MARINE GEOLOGY OF THE ROCKALL PLATEAU AND TROUGH

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The Rockall Plateau is an extensive shallow water area located south of Iceland and west of the British Isles: it is separated from the British Isles by the 3000 m deep Rockall Trough. Rockall Island, composed of 52 ± 9 Ma aegirine-granite is the sole subaerial expression. The Rockall Plateau is interpreted as a continental fragment or microcontinent isolated during the sea floor spreading evolution of the North Atlantic Ocean.

A geological reconnaissance of the Rockall Plateau and Trough has been made by using a 650 cm^3 (40 in^3) seismic reflexion profiling system, supplemented by sparker (8 kJ) profiles on Rockall Bank and arcer (60 kJ) profiles across the margin west of the British Isles. Stratigraphic interpretation of these profiles has been aided by deep sea drilling data, bottom sampling on Rockall Bank and by the relation between the various reflecting horizons and oceanic basement dated by oceanic magnetic anomaly identifications. Analysis of the microtopography of the area has given information on Post-Palaeogene sedimentation processes.

Three major sedimentary basins are present in the area. The Hatton–Rockall Basin is developed in thinned continental crust on Rockall Plateau. The Rockall Trough is developed on continental crust and includes oceanic crust believed to have been generated in Late Jurassic–Early Cretaceous time. The Porcupine Seabight may be developed on thinned continental crust. All three basins have a faulted basement and exhibit a history of progressive and/or intermittent subsidence. The subsidence phases correlate closely with estimated changes in sea-floor spreading rate. This correlation and the regional pattern of uplift and subsidence is discussed with reference to the effects of thermal subsidence and differential loading of the lithosphere beneath continental margins.

Post Upper Eocene sedimentation throughout the area was characterized initially by widespread chert deposition and subsequently by differential deposition of Early Miocene to Recent oozes. The onset of widespread differential deposition in the Early Miocene indicates the present near bottom-circulation was established at this time and may be related to subsidence of the Iceland–Faeroes Ridge. The relation between differential deposition, topography and circulation is discussed in terms of flow around obstacles.

1. Introduction

The Rockall Plateau is an extensive shallow area located south of Iceland and west of the continental margin of the British Isles from which it is topographically isolated by the Rockall Trough: to the west the Plateau is bounded by the flank provinces of the Reykjanes Ridge (figure 1).

The Rockall Plateau is the only major microcontinent known in the North Atlantic Ocean:

its geology and geological evolution are relevant to the study of the geological and physical processes that form a microcontinent and its continental margins; its geology is also relevant to the pre-drift continuity of the North Atlantic continents.

Much geophysical data pertinent to the structure of the Rockall Plateau has accumulated since Bullard, Everett & Smith (1965) postulated that it might be a microcontinent. Deep seismic refraction lines (figures 1 and 2), give continental seismic velocities, and mantle depths of 33 and 22 km beneath the Rockall Bank and the Hatton–Rockall Basin respectively (Scrutton 1970; Scrutton & Roberts 1971; Scrutton 1972). Gravity models of the Plateau and Trough (figure 3*a*; Scrutton 1972) have confirmed the earlier suggestions of Bullard *et al.* (1965) and Roberts (1970, 1971) that the whole Plateau is a microcontinent. The microcontinent may, on recent evidence, extend further north to underlie the lavas of the Faeroe Islands (Bott,

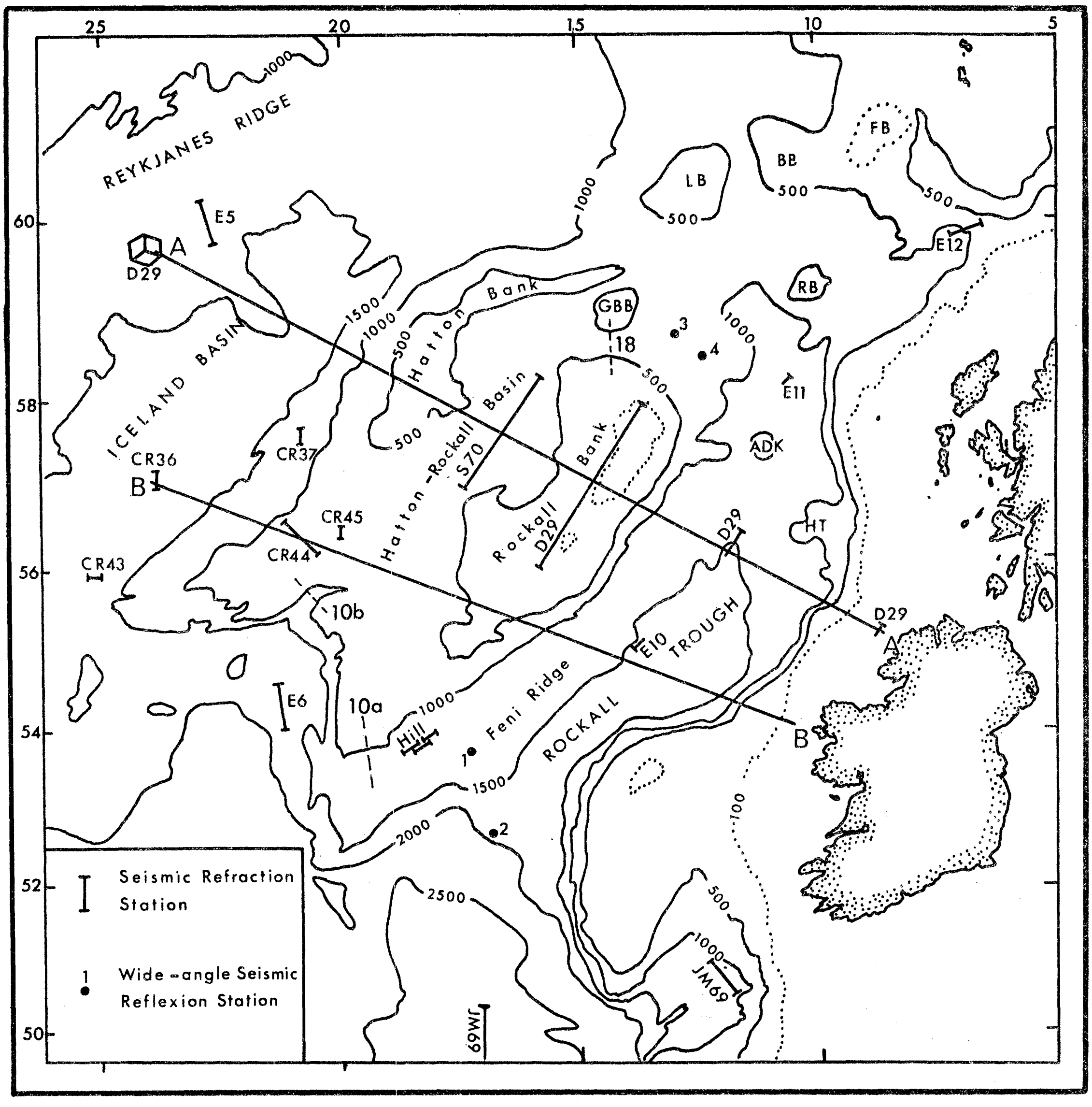


FIGURE 1. Generalized bathymetry of the Rockall Plateau showing the location of the seismic refraction profiles of figure 2, seismic refraction and wide angle reflexion stations. Pecked lines indicate seismic reflexion profiles shown in figures 28*a*, *b* and 24. (Contours in fathoms.)

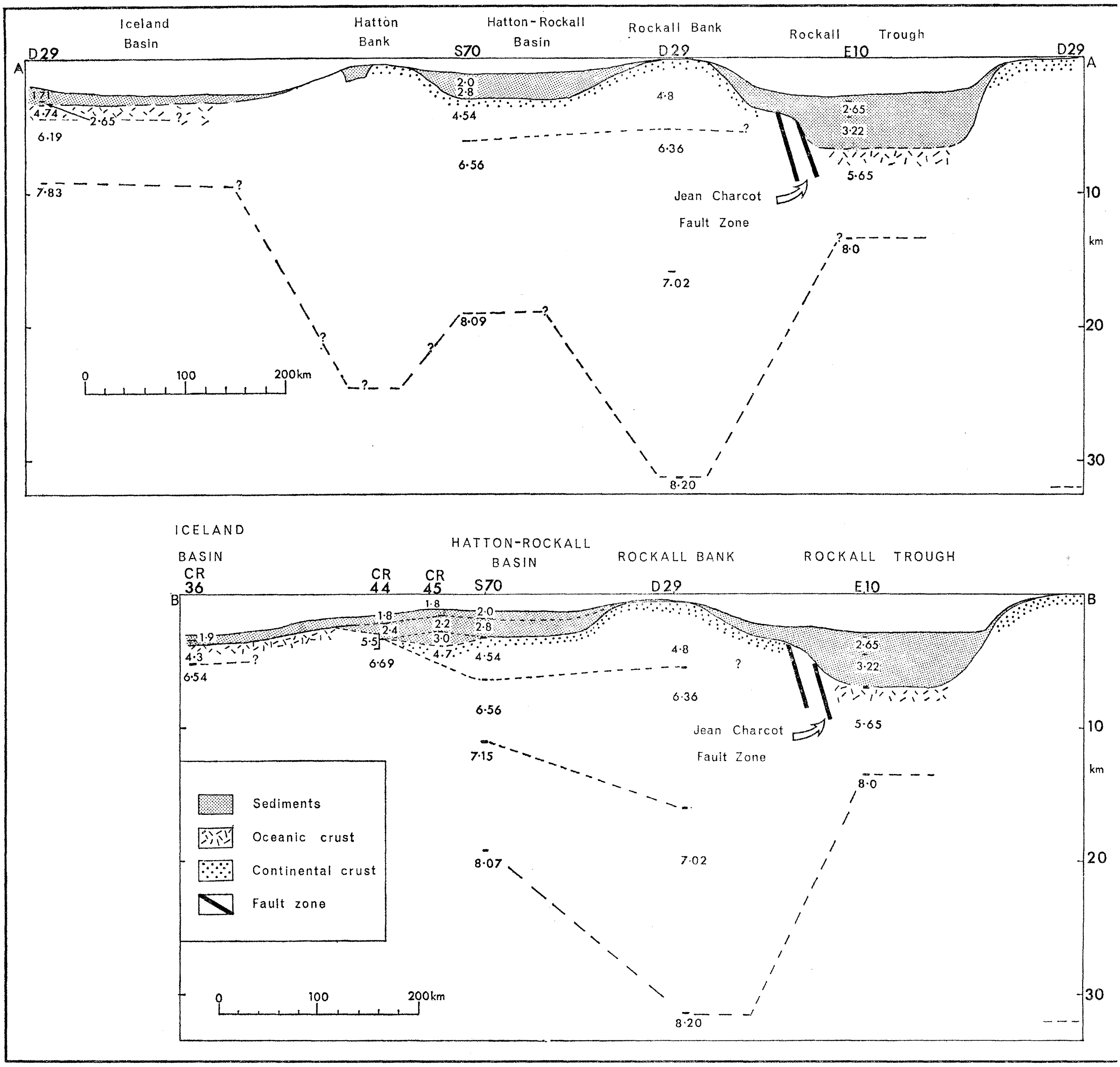


FIGURE 2. Structure of the Rockall Plateau from seismic refraction data; stations and profiles are located in figure 1 (refraction data from Gaskell, Hill & Swallow 1958; Ewing & Ewing 1958; Scrutton 1972).

