in the Trough in support of the Porcupine Fracture Zone proposed by Roberts (1975) or those predicted by Roberts et al. (1981; and Fig. 8.3). At the margins, however, a number of transfer structures mapped below the continental slope (Fig. 5.2) are possibly markers indicating the hidden presence of fracture zones. The abrupt discontinuity marking the northern edge of the North Porcupine Salient (Chart 4 and Fig. 5.22) is probably a good example of a continent-ocean fracture zone, a view inferred previously by Roberts et al. (1981) and communicated by D.G. Masson (pers. comm.). That such a marked feature is not picked up in the Trough as an inactive transform fault and/or fracture zone may be due to the poor resolution of many of the reflection profiles. Alternatively the diffuse style of oceanic accretion discussed above may have been such as to preclude the development of well-defined orthogonal median ridges and fracture zones, as seems to be the situation in the present Gulf of California (Curray, Moore et al. 1982).

As with the gravity and seismic data the magnetic anomaly information is liable to conflicting interpretations. Within Rockall Trough the Barra magnetic anomalies and the weak lineations further to the north-east (Chart 3; Fig. 7.1) have usually been interpreted as ocean magnetic reversals of mid-Cretaceous age (Vogt and Avery 1974; Roberts 1975; Srivastava 1978; Kristoffersen 1978; and Roberts et al. 1981; Fig. 8.3). The magnetic inversion models presented in Chapter 7 show, fairly convincingly, that the large amplitude, arcuate Barra anomalies in southern Rockall Trough are unlikely to be caused by a <u>typical</u> oceanic reversal and crustal structure (Figs 7.4 to 7.7). Rather they can be accounted for by invoking appropriate surface and basal topography on the shallow volcanic basement, together with the inclusion of underlying zones of intrusion. The weaker magnetic lineations elsewhere in Rockall Trough also appear to correlate well with ridges in the subjacent extrusive layer, and it seems unnecessary to advocate a magnetic reversal origin here either.

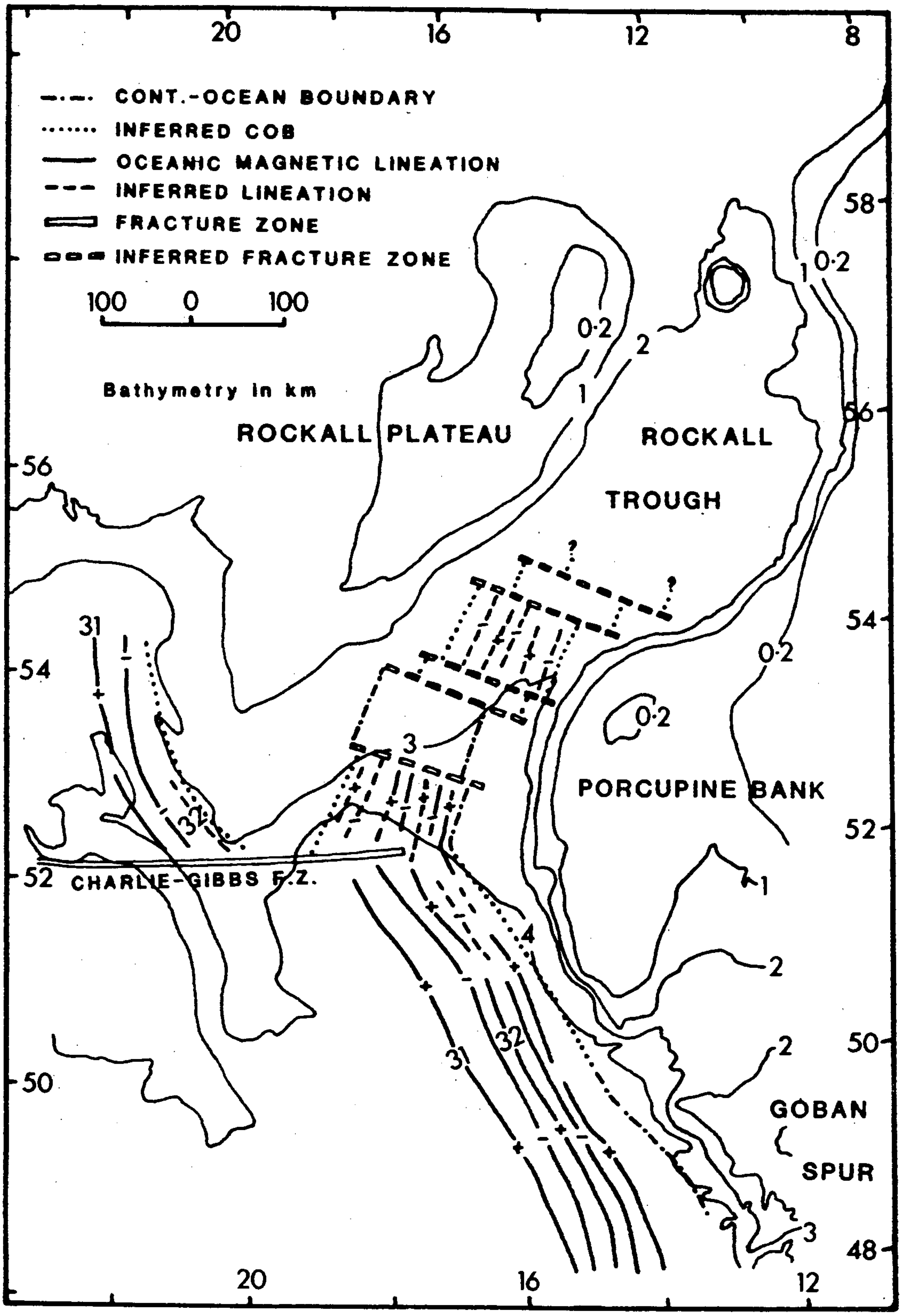


Figure 8.3 The oceanic crustal model of Roberts et al. (1981) in Rockall Trough and off Goban Spur. Magnetic anomalies 31 and 32 are shown. Polarities indicated by + and - signs. Anomalies 31 and 32 were also speculated to be 33 and 34, respectively.

How then does the presence of the conspicuous Barra magnetic anomalies, yet the absence of a clearly developed pattern of reversal lineations throughout the Trough, fit in with this author's contention, based on the gravity and seismic data, that the geological basement here comprises true oceanic crust? The quiet magnetic signature over much of Rockall Trough, though specially adjacent to its margins, is generally attributed to the accretion of ocean crust here during the long interval of normal magnetic polarity in the middle Cretaceous (the K-N polarity period, 118-83 m.y.B.P., of Harland et al. 1982) . This interpretation is dependent to some extent on both the identification of Anomaly 34 - marking the young end of K-N - intersecting the eastern end of the Charlie-Gibbs Fracture Zone, and the belief that the oceanic crust between this anomaly and Goban Spur-Porcupine Bank continues northwards into Rockall Trough (Figs 4.12 and 8.3; Chart 3), as discussed by Roberts et al. (1981).

Although there now appears to be general agreement that the oldest recognisable magnetic lineation here is Anomaly 34 (Kristoffersen 1978; Masson et al. 1985; Scrutton 1985), and not A32 (e.g. Srivastava 1978), there is considerable doubt - as testified by this work - as to the geophysical and geological continuity across the CGFZ-Clare Lineament. Despite this element of ambiguity the argument for the origin of magnetic quiet zones is still applicable since the next oldest reversal (M0 of the M sequence) does not appear until roughly 120 m.y. B.P. (magnetostratigraphy from Harland et al. 1982), some 40 m.y. before the young end of K-N. This is ample time for the narrow Rockall Trough to have opened by sea floor spreading before and independently of the oceanic crust to the south, yet still have accreted during the K-N normal polarity phase.

There are a number of other processes which could each separately provide an explanation for the observed magnetic quiet zones, or which could do so in some combination with the intra-K-N argument described above. In the first instance, if the crust beneath Rockall Trough was accreted towards the beginning of K-N or even earlier then one would not expect to observe a well defined pattern of reversals; this is because it has been demonstrated that the magnetic intensity of ocean crust decreases dramatically with age as a result of submarine alteration, metamorphism and so on.

Secondly, the considerable depth to the oceanic basement and the great thickness of superincumbent sediments, two factors that are related to the antiquity of the Trough, also do not favour the identification of a typical oceanic magnetic pattern. The role of the sediments here is somewhat equivocal: some would argue that the overlying sediments have a negligible effect in attenuating any magnetic signature, others that it is possible for the sediments to effect some measure of thermal blanketing and hence weaker magnetisation of the underlying rocks.

Thirdly, the style of accretion may have been such as to prevent the proper imprinting of a magnetic reversal sequence. This author and others (e.g. Roberts et al. 1981; Price and Rattey 1984) have previously described how the seismic reflection profiles hereabouts tend to favour the accretion of Layer 2 in Rockall Trough in shallow water and in association with rapid clastic sedimentation. This could explain both the anomalous appearance of the seismic basement and the lack of coherent linear magnetic anomalies. Further to this Roots et al. (1985) developed an elegant model (Fig. 8.4) whereby the size of the initial ocean crustal blocks generated adjacent to oblique-to-spreading rifted margins (clearly the case for Rockall Trough) was too small to enable their magnetic signatures to be resolved above background noise (75nT). A normal oceanic reversal sequence was predicted to have formed only after about 100 km distance from the continental margin, by which time the ridge-transform system was able to expand and mature to assume a more usual orthogonal geometry (Fig. 8.4).

Such a model is attractive in explaining the quiet magnetic character of Rockall Trough, even in the absence of the other options mentioned above, because it appears that any accretion that has taken place has done so through small-scale, poorly developed ridge-transform systems that are oblique to the initial rift; there is no seismic evidence for a mature, orthogonal spreading geometry. In addition the width of oceanic crust in Rockall Trough of about 200 km predicted by the gravity models and seismic reflection data is in excellent agreement with the maximum extent of the quiet zone modelled by Roots et al. (1985) - 2 x 100 km.

In complete contrast Megson (1983) modelled the magnetic quiet zones with highly thinned continental crust having a low induced intensity of magnetisation (c. 1.6 A/m). The Barra anomalies were

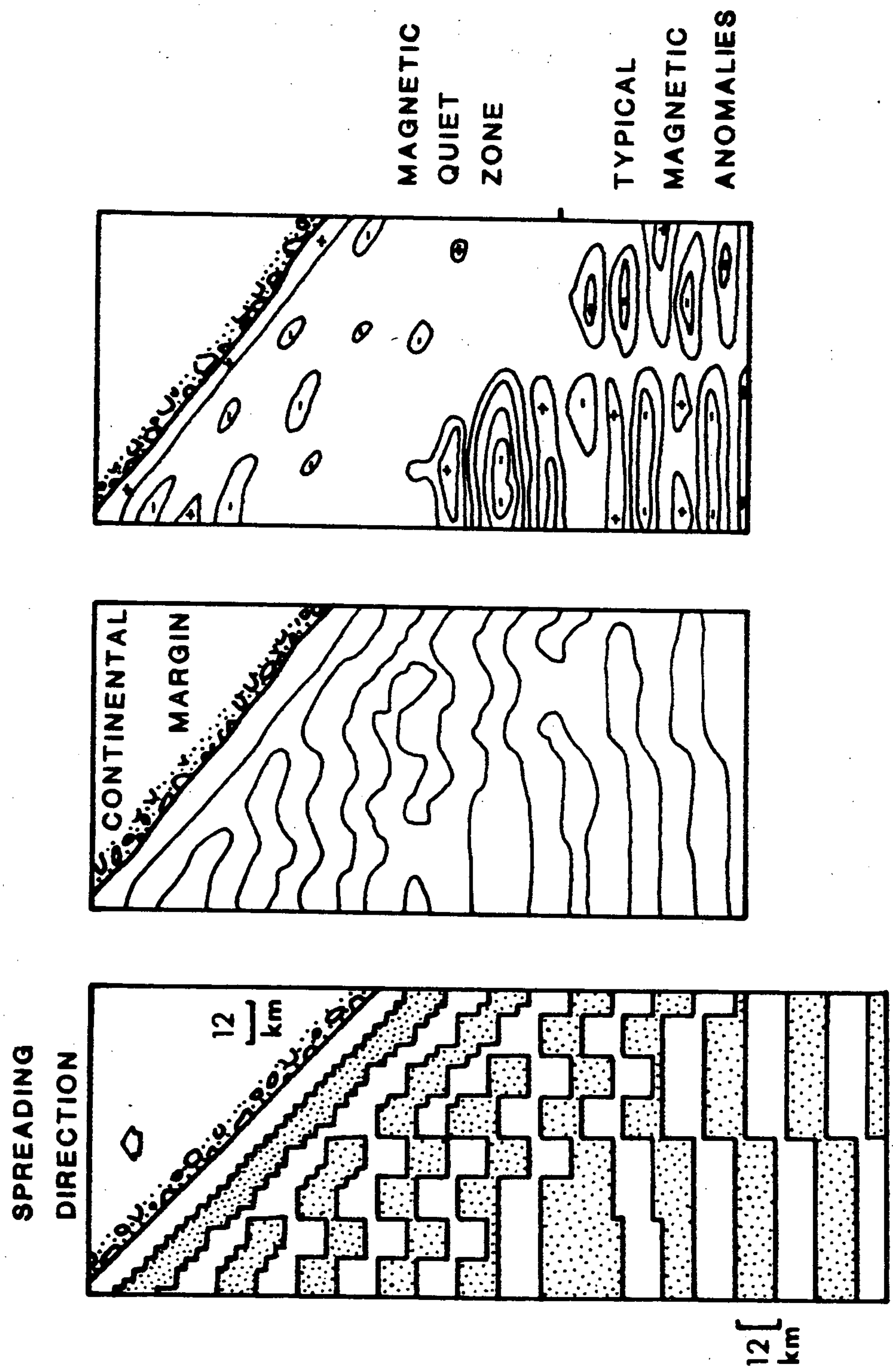


Figure 8.4 A model for the formation of magnetic quiet zones at rifted continental margins. Redrawn from Roots et al. (1985). Left panel: distribution of 1 km thick magnetic source blocks in oceanic crust (6km depth, 5.5 A/m magnetic intensity). Centre panel: Computed magnetic anomaly field over model. 0 nT contour only shown. Right panel: computed magnetic anomaly field contoured every 50 nT above and below $\pm$ 50 nT to highlight magnetic quiet zones as blank areas.

then modelled by introducing intrusions (3.5 A/m) into this stretched crust, in a rather similar fashion to the magnetic models of this study (Chapter 7). In the scheme of the present author the Barra magnetic anomalies are assumed to arise from the curved region of volcanic ridges and deep-seated intrusive zones; a phase of igneous activity that is possibly a diffuse northward extension of the normal Anomaly 34 oceanic crust in the south (Fig. 8.3). But this is difficult to reconcile with the earlier proposals that the remainder of Rockall Trough is also composed of normally magnetised A34 ocean crust, in which event no magnetisation contrast would be expected. Either the ages and magnetic polarities of the two crustal provinces are different, or they are the same but the style of accretion was such as to give the Barra igneous province a better defined magnetic signature. The second option is favoured here, though it should be remembered that the long span (c. 40 m.y.) of the A34 normal polarity phase still allows for possible large differences in the timing of the two igneous episodes.

The final considerations in this discussion of the composition of Rockall Trough are those of regional geometry and scale. The congruency in the outline of the opposing margins of the Trough is a reasonable indicator that sea floor spreading has been in operation. It is not difficult to imagine or eye in the rotation of the two margins together in the classic Bullard et al. (1965) manner. Matching the main inflections in the isobaths of the continental slopes suggests a spreading direction between east and south-east. The width of the Trough, the steepness and cross-sectional shape of its margins, the three or more axially situated seamounts in the Trough, and the high whole-crustal stretching (beta) factor of 4 to 4.5 implied by the gravity anomaly models: all these are in support of oceanic crust underlying Rockall Trough.