Browitt & Stacey 1971; Casten 1973; Bott *et al.* 1974). Laxfordian and Grenvillian granulites dredged and drilled on the Rockall Bank (Roberts, Matthews & Eden 1972; Roberts, Ardus & Dearnley 1973*a*; Miller, Roberts & Matthews 1973) have proved continental rocks as earlier indicated from indirect geochemical evidence (Moorbath & Welke 1969). West of the Rockall Plateau (figure 3*b*), the identification of magnetic anomaly 24 has shown the smooth oceanic basement abutting the west margin is 60 Ma in age (Ewing & Ewing 1958; Godby, Hood & Bower

1968; Vogt *et al.* 1969, 1971; Scrutton & Roberts 1971; Ruddiman 1972; Laughton 1971; Vogt & Avery 1974). The age of the oceanic basement that abuts the southwest margin (Ewing & Ewing 1958) is given as 76 Ma by the position there of anomaly 32 (Vogt *et al.* 1971; Laughton 1971; Vogt & Avery 1974; Jones & Roberts 1975). Early Cretaceous oceanic crust probably underlies the Rockall Trough (Roberts 1971; Scrutton & Roberts 1971; Roberts 1973; Roberts 1974; Jones & Roberts, 1975).

Syntheses of the plate tectonic evolution of the North Atlantic Ocean based on those data demonstrate that three distinct phases of sea floor spreading hived off and structurally isolated the Rockall Plateau (Vogt *et al.* 1971; Le Pichon, Hyndman & Pautot 1972; Laughton 1971). The earliest phase contemporaneously opened the Rockall Trough and Bay of Biscay during Mid-Early Cretaceous time (Roberts 1971; Roberts 1973*a*; Roberts, Flemming, Harrison & Binns 1973*b*; Jones & Roberts, 1975) and thereby split the Greenland–North America–Rockall plate off Eurasia. By 76 Ma, spreading had ceased in the Rockall Trough and the spreading axis changed to the line of the Labrador Sea spreading apart North America and Greenland–Rockall. At 60 Ma, spreading about the incipient Reykjanes Ridge split Greenland and Rockall apart thereby completing the isolation of the Rockall Plateau.

The outline spreading history thus shows the continental margins of the Rockall Plateau formed in three distinct spreading or rifting events. It is therefore probable that the Rockall Plateau has consequently had a substantially more complex stratigraphic and tectonic history than that conceptually implied for continental margins formed by spreading and rifting about a single axis. In such concepts, it is envisaged (see, for example, Schneider 1973), that the continental lithosphere is initially domed by sublithospheric heating, and then rifted apart (with extensive volcanism) until oceanic lithosphere is formed; the later history of the young margin and the oceanic part of accreting plate is governed mainly by erosion, thermal subsidence and epeirogenesis, characterized by regional warping and the development of a thick sedimentary wedge. In real terms, the pattern of early rifting and fracture zone development determines the basic structural framework of a continental margin but is itself controlled by the interaction between the regional stresses imposed during rifting and both the pre-existing inhomogeneities and tectonic fabric of the lithosphere. In the case of a microcontinent such as the Rockall Plateau, we may surmise that these factors will significantly contribute to the basic form of its margins although normal fault fabrics developed during and in response to preceding spreading episodes may further influence the ultimate form of the younger margins and the later history of the older margins. Two other factors may be anticipated to have an important influence on the structure and evolution of the microcontinent; many normal faults may be rejuvenated in response to the processes of hot creep and thermal subsidence inferred to operate beneath continental margins (Bott 1971; Sleep 1971; Bott & Dean 1972) and secondly regional warping, subsidence and epeirogenesis of both the oceanic and continental parts of the plate may occur in response to changes in the rate of spreading at the mid-ocean ridge axis in turn influencing the stratigraphy and distribution of sediments.

We may therefore anticipate that such processes have fundamentally influenced the geology and geological history of the Rockall Plateau. Their importance and interdependence can be best evaluated by examining the structural and stratigraphic history of the Rockall Plateau revealed in seismic reflexion profiles, and in relation to the spreading history of the adjacent oceanic crust.

This report is mainly concerned with a discussion of the topography, structural and stratigraphic history of the Rockall Plateau region and the extent to which the factors outlined above have determined the present geology of the Rockall Plateau, Rockall Trough and continental margin west of the British Isles. A detailed description of several aspects of the microtopography has been given because of the close relation to the ocean bottom circulation that has profoundly influenced sedimentation during the Neogene. Many of the minutiae that would contribute to a complete perspective of the development of Rockall Plateau have been omitted from this report and will be published elsewhere.

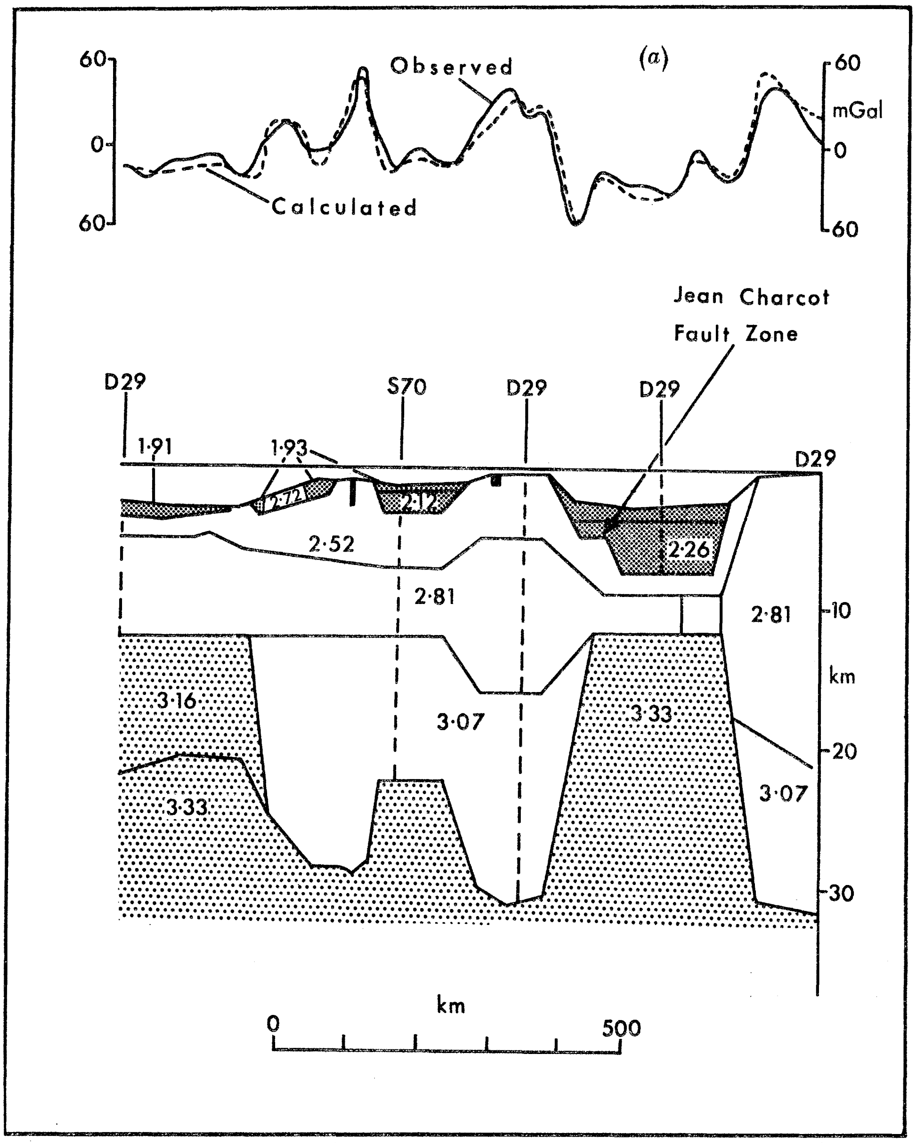


FIGURE 3(*a*). For description see opposite.

2. TOPOGRAPHY OF THE ROCKALL PLATEAU AND ADJACENT AREAS

(*a*) *Data analysis*

The bathymetry of the Rockall Plateau and northeast Atlantic was first outlined by Hill (1956) using rather sparse echo-sounder data. Subsequently numerous soundings have been taken by research ships and collected by the Hydrographic Department (M.O.D.) of the United Kingdom, the Deutsches Hydrographisches Institut, the Lamont–Doherty Geological Observatory, the Institute of Oceanographic Sciences (formerly the National Institute of Oceanography, U.K.) and Durham University. In addition, the southernmost parts of the Rockall Trough, Rockall Plateau and the eastern flanks of the Reykjanes Ridge have been surveyed at 5.5 km (3 n mile) line spacing by the U.S. Naval Oceanographic Office (Johnson & Schneider 1969; Johnson, Vogt & Schneider 1971); Rockall Bank and Rosemary Bank (figure 40) have been bathymetrically surveyed in detail by H.M.S. *Hecla* and H.M.S. *Hydra* (Hydrographer of the Navy 1969, Roberts & Jones 1974).

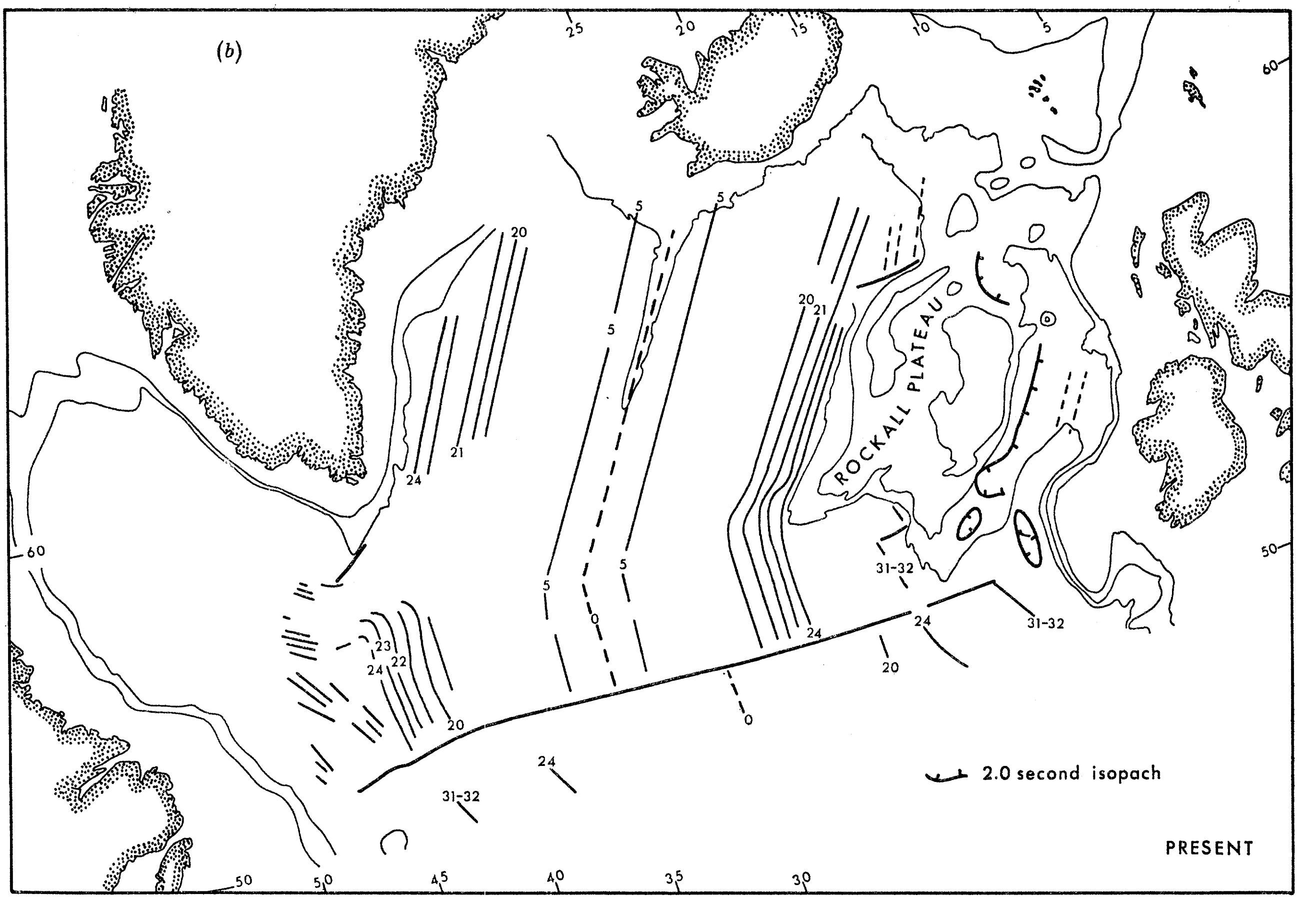


FIGURE 3. (*a*) Structure of the Rockall Plateau modelled from gravity and available refraction data (redrawn from Scrutton 1972). Coarse stipple, mantle; fine stipple, sediment. Densities shown by large numerals. See figure 2 for geographical names. (*b*) Magnetic anomalies on the oceanic crust around Rockall Plateau (data from Godby *et al.* 1968; Laughton 1971; Vogt & Avery 1974; Jones & Roberts 1975).

The bathymetric contour map of the Rockall Plateau (figure 40) was compiled initially at a 1:1000000 scale in two ways; in the case of available detailed surveys, the contours were drawn wherever possible at 5 fathoms (9.1 m) (Rockall Bank) and 50 fathoms (91 m) intervals (Rockall Trough) using the recorded depths plotted at the original survey scale. Otherwise the contours were accepted from published charts if the original survey data were not available (see, for example, Fig. 3 of Johnson & Schneider 1969). The larger scale contour charts were then reduced to a 1:1000000 scale. Beyond the areas surveyed in detail, the chart has been compiled using soundings collected on a 1:1000000 scale and contoured in the manner described by Laughton, Roberts & Graves (1973). The 1:1000000 chart has been subsequently redrawn at a 1:2000000 scale for this report (figure 40).

There is an obvious first order structural relation between the Rockall Plateau and the neighbouring sea floor that permits discussion in terms of three topographically distinct units. The Rockall Plateau and its margins, the ocean basin floor west and southwest of Rockall Plateau and the Rockall Trough.

(*b*) *The Rockall Plateau: Rockall and other banks*

The Rockall Plateau is an extensive shallow area measuring some 400 n mile × 240 n mile (22×10^4 km^2) at the 1000 fathom (1828 m) line and consists of a series of relatively shallow banks disposed around the deeper water Hatton–Rockall Basin (figure 1; figure 40). It is bounded to the west, east and southwest by steep margins falling to depths of 1000–1500 fathoms (1828–2743 m) although to the north the margins are ill-defined and merge with the footslopes of Lousy Bank seamount. Rockall Island is the sole subaerial expression of Rockall Plateau. The shallowest banks are Rockall Bank, Hatton Bank and George Bligh Bank. A number of deeper banks here called Lorien, Fangorn and Edoras Banks (after Tolkien (1959)) mark the southern and westernmost elevations of the Plateau.

The bathymetry of Rockall Bank based on detailed surveys by H.M.S. *Hecla* is shown in figure 4 (Roberts & Jones 1974; Chart C 6091, Hydrographer of the Navy 1975). Rockall Island is the subaerial portion of a series of curvilinear ridges that mark the eroded intrusive centre (Roberts 1969) and collectively form an uneven, possibly wave-cut platform at a depth of *ca.* 60 fathoms (110 m) flanked by slightly steeper talus slopes (figure 4). Beyond Rockall Island, the rocky shoals have steep margins with a characteristic break in slope at their base in depths of 95–100 fathoms (173–183 m) (figure 4). A Flandrian beach conglomerate dredged in this depth at the base of Empress of Britain Bank (56° 15.3′ N 15° 12.0′ W) suggests it may mark subaerial erosion during a Flandrian sea level stance (Roberts *et al.* 1972). The absence of comparable features in depths greater than 100 fathoms (183 m) suggests this depth may mark the lowest sea level stance.

Between rock outcrops and the shelf break, side-scan sonar and echo-sounder traverses, calibrated by bottom sampling and photography (Roberts & Wilson 1971; Roberts 1972*b*; Roberts *et al.* 1972; Roberts & Eden 1973), show a gradual transition from rock outcrop to low rock ridges or boulder fields partly covered in coarse carbonate sand to an almost complete fine carbonate sand cover marked only by dark isolated and rather diffuse echoes on the outer parts of the Bank (figure 5). Deep sea drilling and coring shows the carbonate sand facies is replaced by *Globigerina* ooze by 500 fathoms (914 m) water depths to the west (Roberts 1972*a*; hole 117 in Laughton, Berggren *et al.* 1972). Photographic evidence suggests the carbonate sand is gradually burying the rock outcrops and beach conglomerate (Roberts *et al.* 1972). The lithologic pattern may therefore reflect post-Flandrian transgressive sedimentation. The origin of the dark echoes is not clear; they are not outcrops (Roberts 1972*b*) and their depth clearly precludes an origin by iceberg ploughing. They may be echoes from patch reefs of the cold water coral *Lophelia porifera* recently observed and sampled on Rockall Bank (Roberts & Eden 1973). The side-scan sonar data do not show evidence of a volumetrically significant recent sediment transport regime. Cartwright (private communication) using data from current meters moored near Rockall has derived maximum semi-diurnal current speeds of 35.5 cm s^{-1}

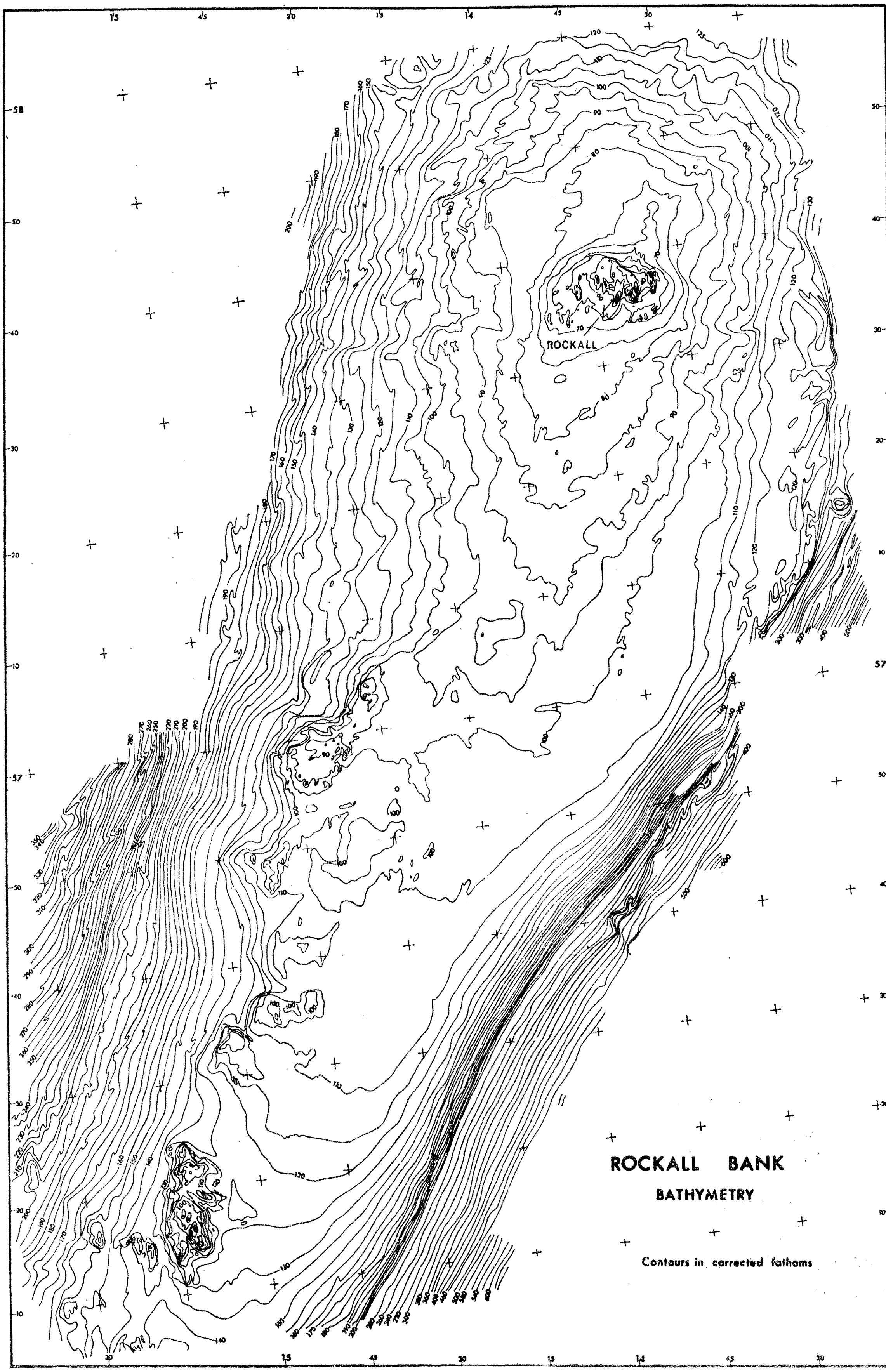


FIGURE 4. Bathymetry of the Rockall Bank. Contours in corrected fathoms at a 5 fathom interval except on the east margin where a 20 fathom interval has been used in depths greater than 200 fathoms (Hydrographer of the Navy 1975).

(136° direction) and diurnal tidal current speeds of 9.1 cm s^{-1} (046° direction). These velocities are clearly inadequate to transport the coarser sands and gravels (Heezen & Hollister 1964) that are present but may suspend and winnow the finer grades. These sediments and others suspended by wave action may be further distributed by the anticyclonic circulation pattern inferred by Hill (1971). It seems clear that, unlike the shelf west of the British Isles, the Rockall Bank contributes little sediment to the adjoining deep basins in accord with the absence of canyons.