On a broader scale it has been established that a long, narrow ocean basin was probably in existence during the Early Cretaceous linking the opening Central Atlantic to the Norway-Faeroe region via the Rockall Trough (Fig. 8.5; Kristoffersen 1978; Srivastava 1978; Graciansky, Poag et al. 1985). The continental reconstruction of the North Atlantic to what is currently believed to be Anomaly 34 (83 m.y. B.P.; Fig. 8.5) is further convincing evidence for an oceanic origin. The manner in which the oceanway narrows towards the Faeroe-Shetland Channel and Norwegian margin is consistent with it

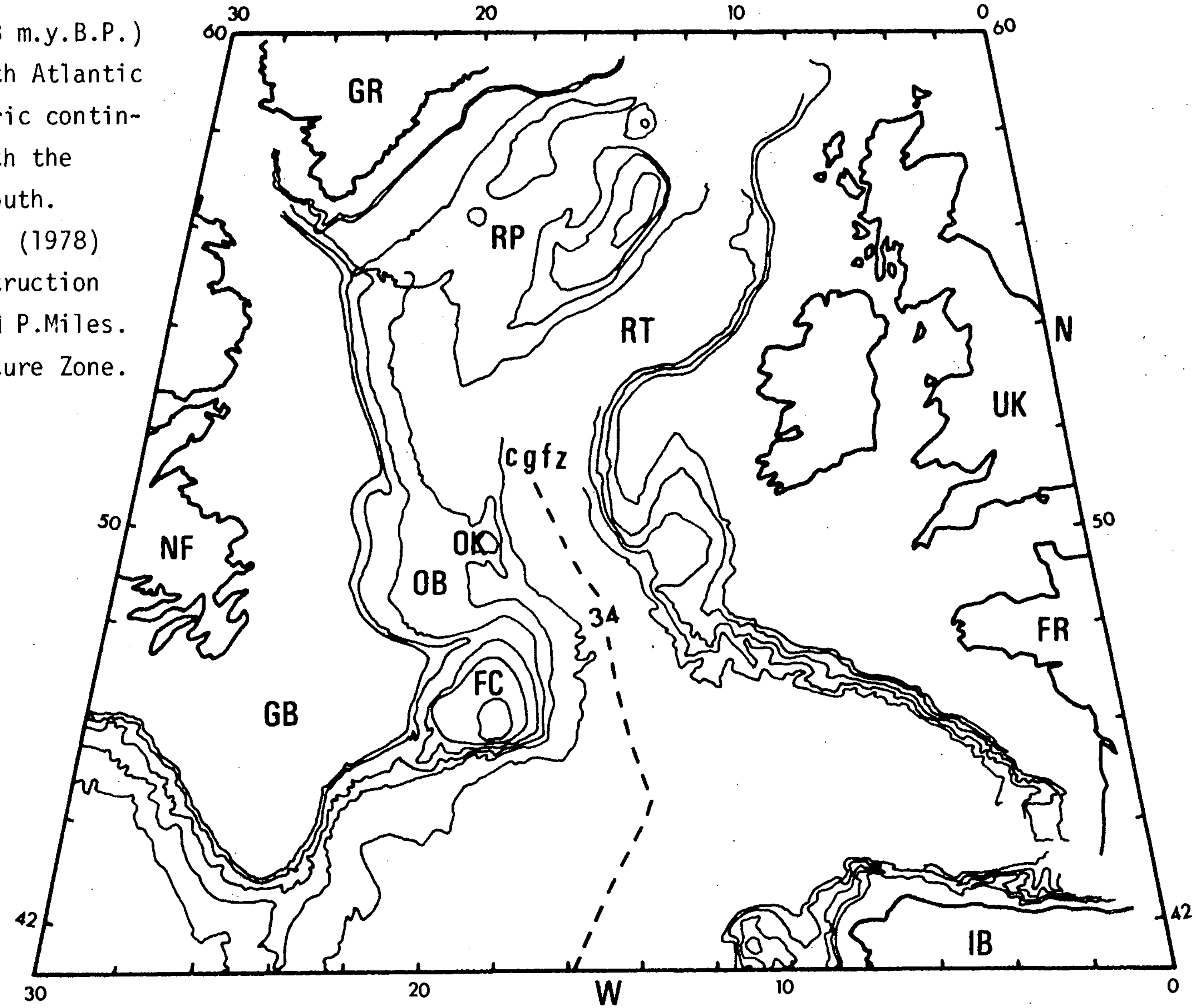


Figure 8.5 Anomaly 34 (83 m.y.B.P.) reconstruction of the North Atlantic illustrating the bathymetric continuity of Rockall Trough with the Central Atlantic to the south. Redrawn from Kristoffersen (1978) and an unpublished reconstruction courtesy of D.G.Masson and P.Miles. CGFZ = Charlie-Gibbs Fracture Zone.

marking a northward propogating rift from the already well developed Central Atlantic, the pole of rotation occupying a position in the Svalbard region (Hanisch 1984; Haszeldine 1984). No suitable analogies to the geometry and scale of this feature are obvious in the truly continental environments, except perhaps the East African rift system.

8.3 Palaeozoic or Mesozoic?

Although we are now fairly confident that Rockall Trough is floored for most of its width with oceanic crust there is still a great deal of uncertainty as to the age of the main phase or phases of continental rifting and oceanic accretion that structured it. At the mouth of Rockall Trough Anomalies 34 to 32 and younger of the North Atlantic Ocean terminate abruptly at the Charlie-Gibbs Fracture Zone. Since it is unreasonable, on geometric and sediment thickness grounds, to expect the Trough to have formed after these anomalies, a latest age for the completion of accretion in the Trough can be put at roughly 83 m.y. B.P. (Anomaly 34 time). Before that time the age of formation of the Rockall channel has been ascribed to various dates ranging right back to the Westphalian.

A late Early to Late Cretaceous age for the formation of the oceanic crust is popularly advocated on the basis of the Anomaly 34 reconstruction of the North Atlantic (Fig. 8.5; Kristoffersen 1978; Roberts et al. 1981; Hanisch 1984; Haszeldine 1984). This model relies on two assumptions: 1) that Anomaly 34 continues past the eastern termination of the Charlie-Gibbs F.Z. and forms the central (youngest) spreading lineation of both the Barra anomalies and the rest of Rockall Trough; 2) the oceanic crust between Anomaly 34 and Goban Spur-southern Porcupine Bank (83-100 m.y. B.P., Masson and Miles 1984; Masson et al. 1985), although gradually narrowing northwards, continues into Rockall Trough where an equal width of contemporary oceanic crust is mirrored on the western half of the Trough (Fig. 8.5). South of the CGFZ a similar half-width of oldest oceanic crust must then be present on the Canadian margin adjacent to Orphan Knoll.

If the Anomaly 34 identification here is correct, as is now widely accepted (Scrutton 1985; Masson et al. 1985; Srivastava 1985), then the Barra anomalies cannot represent sequential magnetic reversal lineations since there are no polarity changes within the 35 m.y. preceding A34. The magnetic charts (Chart 3, Fig. 8.6), however, indicate that the Barra anomalies may consist of four, possibly five, linears. Such a pattern could not be generated by normal sea floor spreading during the time period postulated, unless the area in question featured unusually high spreading rates. Furthermore the Barra anomalies, in this author's view, are not lineations but an arcuate zone of oval to circular positive anomalies (Fig. 8.6).

This study highlights a configuration, an argument, that is in strong opposition to the often cited middle Cretaceous age for Rockall Trough and its contemporaneity with the ocean crust west of the Goban Spur region. That argument is that there is very little geophysical and presumably, therefore, geological continuity across the major structural line marked by the CGFZ and Clare Lineament, such as could be invoked to link the two provinces. On the magnetic anomaly charts (Chart 3, back pocket) this is manifested in a distinct contrast in character north and south of the linear fracture zone. In the Porcupine Abyssal Plain there are clear sea floor spreading lineations dating back to A34, and a tendency towards lineations (induced by basement topography, Scrutton 1985) before that time west of Goban Spur (Roberts and Jones 1975; Graciansky, Poag et al. 1985). Even on the free-air anomaly chart (Chart 2, back pocket), which by its long wavelength content tends to reflect the deeper geology, the change from a mainly negative anomaly field in the south to a predominantly positive field in the Trough also suggests that the CGFZ-Clare Lineament divides provinces of different geology . The 2D gravity and magnetic models provide further support for this conclusion (e.g. Figs 4.10 to 4.14).

But it is the seismic reflection profiles that constitute the most convincing evidence for the lack of continuity, in fact the marked change in seismic geology, between the oceanic Porcupine Abyssal Plain and the enigmatic Rockall Trough. In earlier chapters it was noted that south of the CGFZ the top of oceanic basement exhibits an archetypal volcanic Layer 2 with numerous small scarps facing toward the mid-Atlantic ridge (Figs 4.4c and 4.5). In Rockall Trough, in contrast, the deep layered volcanic basement and that constituting the BVRS is less sharply defined and does not have a

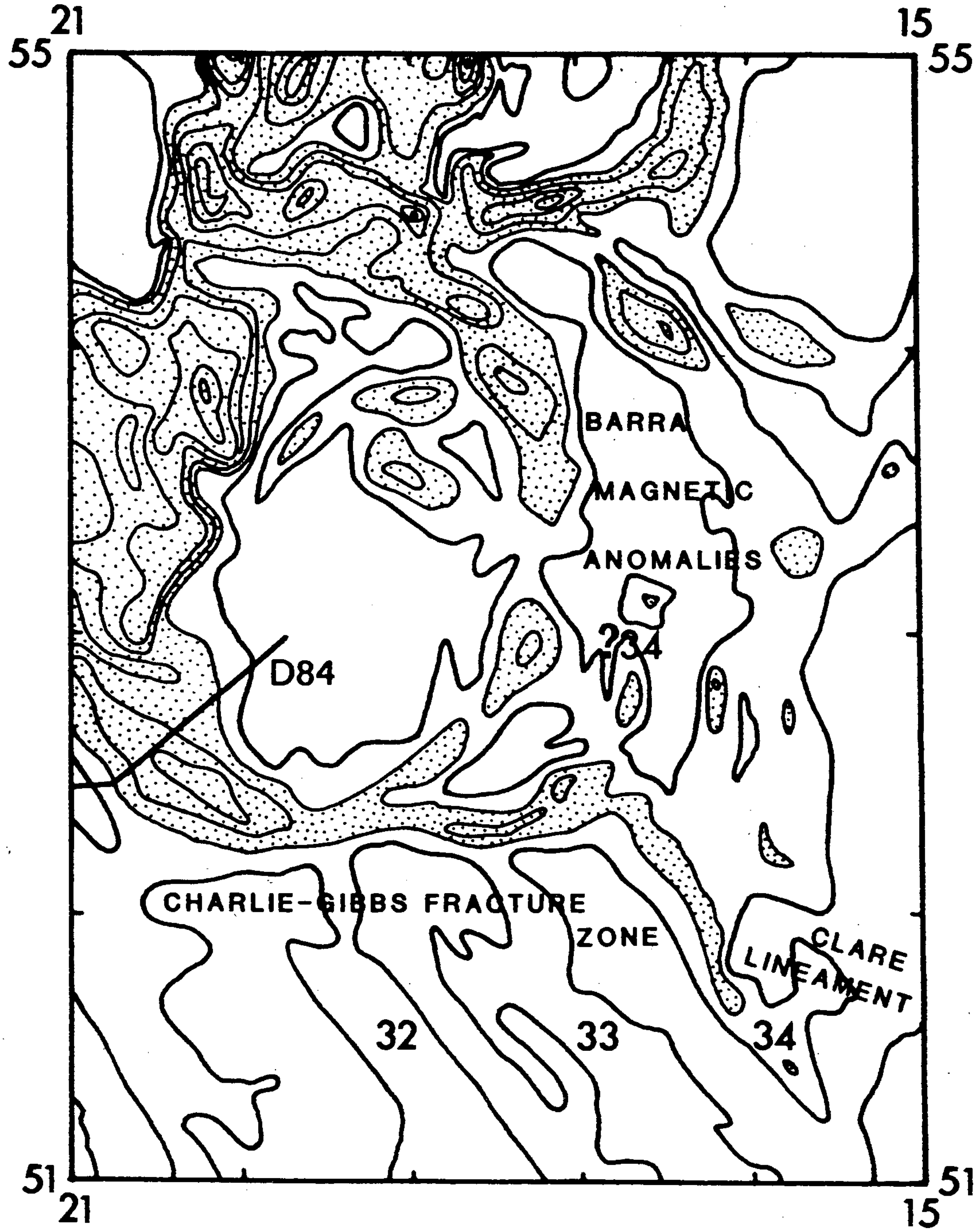


Figure 8.6 Total intensity magnetic anomaly field over southern Rockall Trough. Contours every 200 nT above 0 nT only. Bold contour is 0 nT. Stippled areas greater than 200 nT to highlight both the circular nature of the BVRS and the oval outline of the Barra magnetic anomalies. Magnetic anomalies 32, 33 & 34 are marked. Solid line marked D84 shows section illustrated in Fig. 8.8.

typical Layer 2 relief, except in small patches. It seems most unlikely that this noticeable difference in the character of the seismic basement could have developed in oceanic crust of the same age juxtaposed on either side of the CGFZ-Clare Lineament.

The seismic profiles also show that the basement in the Trough is markedly shallower than to the south (Chart 4, back pocket; Figs 4.4 and 4.5), an observation which, combined with the greater sediment thicknesses in the former locality, accounts for the shoaler bathymetry in Rockall Channel. If the two regions of oceanic crust are contemporaneous then it is necessary to invoke substantial differential subsidence since their initial formation. Is it possible that the oceanic crust in the Porcupine Abyssal Plain has undergone a normal exponential subsidence history, in common with the rest of the North Atlantic Ocean basin, while the Rockall Trough -isolated at shortly after Anomaly 34 time - has had its subsidence pattern hindered by the nearby continental masses?

An alternative explanation for the northward step in the depth to basement is expounded whereby the seismic basement seen on the reflection profiles of this work in Rockall Trough is essentially a superficial volcanic layer with normal ?mid-Cretaceous oceanic crust continuing into the Trough beneath this thick extrusive layer and disrupted only by the Clare Lineament and CGFZ. There is limited evidence in support of this idea from the CM-04 multichannel profile (SPs 3800-4200, Fig. 4.5). But there is no clear evidence to suggest that a sequence of old sediments is present between the layered volcanic basement of this work and some deeper, ancient Rockall oceanic crust. Inspection of all the seismic profiles of this study and also a fair number of multichannel tracks in the central and northern parts of Rockall Trough (courtesy of I.O.S.) indicates the persistence of the proposed layered volcanic basement throughout the Trough, including and continuous with the thicker development within the BVRS.

Thus the volcanic basement can be thought of as the final extrusive product of oceanic accretion in the Trough. It is not known if this upper layer is everywhere roughly synchronous, by some quirk of anomalous diffuse spreading, or if a normal younging of the lavas toward the axis of the basin pertains. The prediction that the thickly developed BVRS is continuous with the surrounding volcanic basement is important because the magnetic anomaly and seismic basement isochron charts (Charts 3 and 4, back pocket) are both strongly in favour of the Barra volcanic ridges extending into the northern ridge N of the CGFZ, and thereon into the ridge (and accompanying 700 nT magnetic anomaly) south of Lorien Bank (Figs 7.1 and 8.8). As the CGFZ was probably initiated at the end of Anomaly 34 a similar age must, by lateral correlation, be applicable to the BVRS and from there the remainder of the Rockall basement. There is little to suggest that the BVRS and its associated magnetic anomalies were already in existence and that the CGFZ grew in length and truncated them. If this does prove to be the case then an investigation of the Orphan Basin-Flemish Cap area of the conjugate Canadian margin would hopefully reveal a similar volcanic/intrusive province. No such structure has yet been reported from this region (Haworth 1977, 1980; Keen 1982).

In the review of the seismic stratigraphy of the Rockall region (Chapter 3) it was remarked that a noticeable thickening of the sedimentary succession occurred immediately north of the CGFZ and Clare Lineament. This increase in sediment thickness is observed on all the seismic profiles across the fracture zone and in itself appears to be a powerful argument for the Rockall Trough basement being older than the oceanic crust to the south (see Figs 4.4 and 4.5). Furthermore the disposition of the regional reflectors indicates that most of the thickening, perhaps all in some places, is caused by the introduction of a deep, often thick R7 to basement sedimentary sequence. This sequence is not recognised over Albian and younger oceanic crust south of the CGFZ, the suggestion therefore being that the pre-R7 sequence documents the age difference between the two crustal regions.

In Chapter 3 it was speculated that the R7 horizon, conspicuous as it is, represents a drifting (post-rift) unconformity marking the onset of oceanic accretion immediately to the south. This would give it a Middle to Late Albian age (c. 101 m.y. B.P.) according to the DSDP Leg 80 drilling evidence (Masson et al. 1985). Collation of the seismic reflection data shows that the pre-R7 succession occasionally exceeds 1.2 s TWT but towards the sides of the Trough is more usually 0.8 s TWT thick (Fig. 5.16). Since the pre-R7 interval everywhere appears to post-date the seismic basement the minimum age of the latter can be calculated very roughly from a knowledge of the sediment thickness and depositional rate. At SP 4100 on profile

CM-04 (FIg. 4.5) approximately 1300 m of pre-R7 sediments overlie the deep ocean-like basement reflector here, assuming a sediment velocity of 3.2 km/s. There is no information available regarding sedimentation rates at these depths in Rockall Trough. Nonetheless if a fairly typical value of 50 m/m.y. is assumed then the basement must be around 130 m.y. old (Hauterivian of Harland et al. 1982). A slower sedimentation rate, say 25 m/m.y., would imply formation of oceanic crust here during the Late Jurassic.

The important point to be gained from all this is the conflicting evidence of the seismic stratigraphy, which appears to indicate a lower-most Cretaceous or older age for oceanic accretion, and the BVRS-CGFZ association which favours the formation of Rockall Trough during the 20 m.y. or so up to the end of Anomaly 34 (101 to 83 m.y. B.P.; Masson et al. 1985). Smythe (pers. comm.) has independently recognised the variations in sediment thickness and seismic stratigraphy across the Charlie-Gibbs F.Z. and believes they are good evidence for a pre-Cretaceous age for Rockall Trough.

For the two conflicing arguments to be consistent it is necessary to invoke either an older age for the BVRS or a younger Late Cretaceous age for the R7 reflector and pre-R7 sediments. Despite the fact that the seismic basement and magnetic anomaly charts (3 and 4, back pocket; Fig. 4.12) appear to show the Barra volcanic ridges curving into the CGFZ, suggesting they are contemporaneous, it is possible that the volcanic ridges are in fact truncated. Slender evidence supporting this idea is seen around 52.5°N 17°W on both charts where the trends of the contours hint at former continuation of the BVRS to the south - presumably now located beneath the East Newfoundland continental margin.

On the other hand, for the pre-R7 sediments to be younger than Anomaly 34 (c. 83 m.y. B.P.) it would be necessary to advocate a very large and abrupt change in sedimentation rate across the CGFZ during the 20-30 m.y. following the formation of the oceanic crust in Rockall Trough and off Goban Spur. There are occasionally patches of seismically layered sediments in depressions in the oceanic crust south of the CGFZ. But there are no examples comparable to the thick pre-R7 succession seen north of that lineation (see Fig. 4.5). On all the seismic profiles that traverse the CGFZ or Clare Lineament (Figs 4.4 and 4.5) a volcanic ridge of some sort is present against which the pre-R7 sequence pinches out on the north side but is absent

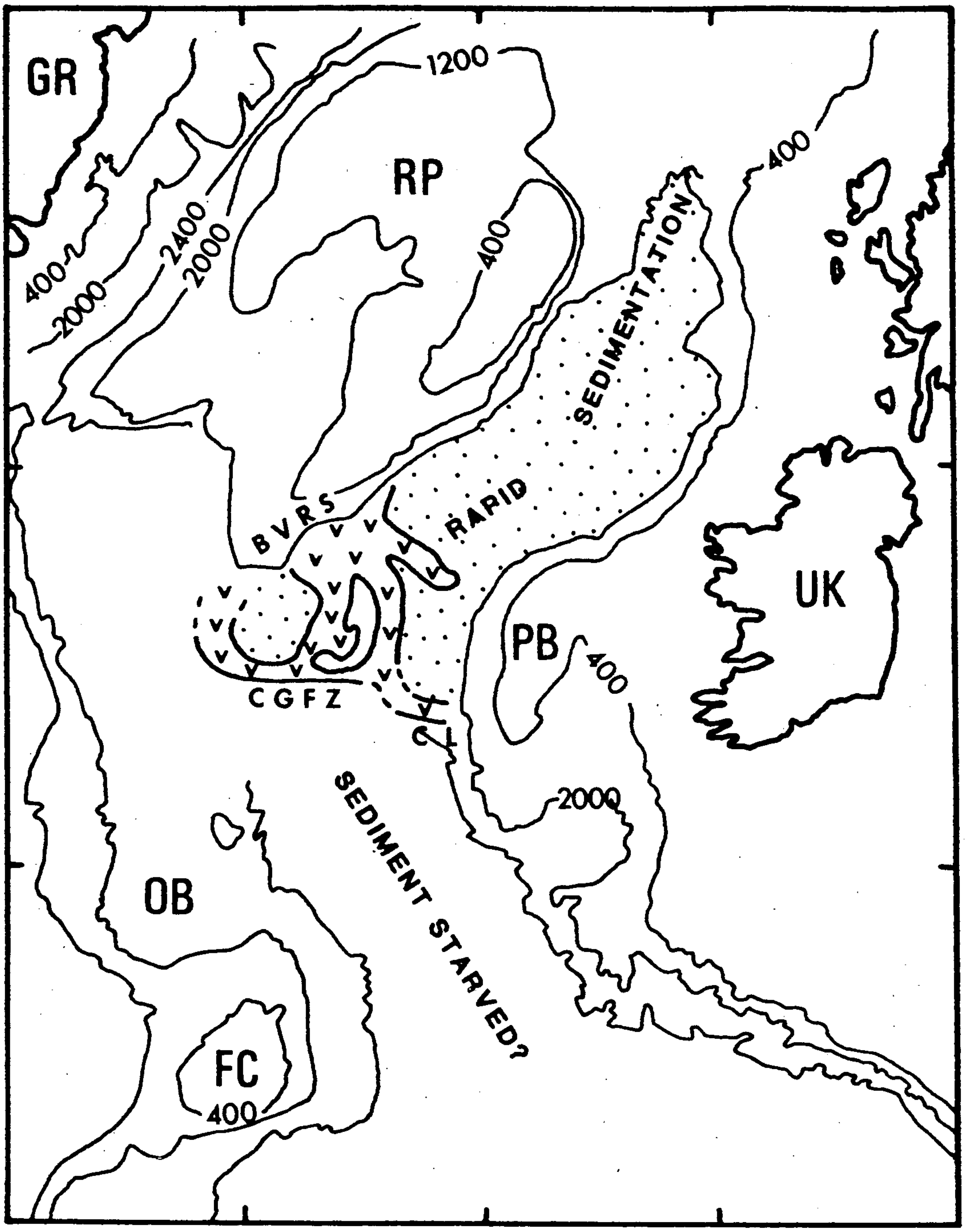


Figure 8.7 Anomaly 34 (83 m.y.B.P.) reconstruction of the North Atlantic (after Kristoffersen 1978) showing the Barra volcanic ridges (V symbol) acting as a possible barrier to sedimentation. CGFZ = Charlie-Gibbs Fracture Zone; CL = Clare Lineament; FC = Flemish Cap; OB = Orphan Basin; PB = Porcupine Bank; RP = Rockall Plateau. Bathymetry in metres.

to the south side over known Late Cretaceous oceanic crust. Is it possible that an unbroken volcanic barrier was associated with the CGFZ right from its inception and that this ridge prevented sedimentation to the south (Fig. 8.7)?

Although the existence of a continuous volcanic ridge appears, from the reflection profiles, to be quite plausible, it is most unlikely that sedimentation during the pre-R7 interval would have been almost entirely absent to the south of the CGFZ. This would demand highly directional sediment supply along the Trough from the north-east. The distribution of the R7-basement succession in the Trough (Fig. 5.16), however, indicates sediment transport from the continental margins, which at this time were presumably recently rifted and uplifted. It is difficult to imagine the absence of similar provenances to the south; Hallam (1971) states that an oceanic seaway with a largely complete marine sequence was present stretching from the Gulf Coast, U.S.A. to the Arctic by the Late Jurassic, and possibly earlier.

An important clue in this debate may be provided by the section of the Discovery 84 single-channel profile from 2000/170 to 0500/171 (Fig. 8.8). Here the seismic profile passes from the enigmatic Rockall Trough into the deeper unnamed basin to the west (extension of Eriador Plain) which is believed to be floored by oceanic crust of roughly A 34 and younger age (Roberts 1975; Kristoffersen 1978; Srivastava 1985). The section of profile (Fig. 8.8) shows broad similarities to those crossing the CGFZ and Clare Lineament. A thick sequence of sediments overlying a poorly imaged but irregular basement terminates against a proposed volcanic ridge or ridges (2300/170 to 0200/171). To the south-west of this ridge a conspicuous step-down in the bathymetry results from thinning of the sedimentary overburden and a lowering of the basement level. Again the pre-R7 succession and probably part of the deeper R5 to R7 succession are absent over the Late Cretaceous oceanic crust. Observing the differences in structure and seismic character across this boundary it is difficult to see how the basement on either side of the volcanic ridge (Chart 3 and Fig. 8.6) can possibly be of the same age. The basement beneath Rockall Trough must surely be substantially older than the Late Cretaceous oceanic crust at the

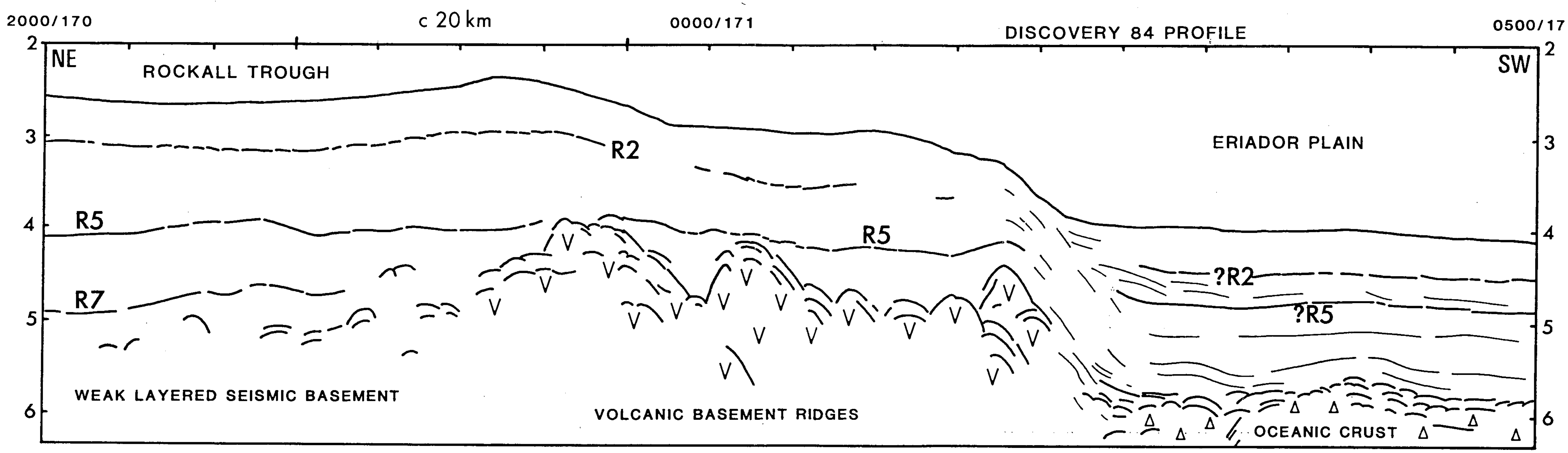


Figure 8.8 Refer to Fig. 8.6 for location

south-western end of the figure. As there is no fracture zone present here then there should be an ocean-ocean contact either beneath the volcanic ridge - if that is superficial - or actually at the edge of the ridge at 0200/171 (Fig. 8.8).

A small number of workers, notably Russell (1976), Russell and Smythe (1978) and Haszeldine (1984), have advocated an older upper Palaeozoic age for the accretion of oceanic crust in Rockall Trough. While many of their arguments favouring a much older spreading history are elegant, especially those of Haszeldine (1984), a popular case refuting such a dating relies on the implication that it would necessitate the formation of a small, narrow ocean basin in the middle of a large continental mass, the Pangaean supercontinent. However, the present author sees no objection to the development of just such a plate geometry. After all the same situation must have pertained during the early stages of rifting and drifting between Africa and North and South America from the middle Jurassic to middle Cretaceous. Indeed if the age of the Trough is put at the Late Jurassic - Early Cretaceous, as determined tentatively from the seismic stratigraphy earlier in this section, then an isolated ocean basin is still required since the crust between Goban Spur and Newfoundland-Flemish Cap had not yet been accreted. Russell and Smythe (1978) and Haszeldine (1984) fulfil the requirement of rigid plate tectonic theory by introducing large transcurrent faults at the northern and southern ends of the Rockall-Faeroe ocean, faults which transfer the spreading motion into the Arctic and Tethyan realms respectively (Fig. 8.9). This does not seem an unreasonable explanation.

But there are other pieces of evidence that detract from an upper Palaeozoic age for Rockall Trough. Foremost of these is the very slow sedimentation rate that would be demanded in the Trough. If the end Stephanian (c. 285 m.y. B.P.) ages predicted for Rockall Trough by Haszeldine (1984) and Russell and Smythe (1978) are applied to the basement observed at SP 4100 on profile CM-04 (Fig. 4.5) then a maximum rate of sedimentation of about 13 m/m.y. can be calculated for the pre-R5 sediment interval. This value seems rather too low, though at DSDP site 550 on oceanic crust off Goban Spur (Fig. 1.5) sediment accumulation rates in the range 5.5 to 35 m/m.y. were reported (Granciansky, Poag et al. 1985, 550 site chapter).

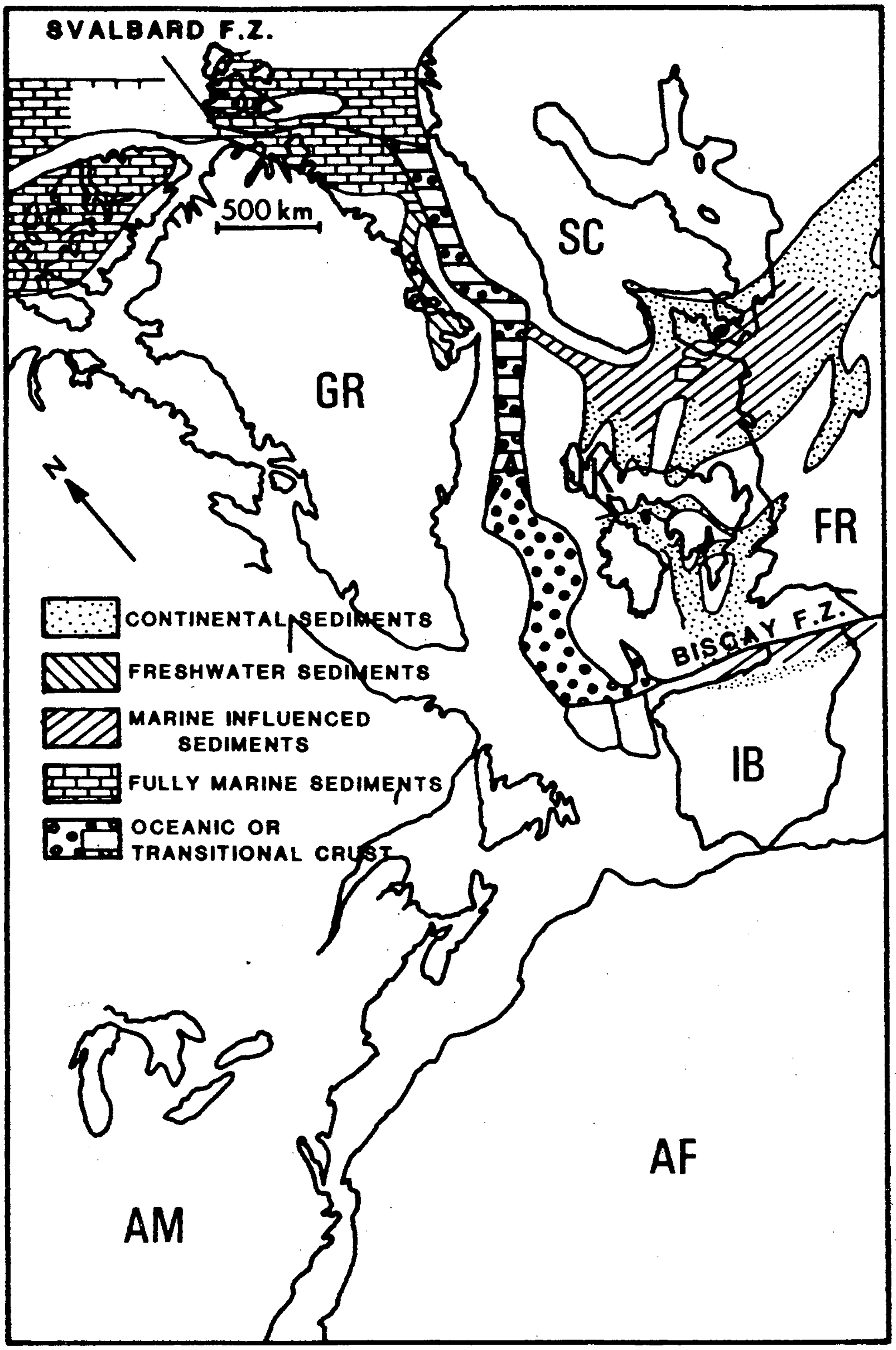


Figure 8.9 The Westphalian - Stephanian Rockall to Norway oceanic rift of Haszeldine (1984). The spreading system is terminated abruptly by the Svalbard and Biscay transcurrent zones.