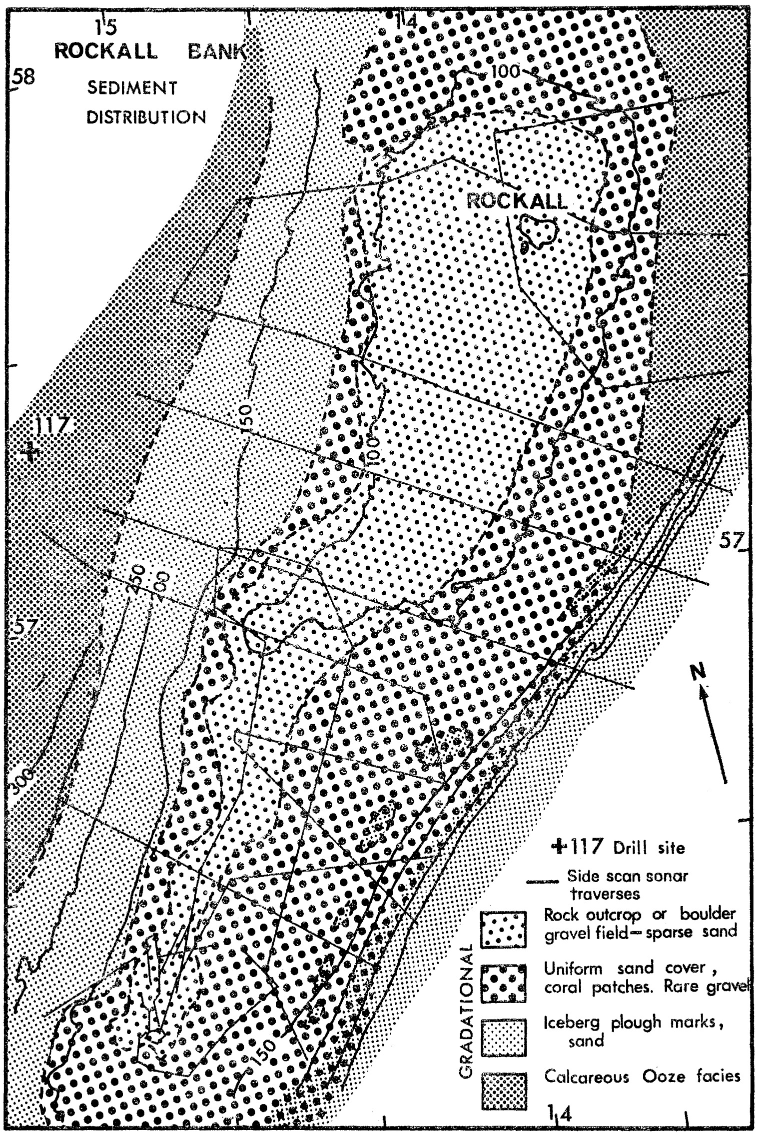


FIGURE 5. Bed form patterns derived from side-scan sonar surveys. Sonar traverses shown as heavy lines. Echo-sounder traverses by H.M.S. *Hecla* omitted for clarity (see Roberts & Jones 1975). (Contours in fathoms.)

In 100–300 fathoms (183–548 m) depths to the east and west (figure 5), there is an unusual relief consisting of a series of troughs separated by often strongly reverberant ridges with a relief of *ca.* 5 fathoms (9 m) and typical wavelength of 0.5 km. Comparable features produced by grounded ice-floes in the Bering Sea (Shearer, Macnab, Pelletier & Smith 1971) led Belder-

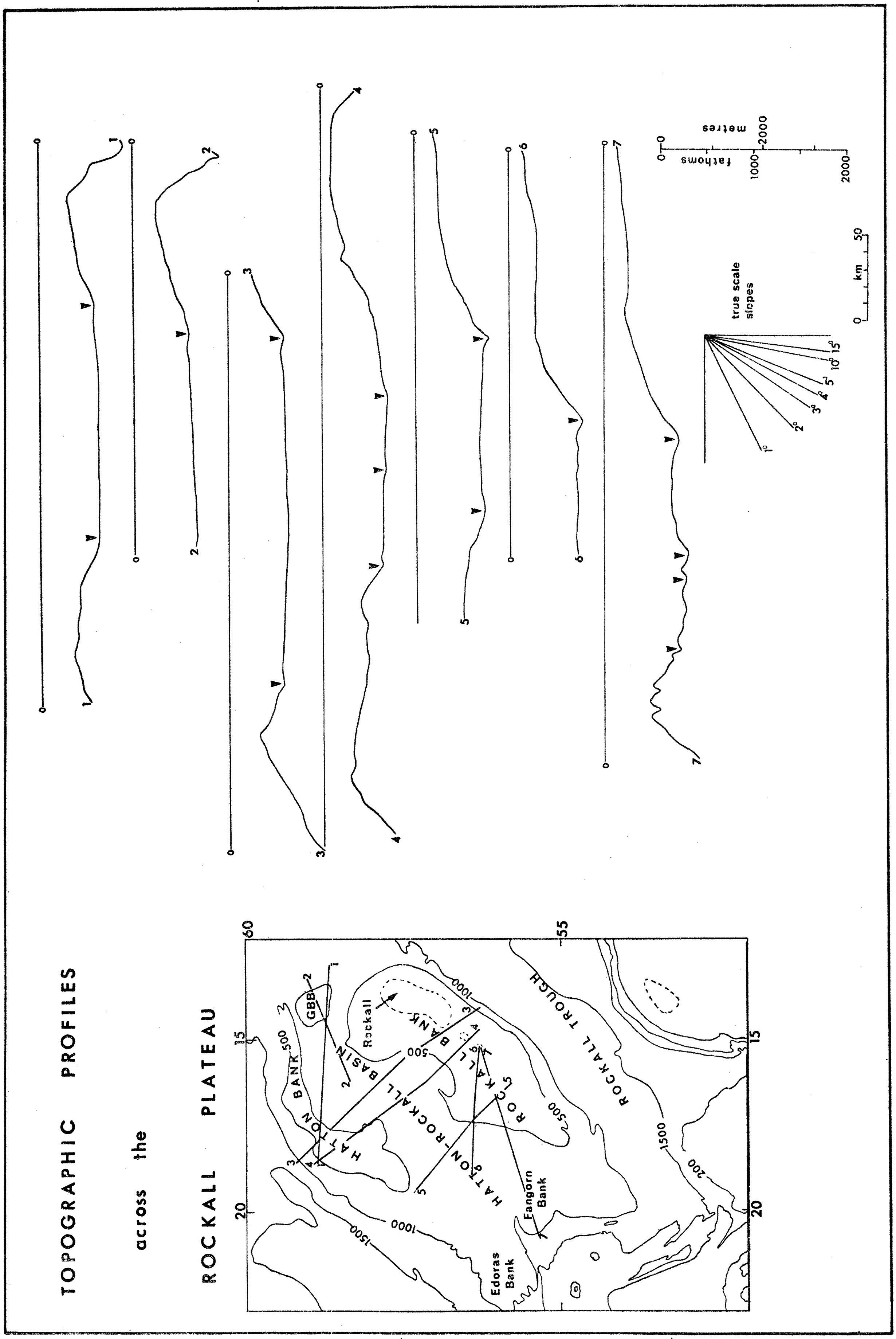


Figure 6. Topographic profiles across the Hatton–Rockall Basin. Inverted arrows indicate axes of marginal channels.

son, Kenyon & Wilson (1973) to postulate a relict origin due to icebergs grounding at a time of lower sea level. In support, an analysis of the echo-sounder records taken by H.M.S. *Hecla* shows the relief entirely surrounds the Bank but is everywhere deeper than the 95 fathom (173 m) fall in sea level indicated by the Flandrian beach conglomerate; quantities of glacial erratics have also been dredged from them (Roberts & Wilson 1971; Roberts 1972*b*).

George Bligh Bank (figure 1; figure 6, profiles 1 and 2; figure 40) has a minimum depth of 234 fathoms (428 m) and is entirely separated from Rockall Bank by a steep sided trough that deepens eastward and connects the Hatton–Rockall Basin and Rockall Trough. The thin sediment cover is formed into a 2–3 fathom (4–5.5 m) relief with a wavelength of between 100 and 200 m. The relief is similar to the iceberg plough marks around Rockall Bank and may indicate grounding on George Bligh Bank.

Hatton Bank has a minimum observed depth of 259 fathoms (473 m) and is a linear ENE–WSW trending ridge characterized by isolated pinnacles with a relief of 5–6 fathoms (9–11 m). It is separated from George Bligh Bank by a linear though narrower trough with a sill depth of 580 fathoms (1060 m).

Fangorn Bank (figure 6, profile 7; figure 40) consists of a series of irregular, steep sided peaks with a relief of *ca.* 40 fathoms (73 m) in contrast to the smooth convex slopes of Lorien and Edoras Banks (figure 28*b*). In these cases, the convexity is associated with a sharp break in slope at the edge of the Banks marking an area where sediments are typically thin or absent compared to the crest and margins of the Bank (figure 28*b*).

(*c*) *The Hatton–Rockall Basin*

The Hatton–Rockall Basin is an irregularly shaped basin elongated in a NE–SW direction, measuring some 240 n mile × 100 n mile (444 km × 185 km) and almost completely enclosed in the north by Hatton Bank, George Bligh Bank and Rockall Bank (figures 1 and 40). There is an obvious spatial relation between the trends of the basin margins and the changes in trend of the west and southwest margins that implies a substantial structural control on the form of the basin.

Although the basin conforms to the accepted topographic definition, it has a number of unusual features. The basin floor is formed into two broad spurs or arches so that it is shallower than the margins by as much as 54 fathoms (98 m) (figures 6 and 40). The spurs are associated with well developed marginal channels (figures 6 and 9) that continue into the northern part of the Basin where the arching is absent. There is considerable variation in the form, width and relief of these channels. In general the channels are broad and have low relief where flanked by gentle basin margin slopes and are narrowest with greater relief where the latter are steepest (figure 6, profiles 1, 2 and 5). However, the presence of channels within the centre of the basin suggests proximity to a slope is not a prerequisite for their development (figure 6, profile 7). These channels differ markedly from the steep sided and square profiles of erosional channels such as the Maury channel (Ruddiman 1972). Seismic profiles given later show they are due to differential deposition.

The microtopography of the channels and adjacent basin floor also shows considerable variation that is pertinent to Neogene sedimentation in the area. The basin floor beyond the channels and flanking ridges is characterized by a series of narrow depressions and wider ridges with an average relief of 10 fathoms (18 m) and wavelength of 0.75–1.0 n mile (1.38–1.85 km) that form an en-echelon pattern trending NE–SW and sub-parallel to the axis of the arch.

Towards Fangorn Bank, these features increase in amplitude and wavelength to form the annular system of ridges and channels around this Bank (figure 6, profile 7; figures 9 and 40). The crest of the ridge flanking the channels is a relatively sharp boundary between the undulations of the basin floor and the smooth relief of the outer wall of the channel. The channel floor also tends to be more acoustically reverberant than the margins suggesting that minor irregularities not resolvable with the echo-sounder are scattering the sound. There is clearly a close spatial relation in the distribution and development of these bedforms. Comparable features are associated with differential sediment deposition (see, for example, Schneider *et al.* 1967; Fox, Heezen & Harian 1968; Le Pichon, Eittreim & Ludwig 1971*a*; Le Pichon, Eittreim & Ewing 1971*b*) and indicate the present morphology is being fashioned by such sedimentary processes.