Haszeldine (1984) proposed an end-Westphalian to end Stephanian (c. 300-286 m.y. B.P.) duration for spreading based mainly on stratigraphic and commercial drilling evidence (Fig. 8.9). Similarly Russell and Smythe (1978) mooted an Early Permian age for ocean formation. Both theses emphasise the possible relation between the mid-Stephanian (295 m.y. B.P.) volcanics and intrusives in central Scotland, northern England and the Oslo Graben, and the high strain rates during the final stages of continental rifting in the Rockall-Faeroe region. However, neither authors provide any direct evidence concerning the nature and age of the basement below Rockall Trough. The evidence and arguments they put forward can equally well be interpreted as marking the initial formation of a Rockall-Faeroe continental rift, not a true ocean basin. Haszeldine (1984) cites commercial drilling evidence for Visean to Westphalian B sediments on the margin of Rockall Trough, and Westphalian to Stephanian strata in the Porcupine Seabight. This information in itself does not tell us anything about the age of oceanic crust in the Trough. But it does suggest that Rockall Trough, and presumably also the Faeroe Basin, have existed as important depocentres and foci for tectonism and magmatism since the lower Carboniferous Period. Thus the longevity of the Rockall-Faeroe channel, first as a continental rift and later as an oceanic basin, now seems in little doubt.

Russell and Smythe (1978) quoted the calculated Permian magnetic vector for Rosemary Bank Seamount (Scrutton 1971) as evidence supporting the Early Permian age of the Trough. But Miles and Roberts (1981) contend that the remnant magnetic inclination of the Seamount is compatible with a Late Cretaceous-Early Tertiary age. However, multichannel seismic profiling has shown that Rosemary Bank is surrounded by Early Tertiary lavas and sills that are presumably coeval with the Thulean igneous provinces of Scotland and Greenland. Hence the age determined by Miles and Roberts (1981) may be erroneous because of these younger volcanics and sills.

The modelling of Miles and Roberts also indicates that the base of the seamount is significantly shallower than the basement in Rockall Trough, and that a layer of old sediments intervenes between them. It appears then that Rosemary Bank Seamount is unlikely to provide a realistic age for Rockall Trough. Instead magnetic modelling of Anton Dohrn seamount, some 200 km to the south, combined with the seismic stratigraphy, may be more appropriate since the reflection profiles here indicate the absence of large amounts of obscuring Tertiary lavas within the surrounding sediments. Or better still a detailed seismic and magnetic survey of the buried seamount, proposed from this study, at 54.6°N 14.7°W should provide a less ambiguous age. The seismic profiles indicate that this seamount rests on the deep layered basement of the Trough (Fig. 5.15), while the magnetic anomaly charts (Roberts and Jones 1975; Max et al. 1982; Chart 3) suggest a comparatively simple normal magnetisation.

It is useful to consider now a number of regional geophysical and geological investigations of the larger basins surrounding Rockall Trough, for there are important consequences bearing on its possible age. Integrated studies of part or all of the region from Rockall Trough to Svalbard, via the Faeroese and Norwegian margins, have been reported by Smythe et al. (1983), Hanisch (1984) and Price and Rattey (1984). Correlation of multichannel seismic profiles with commercial drilling information in the Faeroe-Shetland Channel (Ridd 1983; Smythe et al. 1983) and off mid-Norway (Price and Rattey 1984) indicates late Early Cretaceous and older sediments above a poorly imaged or absent seismic basement (Fig. 8.10). In the Faeroe Basin in excess of 1.5 km of pre-Turonian sediments are predicted, and it is possible to map out the top Early Cretaceous and perhaps top Jurassic horizons (Smythe et al. 1983).

These stratigraphies imply a Late Jurassic or older age for the poorly defined basement in the axial parts of the Faeroe-Shetland Channel (Fig. 8.10). Since it is likely that this area and its basement are continuous with the broader Rockall Trough then a similar age of formation can be accepted for the latter feature. The seismic evidence certainly favours a pre-middle Cretaceous age for the Rockall-Faeroe channel, and in doing so refutes the post-Aptian age (c. 101 to 83 m.y. B.P.) formulated by Roberts et al. (1981). In addition, if one accepts the propogating rift model of Vink (1982), and that of Hanisch (1984) developed specifically for the Rockall-Faeroe rift, then the presumed oceanic crust in Rockall Trough must be slightly older than that further north-east around the Faeroes and mid-Norway.

Hanisch (1984) dates the formation of oceanic crust from Rockall Trough to about 65.5°N loosely as mid-Cretaceous. This is based on the onset of folding in West Spitsbergen at Aptian-Albian times; deformation he believed was a result of compression on the far side

A

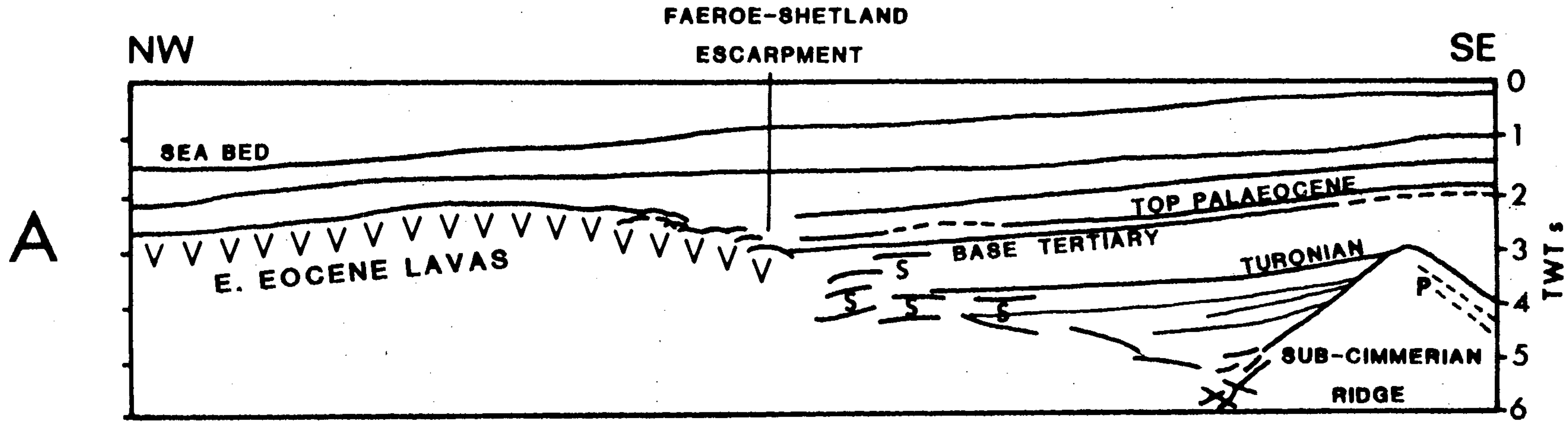


Figure 8.10a. Interpreted seismic profile and seismic stratigraphy across the Faeroe-Shetland Channel (after Smythe et al. 1983). S = sills within seds. P = Permian.

B

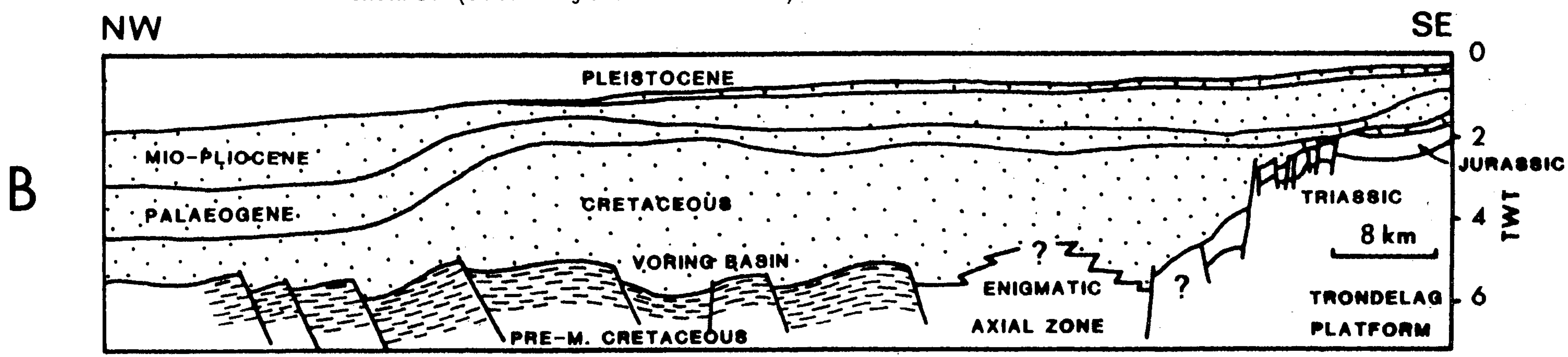


Figure 8.10b. Seismic stratigraphy and structure across the Mid-Norwegian continental margin (after Price and Rattey 1984).

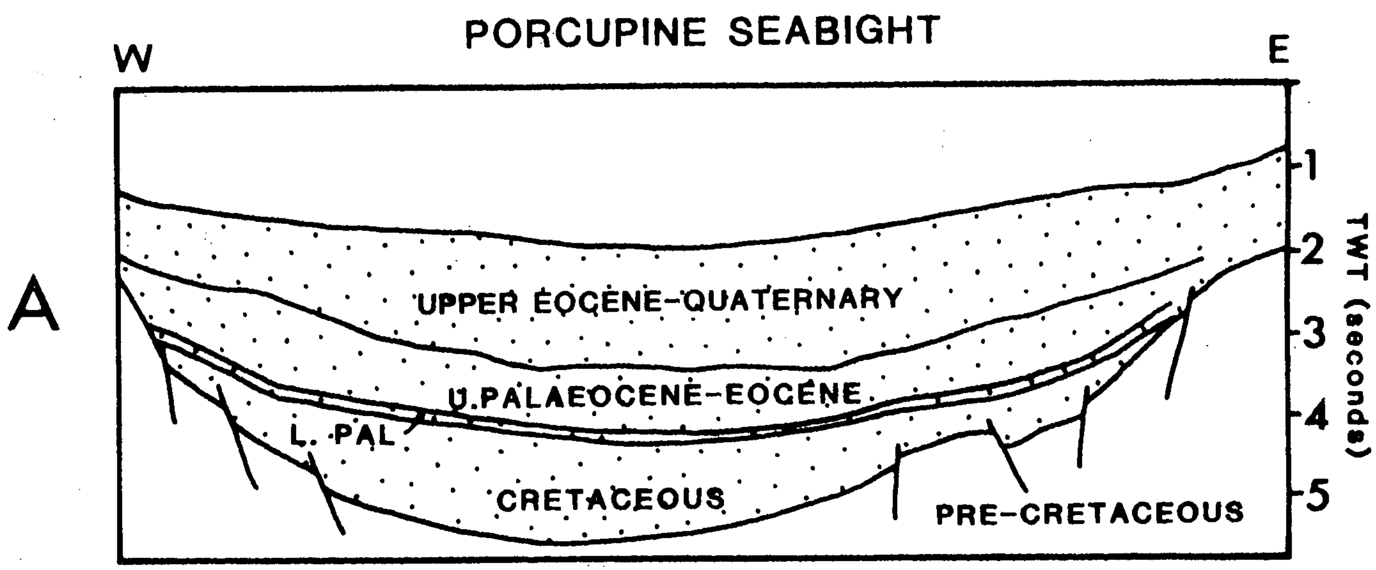
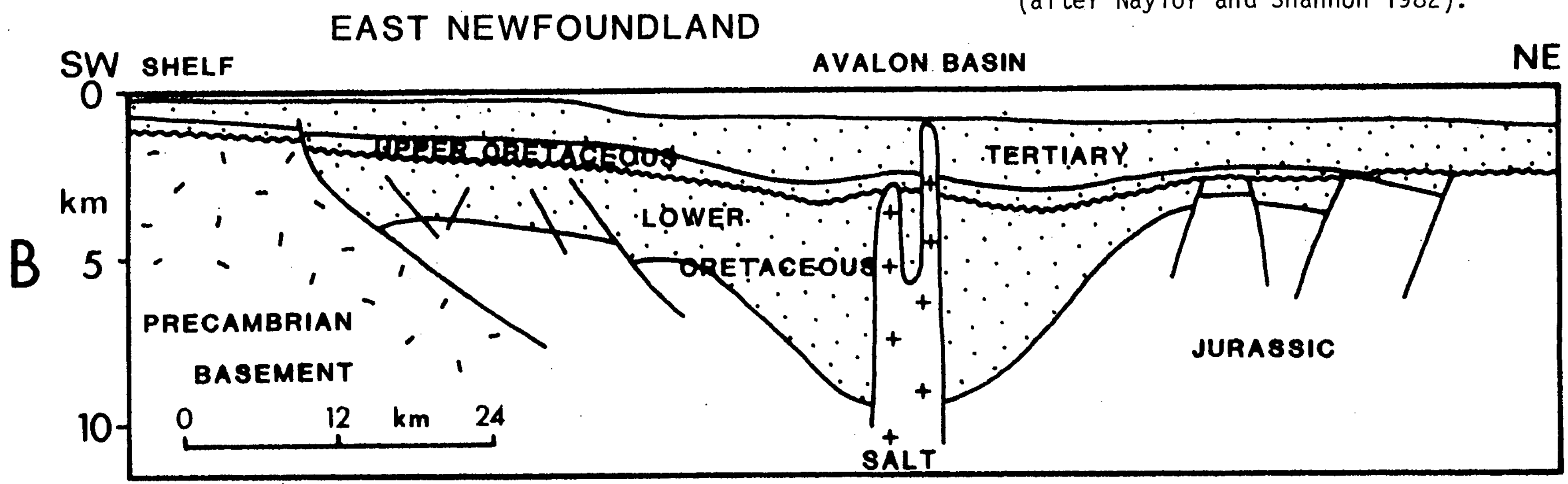


Figure 8.11 A) above. Seismic stratigraphy within the Porcupine Seabight (largely after Naylor and Shannon 1982).
B) below. Seismic stratigraphy and basement geology across the E.Newfoundland continental margin (after Naylor and Shannon 1982).



of the pole of rotation (74°N 21°E). His Coniacian to Early Maastrichtian age for the formation of southern Rockall Trough is based on an erroneous understanding and interpretation of the magnetic lineation sequence put forward by Roberts et al. (1981). To be consistent with his propogating rift model and the age of compressional deformation near the pole of rotation it would have been better for Hanisch (1984) to propose a pre-Aptian or pre-Albian age for Rockall Trough. The timing of rifting and spreading here would be expected to predate slightly that further north-east and presumably also any folding on the compressional side of the pole (at least in its vicinity).

In the reconstructions of Rockall Plateau back against the UK continental margin there is a small but significant zone of overlap between southern Rockall Bank and Porcupine Bank (Fig. 8.16). This observation suggests that Porcupine Seabight was formed by rotation of Porcupine Bank to the north-west contemporary with rifting in the Rockall Trough. The similarities in bathymetric trend and depth of the Trough and the Seabight also supports their having a common history. The drilling and seismic profiling evidence (Naylor and Shannon 1982) indicate Permo-Triassic and younger rift-valley type sediments overlying stretched continental crust. There is little evidence for oceanic crust beneath Porcupine Seabight, although a narrow zone of thin crust (<10 km) in the axial parts of the southern Basin is of enigmatic origin (Masson and Miles, in prep.).

Within the Mesozoic sequence of the Seabight a marked, often faulted, unconformity is present at the base of the Cretaceous - the late Cimmerian unconformity (Fig. 8.11a). Masson and Miles (in prep.) correlate the unconformity with the reactivation of the basin during the development of the north-west trending rift which was later to evolve into the oceanic basin west of Goban Spur. The age of this rifting phase was put at the lower Barremian to middle Albian (Masson et al. 1985). The same pattern of thinned continental crust with a thick Mesozoic and Cenozoic sedimentary succession is observed on the conjugate Canadian continental margin (Fig. 8.11b; Keen and Barrett 1981; Keen 1982; Naylor and Shannon 1982). The base Cretaceous unconformity also seen here is again probably a manifestation of the lower Cretaceous rifting preceding the formation of oceanic crust in the North Atlantic south of the Charlie-Gibbs

Fracture Zone. Thus there is no visible evidence in the stratigraphy of the Porcupine Seabight or Orphan Basin that provides any support for a pre-middle Cretaceous age for Rockall Trough.

8.4 Rockall Trough: A three-fold geological evolution

Early continental rifting

There is now a fair amount of evidence from geophysical, drilling and stratigraphic studies suggesting that the present SW-NE trending Rockall Trough marks a line of continental rifting and sedimentation dating back perhaps to the Early Carboniferous (Fig. 8.12; Haszeldine 1984), but almost certainly to the Permo-Trias (Hallam 1971; Russell 1976; Russell and Smythe 1978; Ziegler 1981, 1982). Support for these early rifting ages comes from seismic refraction investigations over the continental margin west of the Hebridean Islands which define Mesozoic and possibly upper Palaeozoic sedimentary basins predating the formation of presumed oceanic crust in Rockall Trough (Jones 1978; Bott et al. 1979). Smythe et al. (1978) have questioned the Mesozoic age of the basin detected by Jones (1978) on the grounds of the refraction velocities, suggesting instead that the refractor represents widespread Early Tertiary basalts.

Within southern and central Rockall Trough the data available to this study do not permit the age and evolution of these early rifting phases to be adequately discussed. The rapid thinning of largely undeformed sediments from the Trough onto the continental slope precludes any meaningful seismic stratigraphic correlations which might enable us to shed some light on the age of the downfaulted basement beneath the Rockall margins (see Chapters 3 and 5). Likewise no detailed seismic refraction work has been performed over these southern margins which might indicate the presence of substantially older rifted sedimentary basement. For these reasons the reader is referred to Chapter 1 of this work and the syntheses of Ziegler (1981), Naylor and Shannon (1982) and Haszeldine (1984) for a review of the pre-spreading history of the Rockall area.

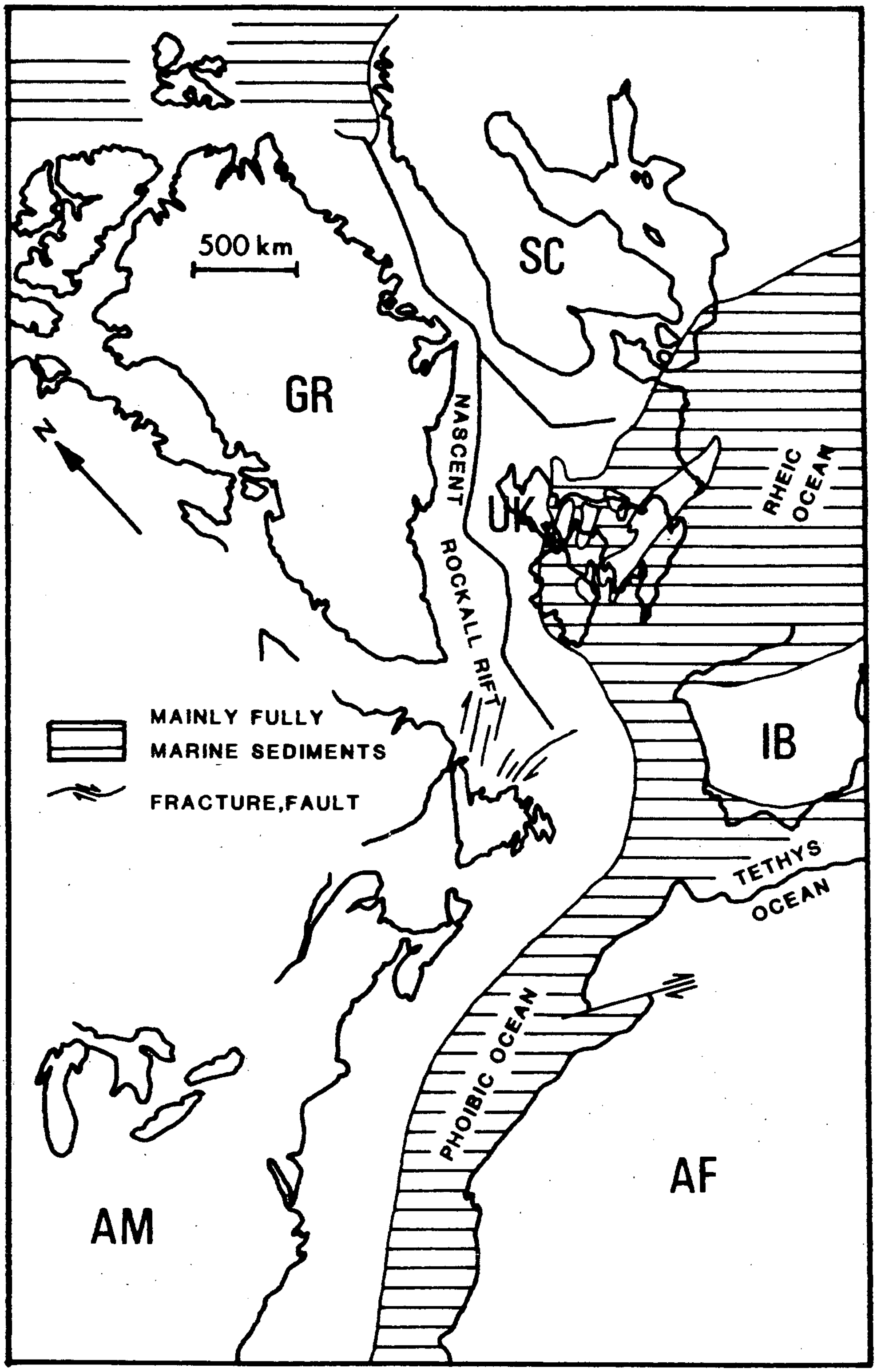


Figure 8.12 Early Visean reconstruction of Laurasia illustrating the earliest extension phase along the Rockall-Faeroe-Norway rift system. Redrawn from Haszeldine (1984).

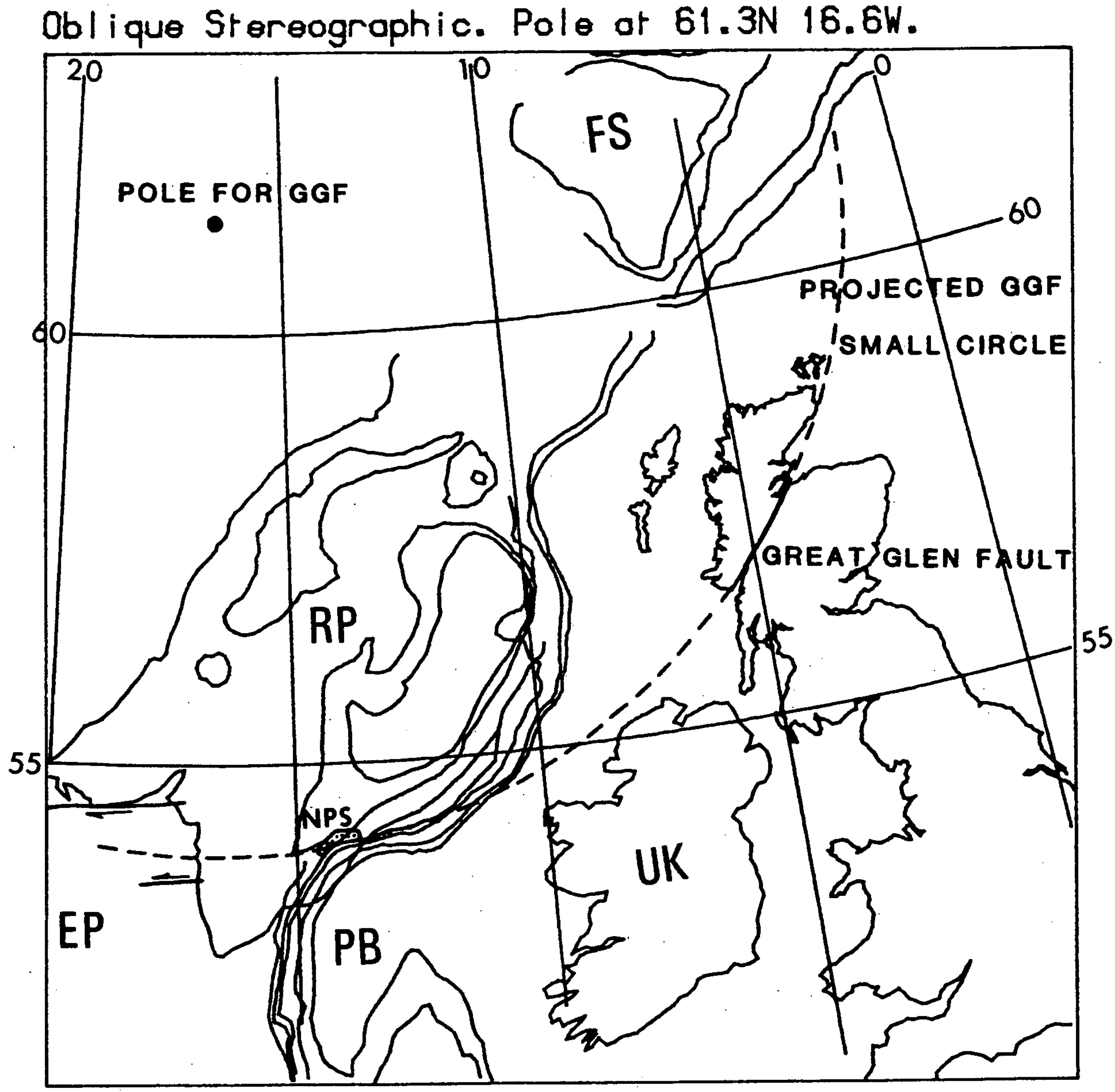


Figure 8.13 Small circle plot of the transcurrent Great Glen Fault superimposed on the pre-spreading position of Rockall Plateau predicted by Hanisch (1984, pole at 74 N 21 E). GGF pole after McQuillin et al. (1982) EP = Eriador Plain; FS = Faeroe Shelf; NPS = North Porcupine Salient; PB = Porcupine Bank; RP = Rockall Plateau.

The SW-NE trend of the two continental margins of Rockall Trough may reflect control on the direction of early rifting by pre-existing Caledonian structures - an influence that is clearly seen in the basement geology of Scotland and north and west Ireland. A number of authors have commented on the possible genetic connection between the position of the main Caledonian lineaments and the shape of and re-entrants into the Rockall and Porcupine Banks (e.g. Bailey 1975; Riddihough and Max 1986; and Megson 1983). McQuillin et al. 1982 postulated that the Great Glen Fault was a major crustal transcurrent feature and accordingly should approximate a small circle about a pole of rotation. This pole they calculated to be at roughly 61.3°N 16.6°W from the surface outcrop of the fault in Scotland and the Shetland Isles. In Figure 8.13 the Great Glen Fault is extrapolated along the small circle locus predicted by the rotation pole. Also shown is a reconstruction of the Rockall Plateau 5° towards the south-east about Hanisch's (1984) pole of rotation - the one that appears to produce a good fit of the isobaths.

It is observed that the projected Great Glen Fault is closely coincident with the northern edge of Porcupine Bank and so may have played a large role in determining the latter's position. It is also noted that the North Porcupine Salient is developed where the GGF leaves the Bank and that between this point and its conjugate site at the north-west margin there are changes in the gravity field and seismic basement contours along the Trough (Charts 2 and 4). Further west the projected GGF may be partly responsible for the location of the Lorien Channel and the development of the offset margin or a speculated transform fault (Roberts 1975) in oceanic crust below Eriador Plain. The geometry depicted in Figure 8.13 is obviously an over simplification; the GGF is unlikely to exactly follow a small circle path, while another complication arises from the fact that Porcupine Bank requires a small degree of translation to the SE to restore the extension in Porcupine Seabight.

Mesozoic development of a small ocean basin

In their synthesis of the Mesozoic development of sedimentary basins across the pre-drift fit of Europe, Iberia and North America Masson et al. (1984; Fig. 1.9) describe a SW-NE trending system of rifts that were active during the Late Triassic to Early Jurassic stretching from the Porcupine Bank-Orphan Knoll-Grand Banks province in the NW to the Aquitaine Basin in the south-east. A second period of rifting in the Late Jurassic-Early Cretaceous was a precursor to the Late Cretaceous-Tertiary opening of the southern North Atlantic. A middle Jurassic tectonically quiet period was believed to have separated the two rifting phases. It seems likely that the Rockall-Faeroe channel was also caught up in the first early Mesozoic rifting phase and it is possible that the basin continued southward to connect with the Orphan Basin.

In section 8.3 of this chapter, however, it was concluded that the controversial oceanic crust in the Rockall-Faeroe system was accreted at least during the Early Cretaceous and quite possibly the Late Jurassic. This implies sea floor spreading contemporaneous with widespread continental rifting over a large area to the immediate south, a situation that is rather unexpected. Furthermore, if the middle Jurassic was a period of tectonic quiessence (Masson et al. 1984) then the rifting phase preceding sea floor spreading in Rockall Trough must have been very short. The abrupt steep margins bounding the Trough and their thin syn-rift sedimentary cover favour a rapid rifting event, though presumably some degree of thinning must have been effected during the Carboniferous and E. Permian extensional episodes (Russell and Smythe 1978; Haszeldine 1984), and possibly during the Late Triassic-Early Jurassic (Masson et al. 1984).

As also stated in section 8.3 the major drawback of advocating a Late Jurassic-Early Cretaceous age for Rockall Trough as an ocean is that at this time there was no oceanic crust between Goban Spur in the north and SE Newfoundland-Iberia to the south (Masson and Miles 1984; Fig. 8.14). Hence it is necessary to explain the isolated accretion of c.200 km of oceanic crust in Rockall Trough while the surrounding regions were just undergoing initial rifting. The gradual narrowing of the ocean north-east through the Faeroe region and on to the Norwegian margin can be adequately explained by the progressive approach toward the supposed pole of rotation near Svalbard (Hanisch 1984). But the abrupt termination of 200-250 km of oceanic basement at the CGFZ and Clare Lineament (Fig. 8.17) is difficult to justify on plate geometric grounds.

One possible explanation supposes that the Rockall-Faeroe channel formed as the earliest manifestation of the northward propogation of the already well developed Central Atlantic (Fig.