(*d*) *The west margin of the Rockall Plateau*

The western continental margin of the Rockall Plateau extends from 60° N to 54° N and its base is clearly defined by the 1500 fathom contour (figures 7 and 40). North of 60° N, the margin merges with the footslopes of the Iceland–Faeroes Rise. It is not a typical continental margin because both shelf and shelf break are absent. The gross physiography consists of a series of poorly defined shelf-like features that are the crests of Hatton, Edoras, Fangorn and Lorien Banks and a 'continental' slope (figure 7, profiles 1–12; figure 40).

Major regional changes in the trend and morphology at 59° 30′ N and 55° 30′ N closely reflect the different ages and tectonic processes that have formed the margin between 55° 30′ N and 59° 30′ N. The margin (figure 3*b*) closely parallels the oldest, 60 Ma, magnetic anomaly 24 recorded in the adjacent oceanic crust (Vogt & Avery 1974). The eastward change in trend and slope at 59° 30′ N may be related to a fracture zone inferred to lie north of Hatton Bank (figure 3*b*; Jones & Roberts 1975). In contrast, the margin south of 55° 30′ N is remarkably rectilinear in plan. The adjacent oceanic magnetic anomalies (figure 3*b*) are truncated at the steep east–west slope south of Edoras Bank (figure 40; figure 7, profiles 9 and 10) and young westward from anomaly 32 close to Lorien Bank to anomaly 24 at 25° 30′ W (Vogt & Avery 1974; Jones & Roberts 1975). The truncated anomalies suggest the east–west margin may have been a fracture zone between 76 and 60 Ma in contrast to the north–south margin, parallel to the magnetic anomalies and thus presumably formed by rifting. Clearly the rectilinear margin is older than the 60 Ma margin north of 55° 30′ N. Le Pichon *et al.* (1972) and Laughton (1971) have previously suggested the rectilinearity marks an offset in the margin of the embryo Labrador Sea that may have been controlled by the eastward continuation of the Grenville Front (the structural boundary between the Grenville and Churchill geological provinces of NE Canada) from Labrador on to the Rockall Plateau (Roberts *et al.* 1973*a*). Finally, the progressive southward increase in depth of the western margin may reflect greater thermal subsidence of the older and cooler part of the margin (Sleep 1971). More probably however, it may reflect the greater extension and therefore subsidence undergone by the southwest margin during the 60 and 76 Ma rifting episodes.

(*e*) *The ocean basin floor west and southwest of Rockall Plateau*

The topography of this area can be reviewed in terms of two provinces divided naturally by the 60 Ma isochron and bounded to the south by the Gibbs Fracture Zone (figure 40). These provinces are the east flanks of the Reykjanes Ridge (post 60 Ma crust) and the ocean

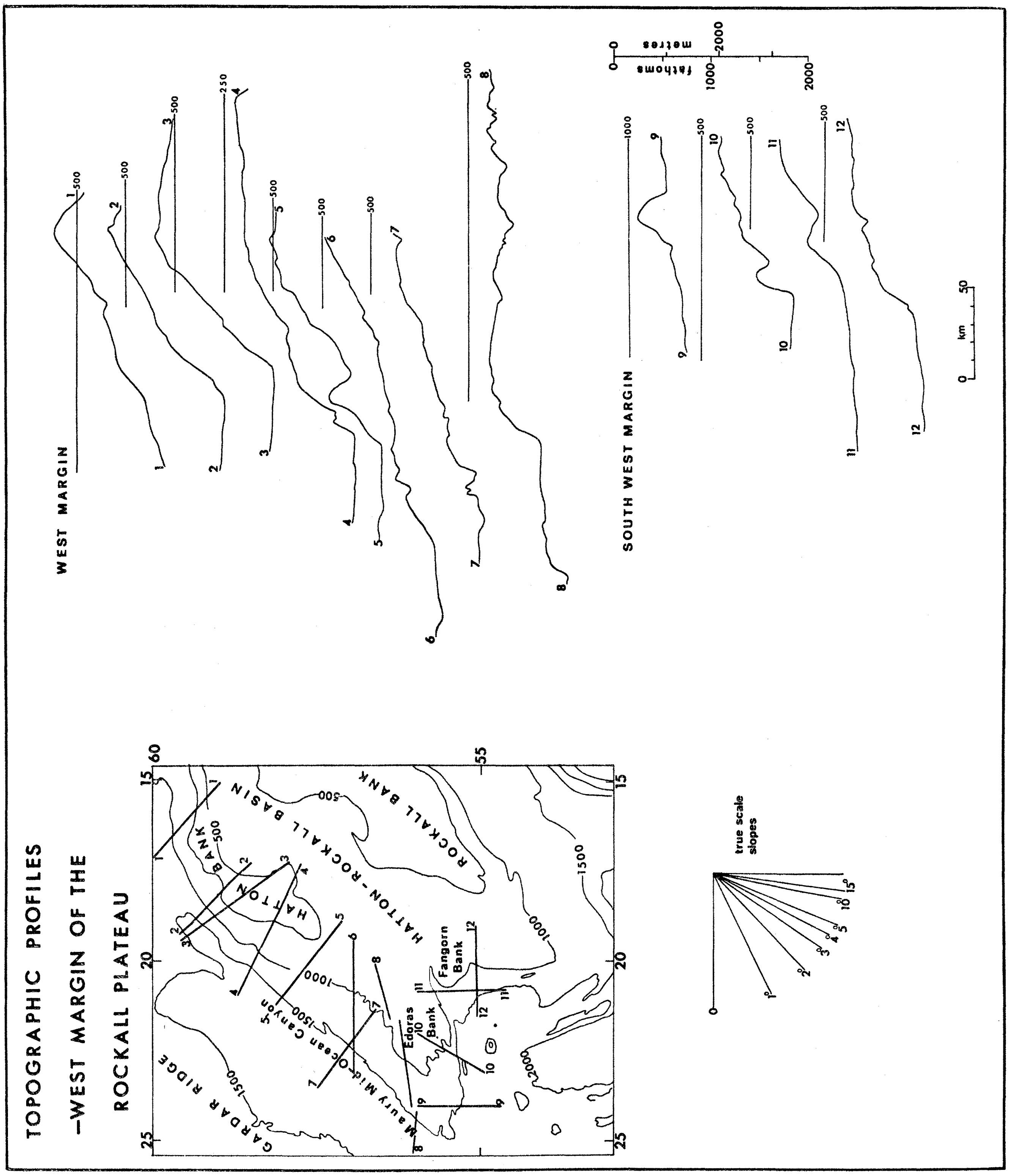


FIGURE 7. Topographic profiles across the west margin of Rockall Plateau.

basin floor southwest of Rockall Plateau (60–76 Ma crust). The topography of the area south of the Gibbs Fracture Zone has been discussed recently by Johnson & Vogt (1973).

(i) *The Gardar Ridge*

The eastern flank of the Reykjanes Ridge is dominated by the broad Gardar Ridge (figure 40) sediment drift and is therefore anomalously smooth in contrast to the well lineated and more typical mid-ocean ridge topography developed south of the Gibbs Fracture Zone (Johnson & Schneider 1969; Johnson *et al.* 1971; Johnson & Vogt 1973). The difference reflects the contrasting sedimentation regimes north and south of the Gibbs Fracture Zone. To the north the well lineated basement relief has been infilled and covered by Gardar Ridge sediments deposited under the influence of southward flowing Norwegian Sea bottom water (Johnson & Schneider 1969; Davies & Laughton 1972; Ruddiman 1972). Little of this water penetrates south of the Gibbs because the main current flows westward through the fracture zone into the Labrador Sea (Worthington & Volkmann 1969; Worthington & Wright 1970) (figure 39). Preservation of the well lineated topography is thus due to pelagic sedimentation unimpeded by bottom flow (Ruddiman & Glover 1972).

The Gardar Ridge sediment drift (figures 9 and 40) extends southward for 1000 km from 61° 30′ N to the Gibbs Fracture Zone. Many features such as moating around seamounts and undulations affirm the importance of differential deposition (Davies & Laughton 1972; Ruddiman 1972). However, the crest and east flank exhibit a remarkably regular sinuosity consisting of a series of NNE–SSW segments always linked by shorter westerly trending segments that locally continue westward as separate spurs (figure 40).

The Maury channel system (figure 40) extends from 60° N to the Gibbs Fracture Zone and may continue southward into the Biscay Abyssal Plain (Johnson *et al.* 1971; Ruddiman 1972; Cherkis, Fleming & Feden 1973*a*). Unlike many mid-ocean canyons, it is not situated between levees but follows the axis of greatest basin depth. It is widest between 59° and 57° N where the channel floor is underlain by a sequence of highly reflective stratified Pleistocene volcanogenic turbidites (Ruddiman 1972; Laughton, Berggren *et al.* 1972). South of 57° N, the channel is narrow and vee-shaped in profile. The sandstones may be derived from volcanic or intraglacial eruptions in Iceland. In support, Roberts, Laughton & Graves (1975) show two canyons originating close to the Vestmannejar (the well known site of the November 1972 eruption) continuing downslope into the Maury channel.

(ii) *The ocean basin southwest of Rockall Plateau*

An undulating relief closely resembling that observed on the Gardar Ridge, in the Hatton–Rockall Basin and Rockall Trough characterizes the ocean basin floor southwest of Rockall Plateau. The importance of differential deposition is demonstrated by the well developed moats encircling Gondor and Rohan seamounts and the low sediment ridges extending southward toward the Gibbs Fracture Zone (figure 9; figure 40). However, there is no major sediment drift other than the limited northern extension of the Feni Ridge (Jones, Ewing, Ewing & Eittreim 1970; Johnson *et al.* 1971; Ruddiman & Glover 1972). Depositional continuity does not appear to exist with the Gardar Ridge however, presumably because the Maury channel is an effective barrier.

The prominent magnetic seamounts (figures 3*b* and 40) between Edoras Bank and the Gibbs Fracture Zone are all situated on 60–76 Ma oceanic crust (Vogt & Avery 1974; Jones & Roberts 1975). Strikingly no comparable features are developed on contemporaneous oceanic crust to the west of Porcupine Bank or on the younger crust south of the Gibbs Fracture Zone (figures 40 and 3*b*). The relation of these seamounts to the spreading history of this area is not clear. The steep western escarpment of Eriador Seamount is parallel to and abutted by anomaly 24 (figure 3*b*) but is truncated at 54° N by an east–west trending fracture zone (Vogt & Avery 1974; Jones & Roberts 1975). To the south of the fracture zone, a low rise surmounted by several seamounts and extending SSE to the Gibbs Fracture Zone may be a homologue. The offset is similar to those observed on young rifted continental margins, e.g. the Gulf of California (Moore 1973), and its presence at the boundary between oceanic crust formed during distinct spreading episodes suggests it may mark an oceanic rift margin. Contemporaneous volcanoes developed in the hinterland of the plate may be represented by Gondor and Rohan seamounts, and perhaps by Fangorn Bank. The prominent SSE trending scarp developed south of the Gibbs Fracture Zone has been attributed to excessive intrusive activity during the 60 Ma reorientation in spreading (Vogt *et al.* 1971).