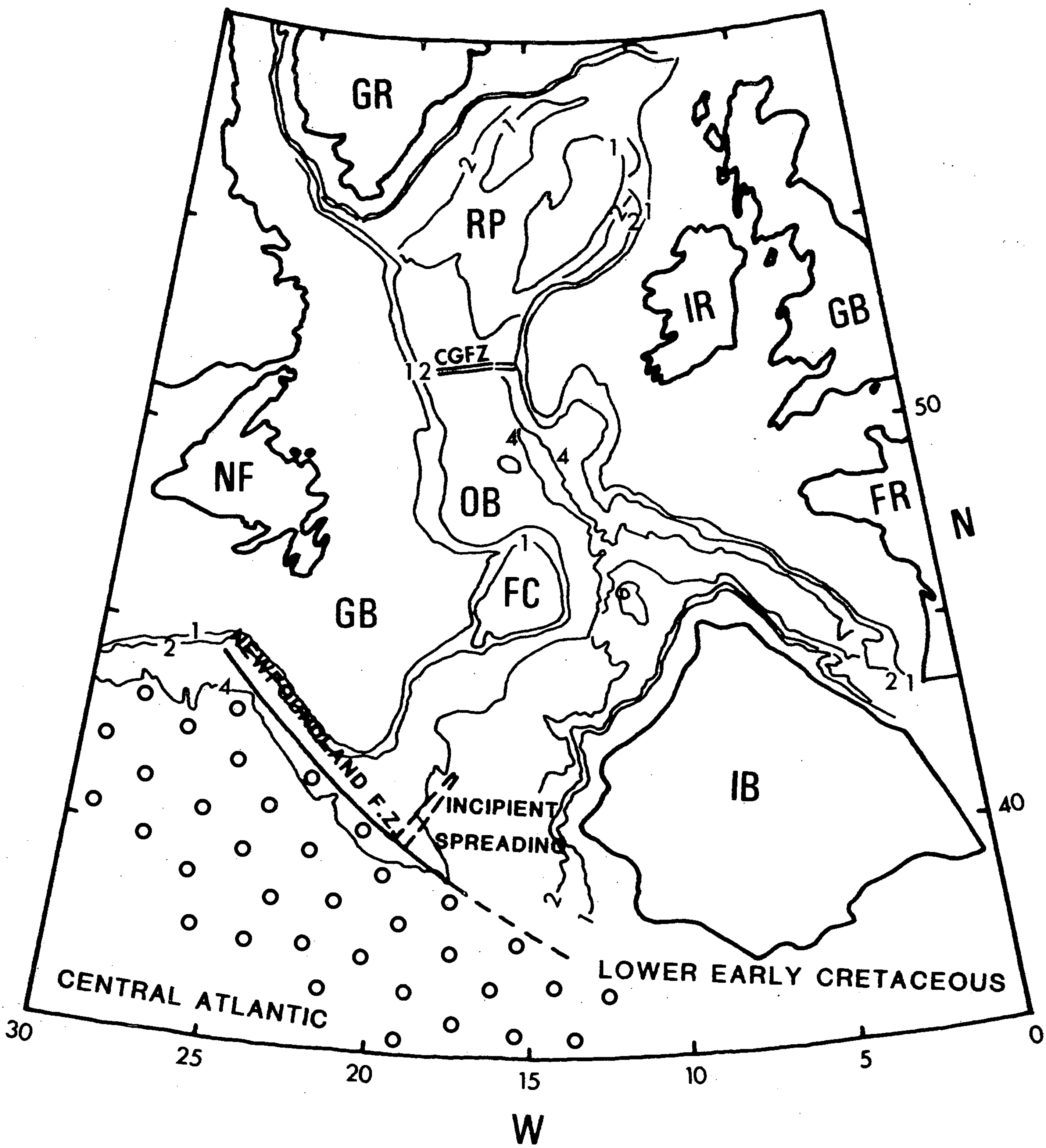


Figure 8.14 Lower Early Cretaceous reconstruction of the North Atlantic region, largely after Masson and Miles (1984). Bold open circles indicate approximate extent of oceanic crust in the Central Atlantic. Bathymetry in km. CGFZ marks future site of Charlie-Gibbs Fracture Zone and Clare Lineament. GB = Grand Banks; FC = Flemish Cap; NF = Newfoundland; OB = Orphan Basin; RP = Rockall Plateau.

8.14; Le Pichon et al. 1977; Emery and Uchupi 1984, Fig. 366). This would require extensive oceanic crustal development and/or continental thinning in the zone from Porcupine Bank-Orphan Knoll to the Newfoundland Fracture Zone. There is no evidence for oceanic crust of pre-Hauterivian age in this region, but there are large areas of presumed distended continental material in the Orphan Basin, below and south of Galicia Bank, and south-east of the Grand Banks - all at intermediate to abyssal depths, similar to Rockall Trough. In Figure 8.15 the continents have been reconstructed to their pre-drift position except the Rockall Plateau which remains in its proposed Late Jurassic-Early Cretaceous position. This schematic figure highlights the possible continuation of the Rockall Trough into the Orphan Basin and from there, tentatively, into the Grand Banks-Galicia Bank region via Flemish Pass and Goban.

While the shape and depth of the isobaths favour simultaneous development of the Rockall and Orphan Basins the geophysical investigations in the latter area support a continental affinity for the deep geological basement (Fig. 8.11b; Keen and Barrett 1981; Naylor and Shannon 1982). This author has speculated as to whether the ambiguities that have plagued our understanding of Rockall Trough have also clouded the interpretations of the thin crust beneath Orphan Basin - as little as 15 km thick. However, the seismic refraction, magnetic and stratigraphic studies here indicate, convincingly, the presence of Palaeozoic continental crust; it is extremely unlikely therefore that oceanic crust continues from Rockall Trough into Orphan Basin.

Nevertheless it is still necessary to account for the extensive continental thinning that has taken place in the Orphan Basin area, a factor that has been ignored in the Anomaly 34 North Atlantic reconstructions so far presented (e.g Le Pichon et al. 1977; Kristoffersen 1978; Srivastava 1978). Masson et al. (1984) include Orphan Basin in the Late Triassic-Early Jurassic rift system connecting Porcupine Seabight to the Grand Banks (Fig. 1.9), and then later in the extensive Late Jurassic-Early Cretaceous rift zone which was to define the opening paths of the expanding North Atlantic Ocean (also Masson and Miles, in prep.). Is it possible that, during this second phase, sea floor spreading could create in excess of 200 km of ocean

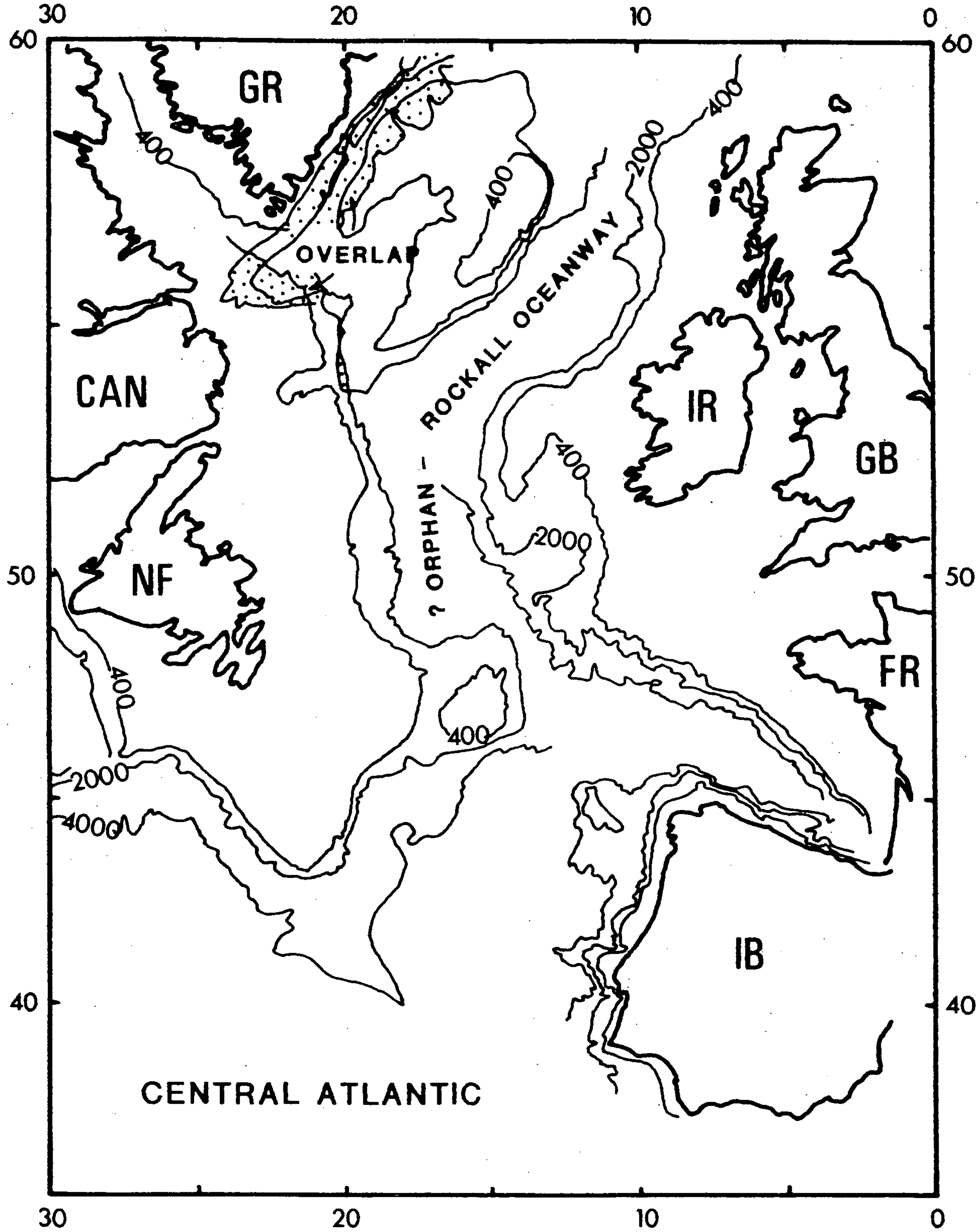


Figure 8.15 Kristoffersen's (1978) pre-drift reconstruction of the N. Atlantic continents but with Rockall Plateau redrawn in its conjectured post-opening position, i.e. as it is today with respect to NW Europe. Stipple indicates region of overlap with Greenland and Canada. Bathymetry in metres. Abbreviations as for Fig. 8.14.

basement in the Trough, while a few tens of kilometres to the south across the CGFZ (more strictly its precursor, the Clare Lineament) all this motion was taken up by continental extension? It does appear rather unlikely: an argument for continental crust in the Trough perhaps?

Russell and Smythe (1978) and Haszeldine (1984; Fig. 8.7) in attempting to account for the same problem, advocated large transcurrent faults at the northern and southern ends of the Rockall-Faeroe ocean, the southern fault transferring the desired motion into the Mediterranean-Tethyan realm along the disputed Biscay transcurrent zone. Although their's are elegant, plausible models the geometries presented in both instances are incorrect. The southern limit of presumed older ocean crust in the Trough is marked by the CGFZ and Clare Lineament (Figs 8.14 and 8.15); it is impossible to define a transcurrent feature that would have been colinear with these discontinuities as well as the northern margin of the Bay of Biscay. One further point: the old position of Rockall Plateau depicted in Figure 8.14 shows a marked amount of overlap with the Greenland and Canadian continental shelves. No allowance was made for this in the schematic reconstruction. While some of the overlap can undoubtedly be accommodated by palinspatically restoring the distension at the continental margins, some adjustment to the positions of the continental blocks is necessary if such a Late Jurassic-Early Cretaceous geometry ever obtained.

Several authors have presented poles of rotation and reconstructions for the Rockall Plateau, some based on regional geophysical and geological constraints, others merely on the need to appropriately occupy the pre-drift gap between Europe, Greenland and North America (Fig. 8.16). Of those depicted here only the reconstructions of Hanisch (1984) and Haszeldine (1984) produce a good fit of the isobaths on each margin (Figs 8.16e and f). Although it is realised that the bathymetric trends do not usually mirror exactly the continent-ocean transition, the distinct inflections in the isobaths in the northern parts of the Trough can be used to assess the quality of the fits. Unfortunately there is scant evidence from the basement in the Trough which might be used to define the spreading direction. The data used in this study, especially the seismic profiles, do not lend any support for the NW-SE fracture zone trends expounded by Roberts et al. (1981).

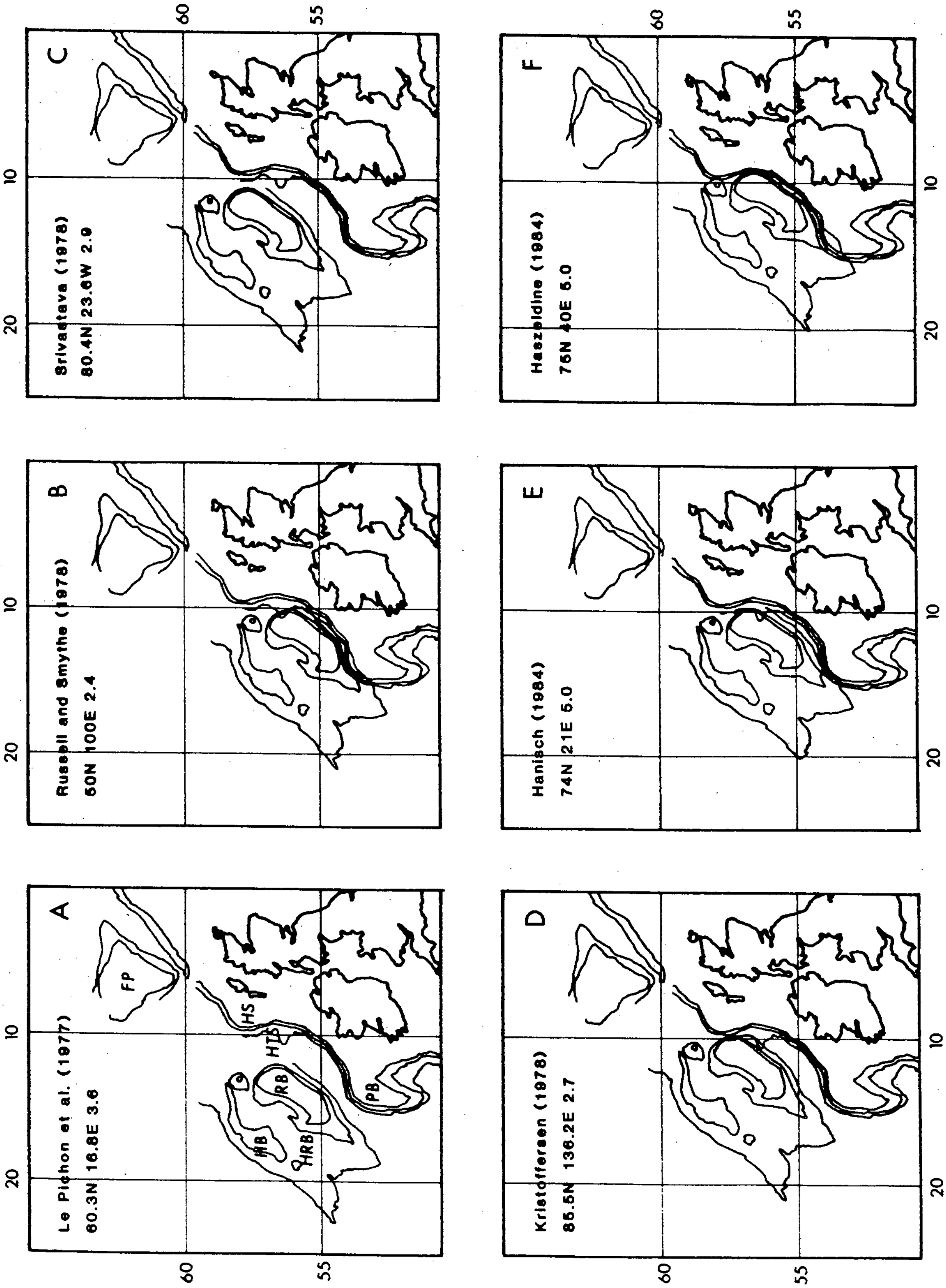


Figure 8.16 Six palaeoreconstructions of Rockall Plateau back towards NW Europe. Refer to text for discussion. Reference and pole & angle of rotation given in each figure. All rotations are counterclockwise. FP=Faeroe Platform; HB=Hatton Bank; HRB=Hatton-Rockall Basin; HS=Hebridean Shelf; HTS=Hebrides Terrace Seamount; PB=Porcupine Bank; RB=Rockall Bank. Computer reconstructions by SUPERMAP Fortran program.

However, there are a number of independent lines of evidence that may suggest a rough opening direction within the Trough. Firstly, the Clare Lineament is believed to represent one half of a fracture zone within the Rockall oceanic crust (Fig. 8.17). The other half must presumably now be located at similar depths near the northern edge of the Orphan Basin. The seismic basement chart (Chart 4, back pocket) gives an orientation of 100°-105° for the Lineament. The Charlie-Gibbs F.Z. is a gently curved continuation (bearing 95° to 100°) of the Clare Lineament, the two features meeting at about 16.7°W. Secondly, a direction of c. 110° was measured from the bathymetric chart (Roberts et al. 1979a) by matching as closely as possible the main inflections on either margin. Thirdly, the trend of any fracture zones in the Trough can be calculated on a stereonet if the pole of rotation is known. Using the poles supplied by Hanisch (1984) and Haszeldine (1984) a spreading direction of 106° and 110°, respectively, was calculated for the Rockall area. This last method is not strictly independent (since it relies on a prior knowledge of the rotation pole) but it does confirm the possibility of a 100°-110° spreading direction (280°-290° in the opposite sense) within Rockall Trough.

Within and around Goban Spur Sibuet et al. (1985) have mapped out a number of SW-NE trending faults, scarps and ridges in both continental and oceanic crust. In particular their newly described Pastouret Ridge was interpreted as a reactivated oceanic fracture zone. The N70°E bearing of this ridge was thought more likely to represent the early spreading direction off Goban Spur and Porcupine Bank, a trend that reflects underlying Caledonian or Hercynian structural control.

If the spreading directions for the two regions are correct, and if the oceanic domains in both are contemporaneous, then the 30° to 40° difference in opening trends is going to necessitate unusual (and unrecorded) extension across the CGFZ-Clare Lineament or, even more unlikely, compression in the European margin. Either one (or both) of the predicted spreading directions is incorrect, or the two oceanic provinces were formed at different times under different conditions. The latter option would agree with this author's speculations concerning the older age of Rockall. A 70° spreading direction in the Trough can be refuted simply on bathymetric grounds.

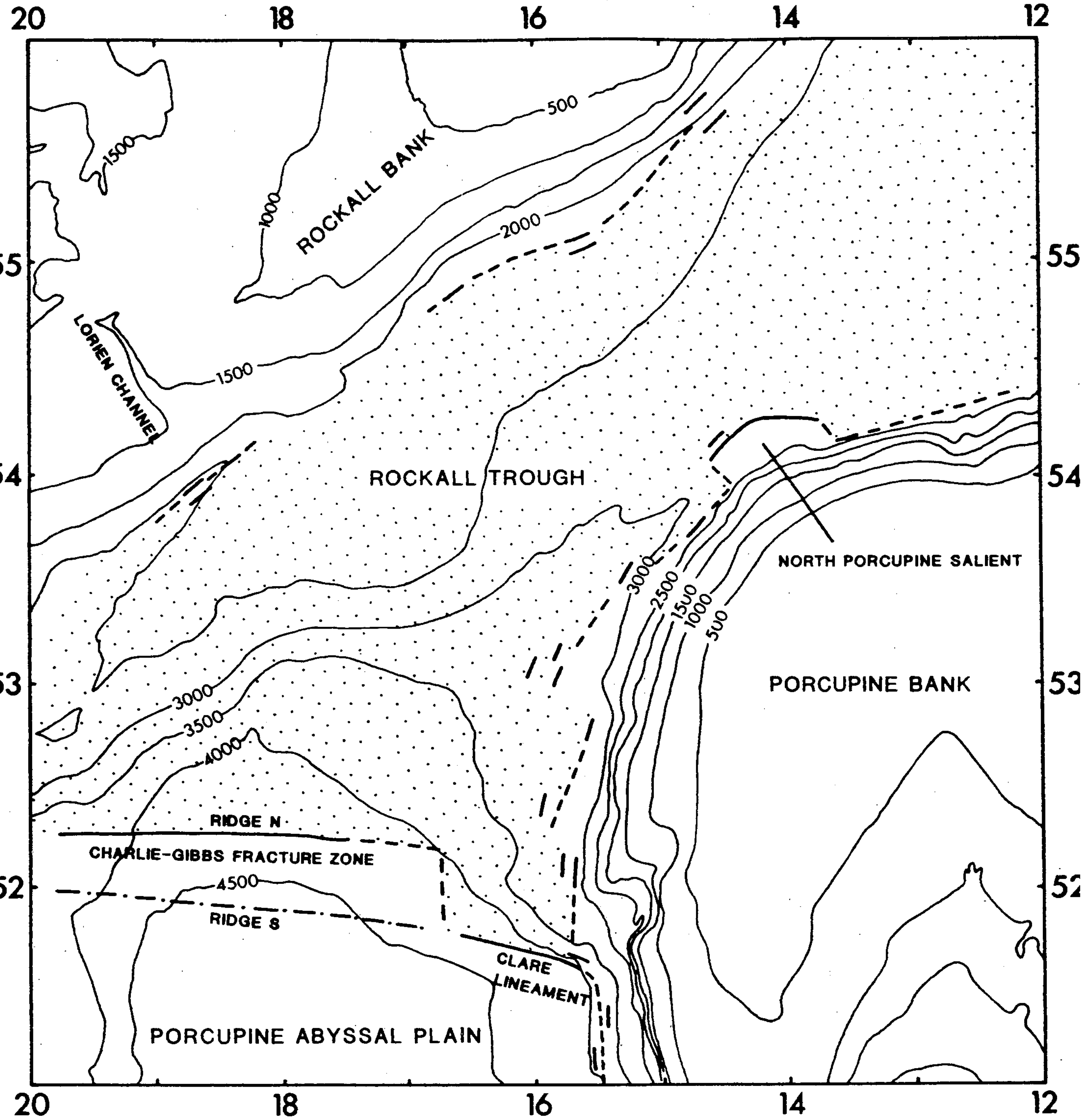


Figure 8.17 Extent of presumed oceanic crust (stippled region) in southern and central Rockall Trough. Refer to text for discussion . Based on the seismic and gravity data of this study. Solid line marks proposed continent-ocean transition, dashed line where inferred or extrapolated. Well documented Late Cretaceous oceanic crust occurs to the south of the Charlie-Gibbs Fracture Zone.

Further conflicting evidence is apparent from the structural investigations of the Porcupine Bank and Seabight. The reconstructions of Figure 8.16 show overlap of the order of 50 km between Rockall and Porcupine Banks which suggests that initial spreading in the central and northern areas of Rockall was associated with continental stretching alone in southern Rockall Trough and Porcupine Seabight. However, the studies of Lefort and Max (1984) and Masson and Miles (in prep.) both reveal due NW-SE rifting trends in the Seabight, still 25°-35° oblique to the direction predicted for the Trough. Here again it is possible that the direction in the Trough is wrong and should be adjusted to NW-SE. Otherwise the Porcupine Seabight would have to predate the Trough and an explanation would then be needed for the variable width of oceanic crust in the latter area. Was there non-uniform accretion? Or is there poor correspondence between the location of the continent-ocean transition and the bathymetric contours (see Fig. 8.17)?

Late Mesozoic-Cenozoic sea floor spreading in the North Atlantic

From roughly Late Albian times (c. 100 m.y. B.P.) onward sea floor spreading between Europe and North America and Greenland led to the progressive isolation of Rockall Trough as a somewhat older ocean basin. The opening history of the region as a whole, as documented mainly from the oceanic magnetic lineations, is treated fairly comprehensively by Srivastava (1978), Kristoffersen (1978) and Emery and Uchupi (1984). For the southern North Atlantic the studies of Le Pichon et al. (1977) and Masson and Miles (1984) should be referred to. Some of the important features in Rockall Trough, such as the Feni Ridge drift or the intra-sedimentary sills and lavas, can be related to distinct stages in the separation of the North Atlantic. The important geological events in Rockall Trough resulting from the various spreading phases are discussed with reference to Fig. 8.18.

During the Late Jurassic-Early Cretaceous it is assumed that some form of oceanic accretion occurred which gave rise to the main structures observed presently beneath the continental margins and the Trough (Fig. 8.18a). The absence of a median ridge or welldeveloped inward-facing scarps, together with the slightly greater crustal thickness of c. 7 km beneath the Trough compared with the more usual 5-6 km in the major oceans, suggests that the manner of accretion was atypical. The analogies that have been drawn with the Gulf of California (e.g. Roberts et al. 1981) are appropriate in so far as there are indications of diffuse accretion associated with rapid clastic sedimentation, perhaps in shallow water. But it should be remembered that the tectonic styles in the two areas are very different.

In southern Rockall Trough approximately 200 km of fairly uniform oceanic crust are indicated by the seismic profiles and the gravity models (Fig. 8.17). In the central parts of Rockall Trough the increased separation of the continental slopes hints at a greater width of oceanic crust. The seismic profiles are inconclusive in this respect because of the widespread presence of obscuring lavas below the north-west margin of the Trough. The nature of the southern termination of the Rockall ocean crust is uncertain. Was there an abrupt change to highly thinned continental crust across a nascent Charlie-Gibbs F.Z.? Or did oceanic crust persist to the south but is now to be found adjacent to Flemish Cap-Orphan Knoll? The evidence for thinned continental crust below Orphan Basin favours the former option.

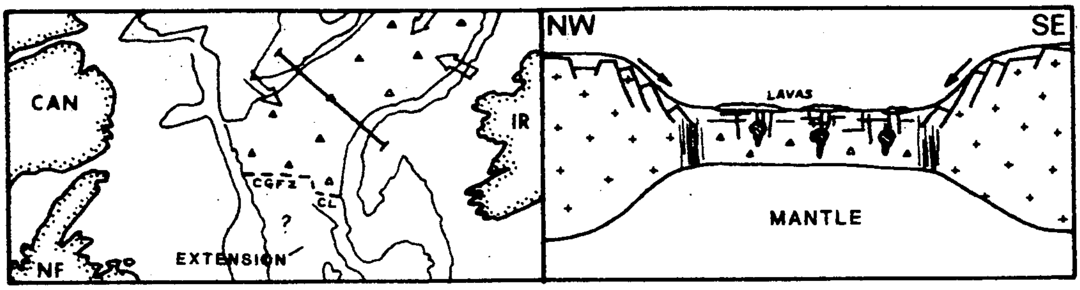
Towards the end of this main phase of sea floor spreading, perhaps during the late Early Cretaceous, a stronger pulse of oceanic accretion is believed to have formed the BVRS and possibly the Moho elevation beneath the Barra free-air anomaly (Fig. 8.18b). The thinner crust here (c. 5 km, Fig. 6.2) and the coherent magnetic signature are consistent with the BVRS representing a more mature oceanic crustal structure, even though the magnetic inversion modelling does not support a polarity reversal origin for the arcuate anomalies (Chapter 7). It is possible that the BVRS was formerly circular in shape but has since been disrupted by younger sea floor spreading. If so then it may mark the site of an old mantle hot spot and up-doming, perhaps associated with some form of triple junction.

At about this time (100 m.y. B.P.) rifting ceased and sea floor spreading commenced between Porcupine Bank-Goban Spur and Orphan Knoll-Flemish Cap (Fig. 8.18b). It is uncertain whether the Charlie-Gibbs F.Z. and Clare Lineament were fully structured in this early spreading stage, or whether the developing ocean tapered to an end west of Porcupine Bank. By the end of Anomaly 34, however, most workers believe there was oceanic accretion west of Lorien Bank, forming what is now Eriador Plain (Fig. 8.18c). Owing to their

Figure 8.18a - e. (opposite). Schematic five-stage evolution of the Rockall Trough. Refer to text for full discussion. Cross-sections in right-hand panels are located by solid lines in left-hand panels. Small arrows = spreading directions; open arrows = sediment influx; crosses = continental crust; triangles = oceanic crust; V = volcanic basement; stipple = sediment; hachured zones = intrusions. 0, 24 & 33 = magnetic anomaly identifications; 1, 2 & 3 = Barra volcanic ridges (BVRS). CAN = Canada; CGFZ = Charlie-Gibbs Fracture Zone; CL = Clare Lineament; GR = Greenland; IR = Republic of Ireland; NF = Newfoundland; OK = Orphan Knoll.

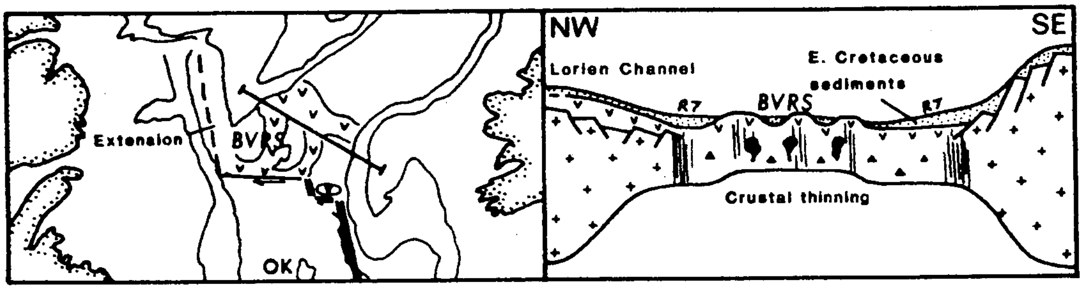
Late Jurassic - Early Cretaceous (c.140 myBP)

A



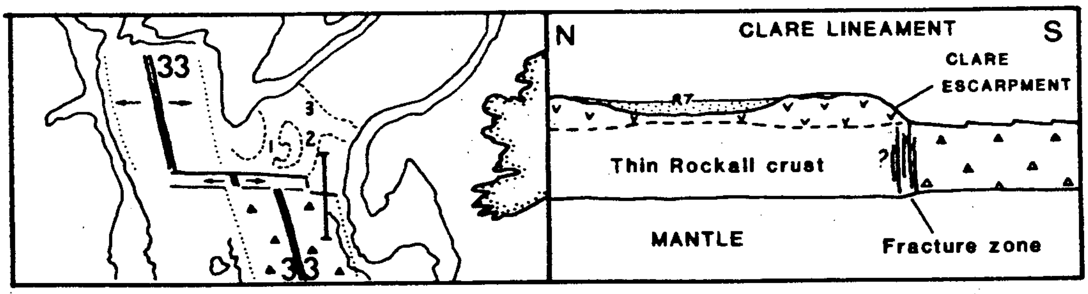
late Early Cretaceous (100 myBP)

B



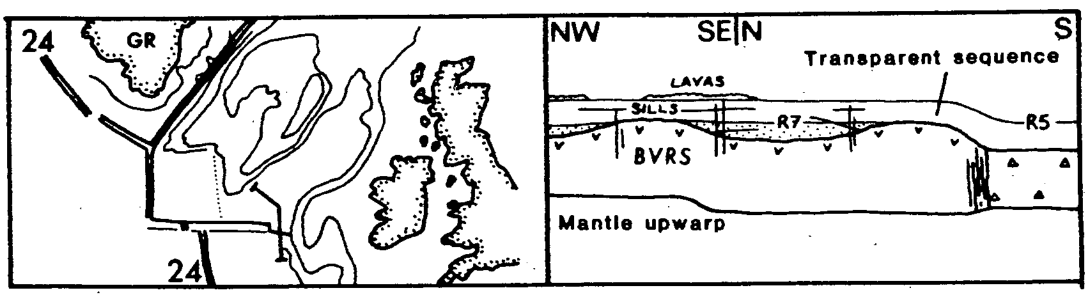
Anomaly 33 time (75-80 myBP)

C



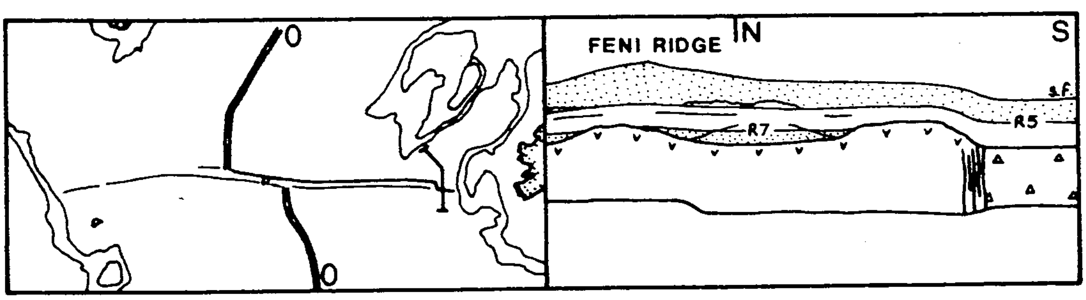
Palaeocene - Eocene, Anomaly 24 time (55 myBP)

D



Present day geometry

E



similar seismic basement and magnetic trends it is very tempting to associate the formation of the BVRS to the progress of the rift tip here along the Clare Lineament and then the Charlie-Gibbs Fracture Zone. Unfortunately the seismic stratigraphy indicates that Early Cretaceous and older sediments were already present over the deep Rockall basement and on the flanks of the BVRS (Figs 8.18b and c). This evidence precludes the formation of the BVRS as a northerly extension of an Anomaly 34 (or younger) median rift.