(*f*) *The Rockall Trough*

In this report, the Rockall Trough is considered to extend from the Wyville–Thomson Ridge to 53° 30′ N and as a subdued feature to 50° 30′ N. It may continue north from the Wyville–Thomson Ridge system to the Faeroe–Shetland Channel though that region will not be discussed here. For convenience the topography is reviewed in terms of four provinces; the Wyville–Thomson Ridge system, the east and west margins of the Rockall Trough and the seamounts situated within the Trough.

(i) *The Wyville–Thomson Ridge System*

The WNW-trending Wyville–Thomson Ridge system completely divides the deeper Rockall Trough from the shallower Faeroe Bank and Faeroe–Shetland Channel system to the north (figure 40) and is therefore of considerable geological and oceanographic importance (Ellett & Roberts 1973). In particular its effectiveness as a barrier to overflow of Norwegian sea water from the Faeroe Bank Channel into the Rockall Trough is directly relevant to Neogene sedimentation.

In detail, it consists of two distinct ridges, called the Wyville–Thomson Ridge and Ymir Ridge, separated by a deeper trough or gap that is closed to the Faeroe Bank Channel but is directly connected with the Rockall Trough (figure 40). The maximum depth in the gap (1722 m) is considerably greater than the adjacent Rockall Trough and is located close to the most constricted part of the gap. Ellett & Roberts (1973) considered these features and sediment drifts within the gap to indicate a strong westerly current driven by intermittent overflow of Norwegian sea water from the Faeroe Bank Channel into the gap via one of the depressions in the Wyville–Thomson Ridge. In support, near bottom current speeds in excess of 80 cm s^{-1} (R.R.S. *Shackleton*, cruise 6/73, unpublished data) have recently been measured in the gap. In this context, it is noteworthy that the Feni Ridge sediment drift begins close to the gap.

The margins of the Rockall Trough. There are many important physiographic differences between the east and west margins of the Rockall Trough. For example, the complete sequence of shelf slope and rise is continuously present only west of the British Isles (figure 8). The axis of

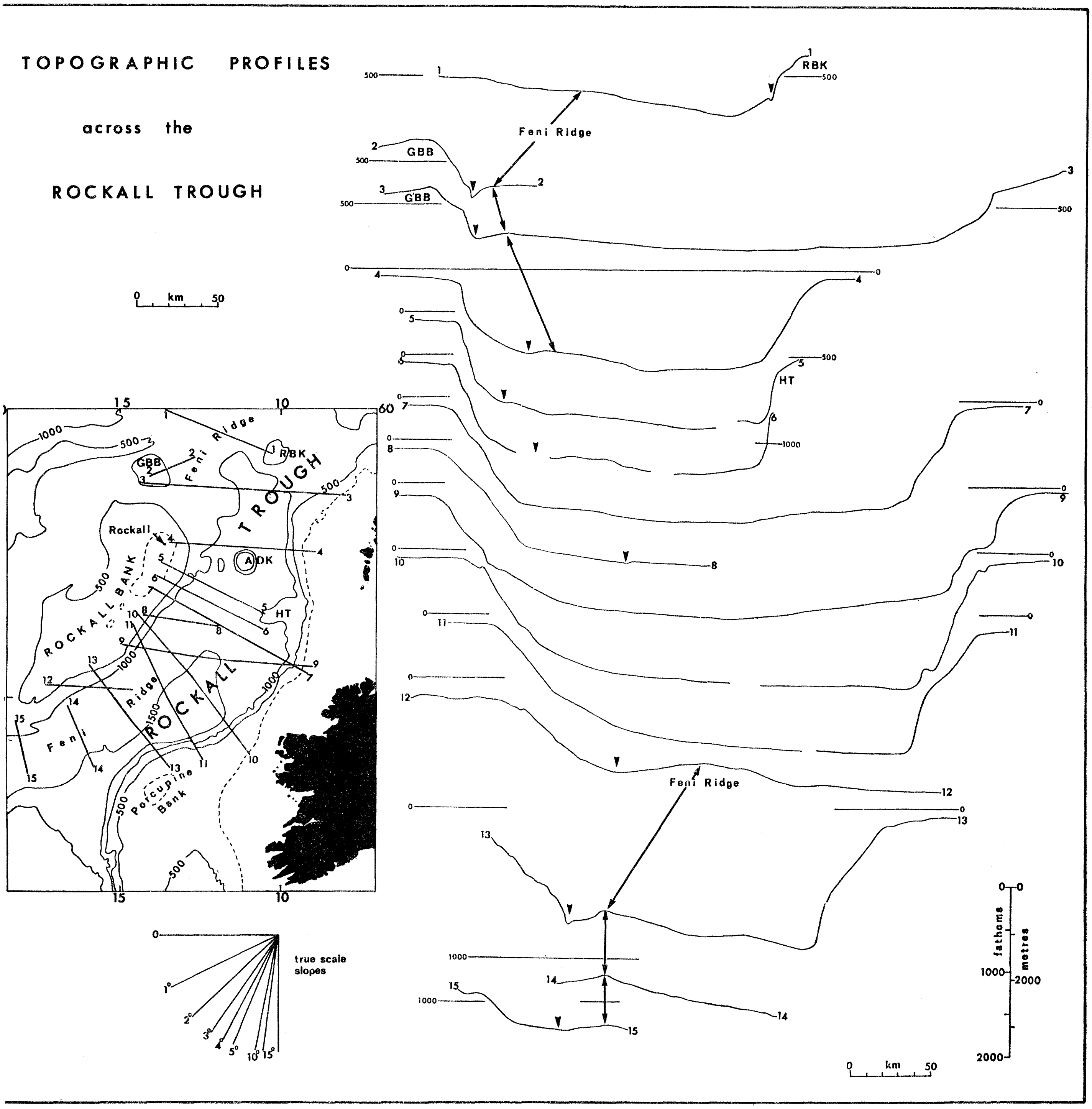


FIGURE 8. Topographic profiles across the Rockall Trough.

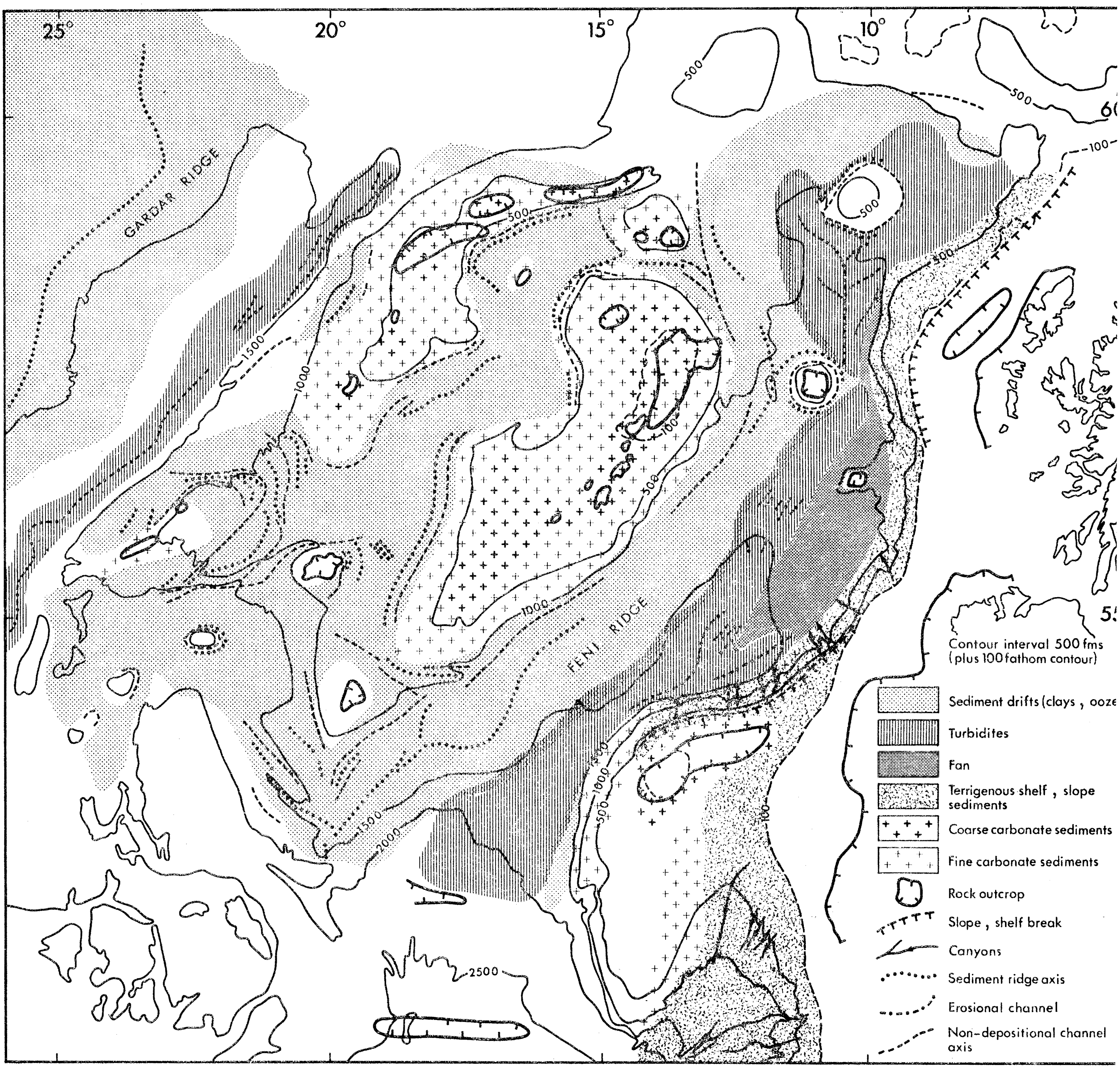


FIGURE 9. Distribution of sea bottom lithologies, sediment drifts and fans on and around Rockall Plateau; also rock outcrops, slope, shelf break, canyons and bottom features. Based principally on seismic profiles shown in figure 12, and unpublished echo-sounder traverses of R.R.S. *Discovery*, R.R.S. *John Murray*, M.V. *Surveyor* and R.V. *Erika Dan*.

maximum depth provides a natural boundary between east and west margins in the following review (figures 8 and 40).

(ii) *The west margin of the Rockall Trough*

The shelf, slope and rise provinces of a typical continental margin (Heezen, Tharp & Ewing 1959) are not fully developed and the margin consists of three distinct segments characterized by the presence or absence of shelf and slope. A broad rise comprised almost entirely by the Feni Ridge is common throughout and extends to the axis of the Rockall Trough.