With the movement of the mid-Atlantic ridge away from the Rockall area sedimentation was able to continue equally on both sides of the CGFZ line (Fig. 8.18d). The conspicuous change in basement character and sediment thickness across this discontinuity should hopefully find an analogy somewhere beneath the East Newfoundland continental margin.

The distinctive R5 reflector is conjectured to have developed at about the Palaeocene-Eocene boundary (c. 55 m.y. B.P.; Fig. 8.18d). The widespread sills and lavas associated with this horizon and the underlying R6 event are almost certainly a manifestation of the onset of sea floor spreading between Greenland and Rockall Plateau-Europe at this time, in common with the extensive Thulean igneous rocks of the Faeroes and NW Britain. The close spatial association between the lavas and sills in southern Rockall Trough and the BVRS is remarkable considering the long lapse in time between their formation, approximately 50 m.y. according to the stratigraphy (compare Figs 8.18b and d). The location of the annular sill complex in southern Rockall trough over the oval mantle upwarp (Chart 7, back pocket) could be interpreted as indicating a lower Tertiary age for the formation of the latter, instead of, or in addition to, the crustal thinning episode proposed in the Early Cretaceous (Fig. 8.18b). However, it is difficult to envisage how crustal accretion 500 km away on the north-west margin of Rockall Plateau could induce up to 2 km of localised crustal thinning in S. Rockall Trough.

The seismic reflection profiles provide evidence for the onset of sedimentary drift construction soon after the formation of reflector R5; that is, from the beginning of the Eocene. This is at variance with the studies of Jones et al. (1970), Roberts (1975) and Masson and Kidd (in press) which agree on a late Eocene-early

Oligocene age for the initial development of Feni Ridge. But seismic profiles across the Gardar Drift on the SE flank of Reykjanes Ridge appear to favour a pre-Oligocene age for differential deposition (Ruddiman 1972; Roberts 1975), an observation supporting the older age predicted from this work. The build up of Feni Ridge drift is generally attributed to the sporadic overflow of cold, sediment-laden Norwegian Sea water across the Wyville-Thomson Ridge (Jones et al. 1970; Roberts 1975). However, it is also apparent that there were several other influences acting during the complex development of the Atlantic Ocean in the Early Tertiary; such as the relative contributions of Norwegian Sea and Labrador Sea waters, the subsidence of the Greenland-Scotland Ridge, and the tectonic developments around the Panama isthmus and the Tethyan oceanic domain (Roberts 1975).

The geophysical and geological investigations of Jones and Ramsay (1982) and, particularly, Roberts et al. (1983) support a Palaeocene age for the Wyville-Thomson Ridge, with its 12 km or less of constituent volcanics being mostly subaerially extruded. It is extremely unlikely that such a huge pile of volcanics could have subsided sufficiently by the beginning of the Eocene to permit overflow of dense Norwegian Sea water into the Rockall Trough. Thus the Eocene build up of Feni Ridge may have been sourced by waters from the south and/or west, e.g. Labrador Sea water. From the early Oligocene an increase in the influence of the Norwegian Sea water would have been made possible by the proposed gradual subsidence of the Wyville-Thomson and Iceland-Faeroe Ridges at about this time (Jones et al. 1970; Roberts 1975; Scrutton and Stow 1984). Obviously a lot more work needs to be carried out relating the regional development of the opening Atlantic and its effect on broad-scale oceanographic circulation, subsidence of ocean barriers, and seismic stratigraphy and sequences before we can fully document the age and evolution of Feni Ridge and other similar sediment drifts.

8.5 SUMMARY

1. Rockall Trough is a small aseismic basin at intermediate depths (2000-4000 m) that separates Rockall Plateau from the continental shelf west of Scotland and Ireland. Gravity anomaly and seismic refraction information provide fairly unambiguous evidence for the existence of unusually thin crust flooring Rockall Trough. In

the southern part of this research area the width of thin crust is about 200 km. In central Rockall Trough it is probable that this width expands to 250 km.

2. Rockall Trough is bordered by classic rifted passive continental margins. Attenuation of continental basement from about 30 km to 7 km is achieved across a horizontal distance of 50 km or more. The two margins are asymmetrical: more abrupt attenuation across the south-east margin of the Trough results in the steeper continental slope of Porcupine Bank, compared with the gentler Rockall Bank. The marked thinning across both margins is achieved partly by downfaulting and rotation of large blocks of continental basement in the upper crust. The regional development of the Northwest European margins provides evidence for rifting in the Rockall area during the upper Palaeozoic, lower Mesozoic and Late Jurassic-Early Cretaceous. The extensive thinning was not fashioned by one major rifting phase. The thin syn-rift and post-rift sedimentary cover to the fault blocks and in the intervening half-grabens implies that the continental margins have been sediment-starved for much of their history. Transfer structures mapped out across the Porcupine Bank permit changes in geometry along the margin and may indicate the existence of fracture zones in the deeper Trough.
3. From an integration of all the geological and geophysical information the 6-8 km thick crust beneath Rockall Trough is interpreted as being of oceanic origin. However, a continental origin for this crust, while not favourable, cannot be entirely excluded. The absence of a median ridge, of axially-facing fault scarps, and of a distinct Layer 2 basement reflector indicate atypical suppressed oceanic accretion. Simple analogies can be made to the diffuse style of crustal accretion recorded in areas such as the Gulf of California (Fig. 8.19) and the Red Sea (Bonatti, 1985).
4. A newly described igneous province, the Barra volcanic ridge system (BVRS), in southern Rockall Trough is related to the last pulses of oceanic accretion. The seismic profiles define at least three conspicuous volcanic ridges which follow a curved path from the Charlie-Gibbs Fracture Zone (CGFZ) onto south-west

Rockall Plateau. Gravity and magnetic modelling in two dimensions indicates that the volcanic ridges are underlain by single, large intrusions or, more likely, by zones of pervasive minor intrusion such as would be expected below a spreading centre. However, inverse magnetic modelling demonstrates that the large magnetic anomalies accompanying the Barra volcanic ridges are not caused by typical polarity reversals. The BVRS may originally have had a circular outline, in which case it could mark the site of an aborted triple junction and mantle hot spot.

5. Reconstructions of the North Atlantic continents back to Anomaly 34 time are in strong support of the contention that the oceanic crust in Rockall Trough is a northerly extension of the Late Cretaceous crust off Goban Spur. However, the seismic stratigraphy outlined in this study, together with the lack of geophysical or geological continuity across the Clare Lineament and CGFZ, indicates an older age. There is no evidence supporting a Carboniferous or Permian Rockall ocean, although this is probably the earliest time when the Rockall-Faeroe rift was established as a locus of tectonism, magmatism and sedimentation. The thickness of older sediments present in the Trough but absent over Late Cretaceous oceanic crust south of the CGFZ favours a Late Jurassic-Early Cretaceous age for Rockall.
6. Detailed mapping of the basement in southern Rockall from seismic profiles shows the Clare Lineament to be a continuation of the long Charlie-Gibbs Fracture Zone, hence proving its persistence right up to the continental margin below Porcupine Bank, where before it was believed to terminate at around 17°W. There is no marked change in the trends of the two structures. The CGFZ has a bearing of 95°-100°, the Clare Lineament a bearing of 100°-105°. Together the two major discontinuities mark the boundary between oceanic crust of different ages to the north and south. They also provide an explanation for the short ridge segment detected recently within the current transform section of the CGFZ at 31.7°W.
7. The geophysical data used here do not resolve any structures in the Trough that could be interpreted as oceanic fracture zones. But certain other pieces of evidence appear to favour a spreading direction of 100°-110°, a range that is compatible with the best

reconstructions of Rockall Plateau presently available. The absence of well marked fracture zones (or even median ridges) within the Rockall basement may be a result of the proposed diffuse style of accretion.

8. A revised seismic stratigraphy for Rockall Trough is presented based on correlations, mainly through multichannel reflection profiles, with several nearby DSDP drill sites. The results of this highlight a seismic sequence overlying basement in Rockall Trough which is often thickly developed but which is absent over mature ocean crust to the south. A pre-Albian age is tentatively given to this distinctive seismic sequence. The stratigraphy indicates differential deposition and consequent upbuilding of Feni Ridge sediment drift from the lower Eocene to the present, rather than from the lower Oligocene. The base of the drift is marked by a conspicuous regional reflector, R5.
9. Interpretation of the c. 8600 km of seismic profiles available to this study proves the existence of widespread igneous reflectors within the sediments filling the Trough. Some of these were previously interpreted as lavas but evidence given here suggests that they are predominantly thin intrusive sills, sourced below by vertical feeder dykes. This igneous episode is dated as roughly lower Eocene or older based both on the seismic stratigraphy and comparisons to the lower Tertiary igneous activity widespread around NW Britain, Faeroes and Greenland. There may be a temporal and spatial association between the crustal thinning postulated below the BVRS and the intrusive/extrusive episode.
10. Within Rockall Trough the largely undeformed sedimentary succession may exceed 5 km of assumed mainly post-Jurassic strata. Slumping, sliding, localised faulting and so on within the sediments appear to be confined to the sections at or near the base of the steep continental margins or along the CGFZ-Clare Lineament. There appears to be little prospect of important hydrocarbon reservoirs in the Trough. The seismic stratigraphy and poorly known sedimentation history do not lend optimism for there being suitable source rocks here. Moreover the reflection profiles do not define any appropriate structural traps or

closures in the Trough. The older rifting history of the continental margins, with their faulted half-grabens and more varied sedimentation pattern, may prove to hold more potential for hydrocarbon exploration.

8.6 Epilogue

From this study it is shown that Rockall Trough is floored by upwards of 200 km of abnormal perhaps transitional oceanic crust and that it has subsequently been mantled by a thick cover of drift-phase sediments. In this light it is interesting to compare Rockall Trough with the Gulfs of California and Aden (Fig. 8.19). In the first example the tectonic style, strike-slip dominated motion, is very different to that in Rockall but the end product appears to be similar - crustal accretion suppressed or diffused by heavy sedimentation. In the second example the geometry of spreading is closely comparable, but in the Gulf of Aden numerous workers have reported mature, unequivocal oceanic crust. Nevertheless the analogy between Rockall and the Gulf of Aden is striking in terms of their shape, size and the oblique-to-margin spreading (Fig. 8.19). The analogy would be enhanced if the Gulf were to cease spreading now and receive sediments for at least the next 130 million years.

Thus as a long inactive, heavily sedimented ocean basin, floored as it is with enigmatic oceanic crust that seemingly did not have time to develop into a mature orthogonal spreading system, the Rockall Trough occupies a remarkable and unique status in the plate tectonic scheme. While there are other ocean basins that have also become inactive and later covered with sediments, for example the Labrador Sea and Bay of Biscay, these areas are younger and/or are typified by a normal oceanic crustal structure. As such this author believes that the Rockall-Faeroe Channel represents an interesting target for deep drilling. The Trough appears to have halted its evolution somewhere between the stages represented by the Red Sea, where the pattern of oceanic accretion is likewise unordered, and the Gulf of Aden where a near-orthogonal arrangement of ridges and transforms is developing (Fig. 8.19).

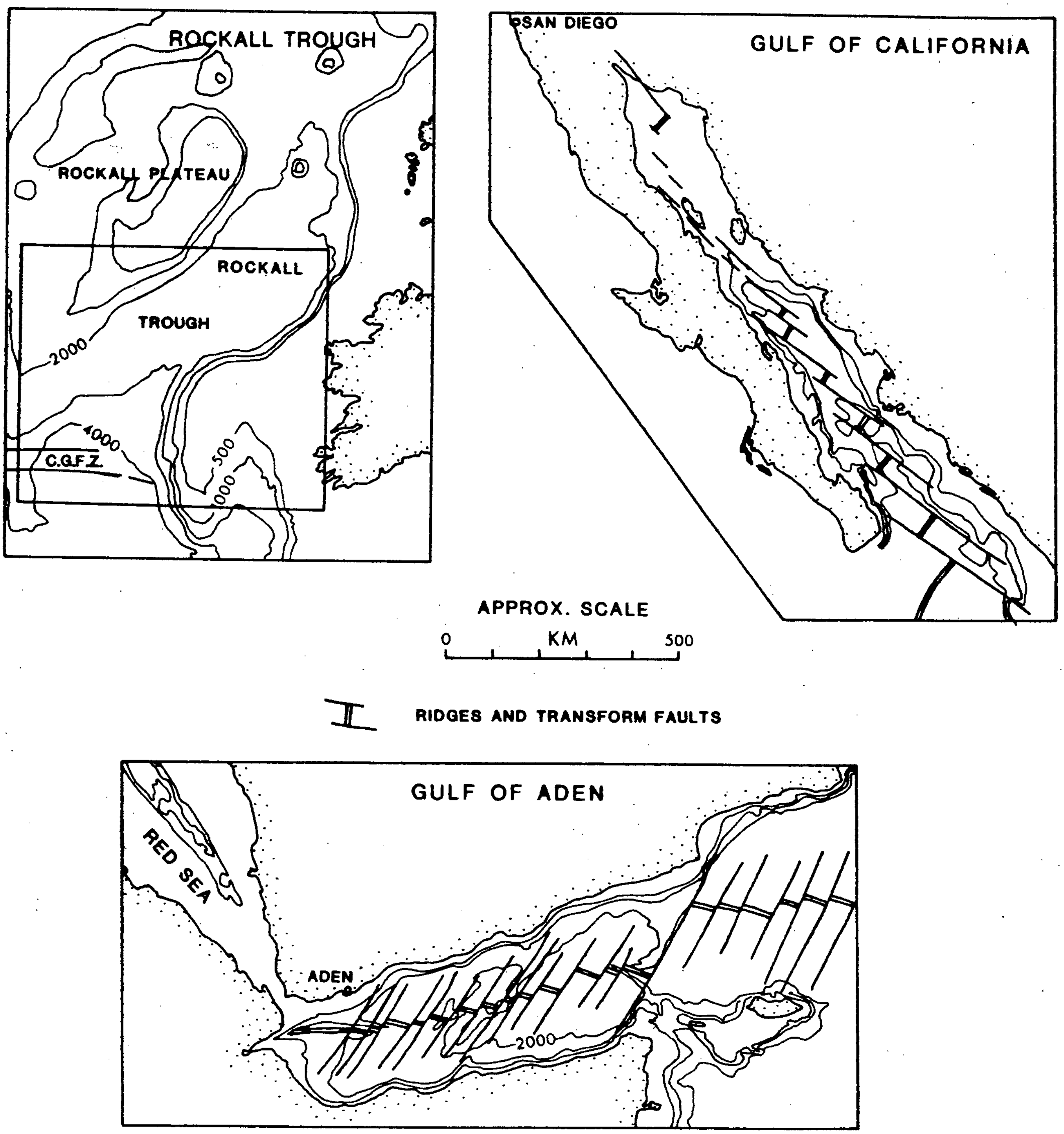


Figure 8.19 Scale comparison between the Rockall Trough, Gulf of California, North America and Gulf of Aden, NE Africa. All figures drawn approximately to the same scale. CGFZ = Charlie-Gibbs Fracture Zone. Oceanic tectonic elements: double solid line = spreading ridges; single solid line = fracture zones. Bathymetry in metres. Inset box in top left panel is the region covered by this research study.

Rockall Trough remains the most poorly understood large basin around the UK. For this situation to improve it will be necessary to integrate further detailed geophysical investigations with drilling programmes. State-of-the-art seismic reflection and refraction studies, combined with gravity and magnetic data, should be able to provide more tightly constrained crustal models that can be interpreted with less ambiguity. To achieve this the author recommends that any seismic studies are observed across the Trough in the region north-east of the BVRS and north of Porcupine Bank. Elsewhere in the Trough it is anticipated that the widespread Tertiary lavas and sills within the sediments will serve only to distort the results and any interpretations from them. The window proposed above is largely free of these superficial events so it should be possible to obtain a clearer picture of the deeper thin crust.

There is a strong need to define the chronostratigraphy and lithostratigraphy in the Rockall basin much more accurately. This can hopefully be achieved through the efforts of the Ocean Drilling Project or its successors, and there is further cause for optimism in the steady expansion of the hydrocarbon exploration industry into these deeper waters. Until then ...

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