The northernmost of these segments extends from the Wyville–Thomson Ridge system to George Bligh Bank and is a broad ridge surmounted by the Faeroe Bank (figure 1), Bill Bailey and Lousy Bank Seamounts (figure 40). The ridge seems an obvious prolongation of Rockall Plateau though it is narrower and has no comparable relief (figure 40). The seamounts are probably composed of basic igneous rocks (Bott & Stacey 1967; Avery, Burton & Heirtzler 1968) and collectively form an ENE trend (figures 1, 3*b* and 40) alined with the coeval (?), Palaeocene igneous platform of the Faeroes (Tarling & Gale 1968; Rasmussen & Noe-Nygaard 1970). The Rockall Plateau microcontinent may continue beneath them forming the continental crust beneath the Faeroe Islands (Bott & Watts 1971; Bott *et al.* 1971; Casten 1973; Bott *et al.* 1974).

The gaps between the seamounts have been previously considered as possible entrances into the Rockall Trough for Norwegian sea water that has spread southwestward from the Faeroe Bank Channel (Steele, Barrett & Worthington 1962). However, the sill depths are everywhere shallower than the channels associated with that outflow of Norwegian sea water (Roberts 1973*b*). In support recent hydrographic surveys made during the ICES Overflow 1973 expedition have not revealed Norwegian sea water between Lousy Bank and Bailey Bank Seamounts (R.R.S. *Shackleton*, cruise 6/73, unpublished data).

The section of the west margin most closely resembling a typical continental margin extends between George Bligh Bank and Lorien Bank (figures 40 and 8). For the most part, it consists of a relatively steep slope forming the edge of the Rockall Trough though the slope is intersected by easterly trending depressions at 58° 30′ N and 54° 30′ N. The only true shelf is developed on Rockall Bank between 58° and 56° N in depths less than 120 fathoms (219 m). The steepest slope is also developed there and is characterized by a distinctive notch (figure 8) profile 10) at depths between 287 and 390 fathoms (525 and 713 m) and impersistent minor irregularities that are the topographic expression of two large slumps (Roberts 1972*a*). The slumping is restricted to the area of optimum shelf development and is absent to the north and south. Roberts (1972*a*) related the slumping to the lower Flandrian sea level and suggested it was initiated by local loading of the upper slope due to increased deposition. The absence of slumping to the south may reflect reduced sediment supply available in the greater depths.

South of 56° 40′ N, the slope is typically convex and no canyons are present (figure 8, profiles 4–11). Changes in trend of the slope to the south of 55° 30′ N are closely followed by the Feni Ridge axis. There is also a change in the microtopography indicated by the appearance of small discrete ridges and channels trending NNE (figure 40; figure 8, profile 13).

(iii) *The Feni Ridge*

A broad sediment drift called the Feni Ridge is present in the Rockall Trough and is best developed between 56° N and 52° 30′ N (figures 40, 8 and 9). Like other sediment drifts in the ocean basins, the Feni Ridge has a number of diagnostic physiographic and stratigraphic characteristics that include an undulating, weakly reflective seabed, the presence of numerous hyperbolae on echo-sounder records, non-depositional and/or erosional channels, and a marked unconformity between the seabed and underlying basement or reflector (Schneider *et al.* 1967; Jones *et al.* 1970; Johnson *et al.* 1971; Le Pichon *et al.* 1971*a*, *b*). Numerous unpublished *Discovery* echo-sounder traverses and the seismic profiles given here have been examined for these characteristics (figure 9). Such features characterize the whole length of the western Rockall Trough though a ridge is not present throughout.

(*a*)

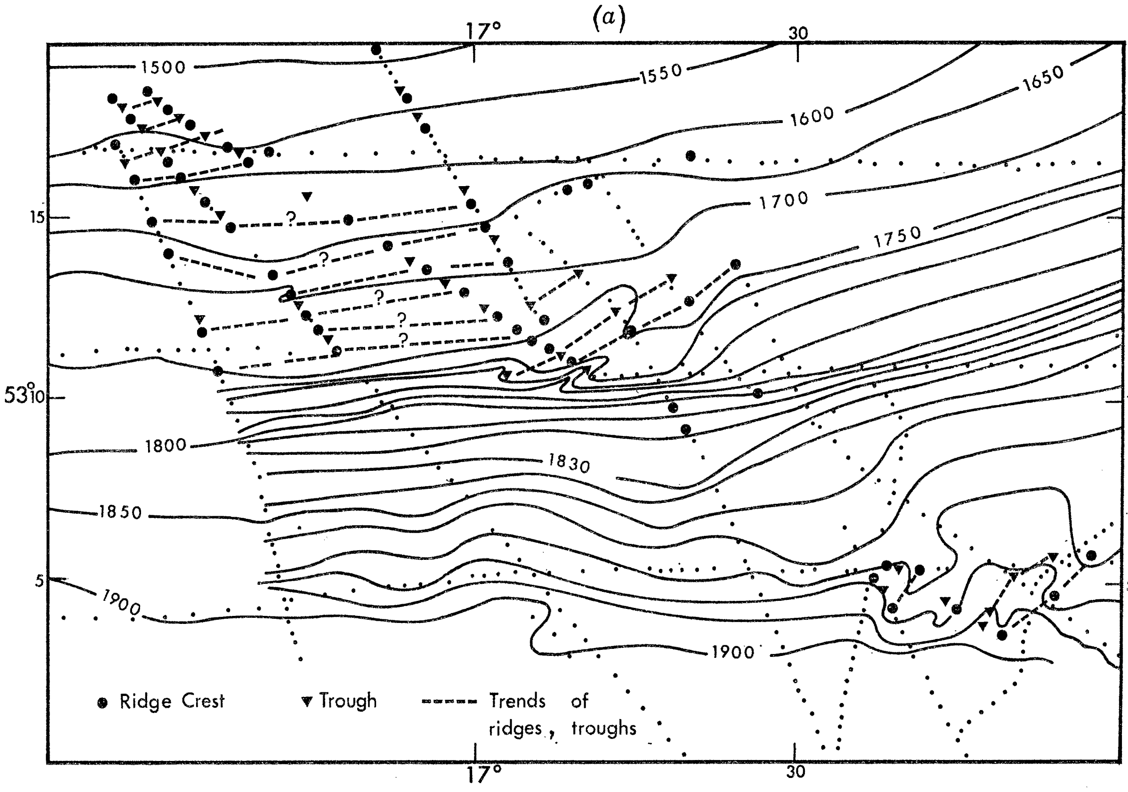


FIGURE 10(*a*). For description see opposite.

The Feni Ridge begins as an arcuate ridge close to the gap between the Wyville–Thomson and Ymir Ridges (figure 40). Over the length of the drift, trend, relief and microtopography seem closely related to the nature of the adjacent slope so that a ridge is best developed where the slope steepens and/or changes trend (figure 40; figure 8, profiles 2, 3 and 13). The development of the hyperbolae and small scale bottom roughness (4–6 m) relief (wavelength 0.2–0.6 km) and undulations (36–54 m relief, wavelength *ca.* 2 km) show a similar relation. Where the ridge is poorly developed on a gentle slope (e.g. near 59° N 12° W), the undulations are sparse and a well developed small scale roughness is present; but where the ridge is prominent, its east flank is characterized by large undulations. These undulations can be shown to trend parallel or sub-parallel to the Feni Ridge axis even as it changes trend by 90° (figure 10*a*, *b*). It is worth emphasizing that a particularly close relation (discussed in detail later) to the trend and form of the adjacent slope is shown between 55° 30′ and 53° N (figures 40 and 9). As the Feni Ridge diverges from the slope, it initially widens and increases in relief but there-

(b)

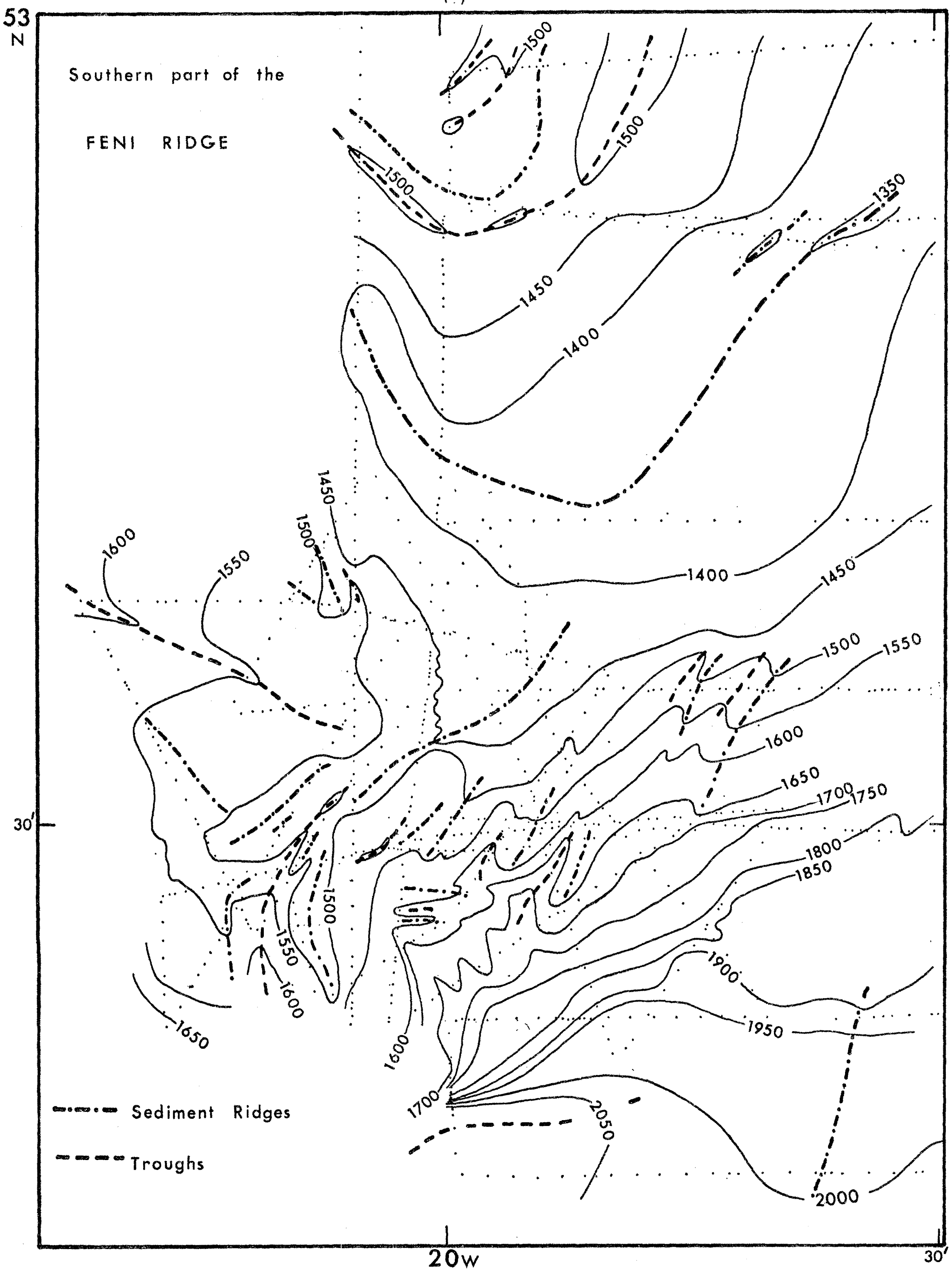


Figure 10. (a) Trend of undulations on the east flank of the Feni Ridge. Dots indicate echo-soundings used in contour compilation. (b) Trend of undulations on the south Feni Ridge. Dots indicate echo-soundings used on contour compilation.

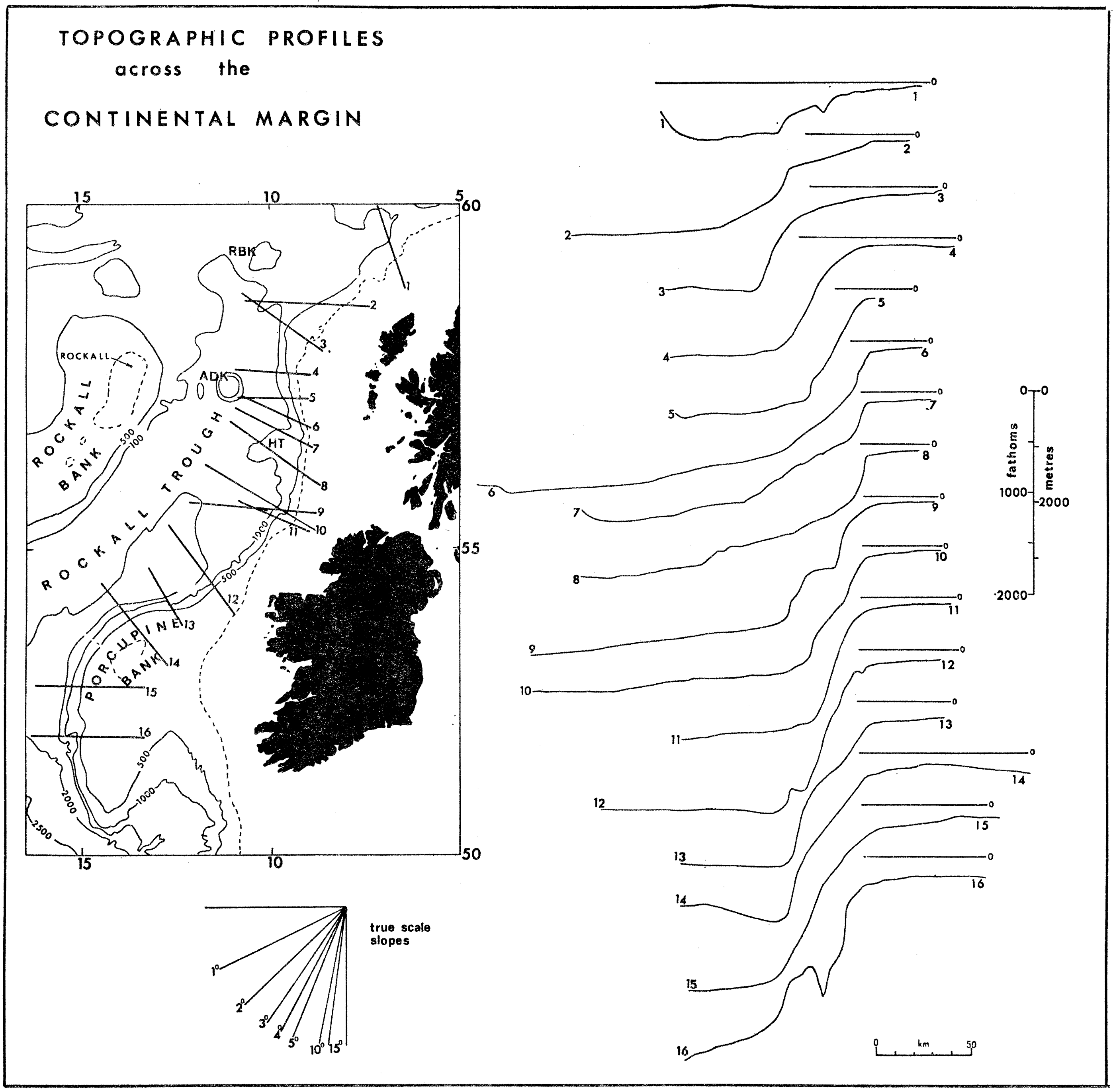


FIGURE 11. Topographic profiles across the continental margin west of the British Isles.

after its sinuosity is remarkably parallel to the changes in the trend of the adjacent slope. An unusual feature is the semi-circular depression enclosed by the sinuous ridge (figure 40).

The eastern boundary of the Feni Ridge sediment drift effectively corresponds to the axis of maximum depth in the Rockall Trough (figures 40 and 9). South of Anton–Dohrn seamount, there is a sharp boundary with a flat strongly reflective seabed cut by small channels (relief *ca.* 15 m) probably extending axially down the Rockall Trough. The eastern edge of this zone marks the distal edge of the Barra and Donegal Fans (figure 9). Axial transport from the fans may have inhibited further eastward extension of the Feni Ridge. South of 53° N, there is a sharp boundary between the undulations and a smooth relief (figure 9) which merges with the continental rise to the west of Porcupine Bank. A 500 fathom (914 m) high ridge extending southward from 52° 30′ N is a sediment drift that may be an extension of the Feni Ridge (Johnson *et al.* 1971).

(iv) *The continental margin west of the British Isles*

The continental margin northwest and west of the British Isles consists of a broad continental shelf, a narrow continental slope and a broader continental rise (figures 40 and 11).

Between 55° and 60° N, an inner shelf and outer shelf are naturally divided by the Outer Hebrides and the low sill connecting those islands to Northern Ireland. The only major relief on the outer shelf is the shallow basin separating the distant islands of Rona, Flannan and St Kilda from the Outer Hebrides. Irregular topography between Ireland the Outer Hebrides may in part reflect southwest convergence of the Minch and Great Glen Faults (McQuillin & Binns 1973).

The superficial sediments of the shelf and their distribution have been examined by Kenyon & Stride (1970) and Belderson, Kenyon & Stride (1971). Their results show the sediment transport paths lie mainly parallel to the shelf edge indicating little transport beyond the shelf. Belderson *et al.* (1973) have shown iceberg plough marks occur in depths exceeding 150 m on the outer shelf and upper slope.

Between 58° and 60° N, there are two distinctive breaks in slope (figures 40, 9 and 11). The shallowest occurs at about 100 fathoms (183 m) and corresponds to the well defined shelf break further south. The 100 fathom (183 m) contour also outlines basins some 20 km in width and 50 fathoms (91 m) in relief (e.g. at 59° N 7° W) that trend perpendicular to the shelf edge and across the NNE trending basement structure (Bott & Watts 1971). They are comparable and may be akin to the glacially overdeepened shelf troughs found off Norway and Labrador (Holtedahl 1970; Grant 1972). The deeper slope break at about 300 fathoms (548 m) marks a partially buried, down warped older shelf (Stride, Curray, Moore & Belderson 1969). South of 58° N there is a well developed shelf break at 100 fathoms (183 m) but the 300 fathoms (548 m) slope break is weakly developed (figures 40, 9 and 11). The linearity of this slope is apparently controlled by an important fault. Between 57° and 54° N, many canyons begin in depths of 300–500 fathoms (548–913 m) that are similar to the slope break to the north and may be related to downwarping of the margin.

A small shelf is developed on Porcupine Bank but further south the deeper crest has a convex profile (figure 11, profiles 14, 15 and 16). The absence of canyons on the east and west sides of the Bank is undoubtedly a reflexion of the limited sediment supply available on the deep Bank. A large valley near 52° N has some resemblance to a canyon but contains little sediment and lies within a tilted fault bounded block. This fault may control the extremely steep linear slope to the south (figure 40; figure 11, profile 16). This valley also lies close to the projected trend of the Gibbs Fracture Zone. However, the offset in the margin is minimal compared to the offsets in the ridge axis and anomaly 32 (76 Ma) (figure 3*b*) and suggests no direct correlation is present (cf. Fleming, Cherkis & Heirtzler 1970; Le Pichon *et al.* 1972; Cherkis, Fleming & Massingell 1973*b*).

The pattern of canyon development seems to bear a close relationship to the shelf topography and the shelf transport paths discussed by Kenyon & Stride (1970). The development of canyons north of 59° N may be related to shelf transport paths which there lie oblique to the shelf edge, and their absence to the north and west of St Kilda to the effective trapping of sediments in the basin between St Kilda and the outer Hebrides (figure 40). The optimum development of canyons and fans occurs off northwest Ireland where the transport paths lie perpendicular to the shelf break. One large canyon at 55° 30′ N 9° W is associated with an offset in the margin structurally controlled by the Great Glen Fault (Roberts 1969; Bailey, Crzwacz & Buckley 1974; this paper, figure 21). In contrast, the more numerous and larger canyons cutting the east slope of the Seabight undoubtedly reflect the greater sediment supply from the Irish Shelf and Celtic Sea.

The physiography of the continental rise between 57° 30′ N and 54° N is dominated by two large fans, here called the Barra and Donegal Fans that extend to the axis of the Trough. The area of the Barra Fan is approximately 2500 km^2 and the Donegal Fan *ca.* 9000 km^2. The fans are cut by minor channels and steps; one channel extends along the base of the slope to join the axis of the Trough north of Porcupine Bank (figure 11, profile 12). The relation of the fans to the strongly reflective axial area has been mentioned previously and is summarized in figure 9. West of Porcupine Bank, local steepening of the rise in a N–S direction occurs at 52° N and 50° 30′ N and, in the former case, may reflect overspill of sediments across the almost completely buried Gibbs Fracture Zone (Cherkis *et al.* 1973*b*).

(v) *Seamounts*

Three seamounts are present within the Rockall Trough. The topography of the axial Rosemary Bank and Anton–Dohrn seamounts has been discussed by Ulrich (1964), Dietrich & Ulrich (1961) and by Roberts *et al.* (1974). Both seamounts are surrounded by a well developed moat flanked by an outer ridge. In contrast, Hebrides Terrace Seamount is situated at the base of the slope and is not associated with a moat. Both Hebrides Terrace Seamount and Anton–Dohrn Seamount are flat-topped.

(vi) *Comparison of the east and west margins*

There is an approximate congruency between the facing margins of the Rockall Trough that was used by Bullard *et al.* (1965) to constrain their fit of the North Atlantic Ocean. In detail, however, the margins converge at several points; for example between 58° and 57° N. The most important convergence is at 54° N where the east–west trend of Porcupine Bank contrasts with the NE–SW trend of Rockall Plateau. This convergence is the cause of the overlap in the fit of the opposite margins based on the 500 fathom (914 m) contour. The overlap indicates either substantial rotation of Porcupine Bank or a different sub-surface shape of the continent-ocean boundaries in the Rockall Trough to that implied by the bathymetry.

Secondly it is striking that no canyons are developed on the Rockall Plateau. Their absence implies little sediment has been contributed to the Trough by erosion of Rockall Plateau and may therefore account for the preservation of the Feni Ridge sediment drift in contrast to the development of the Barra and Donegal Fans in the area of greater sediment supply. A similar contrast in depositional regime is implied by the absence of canyons on the west slope of the Porcupine Seabight.

3. Seismic reflexion profiles

(a) *Instrumentation and data analysis*

The seismic reflexion profiles discussed in this report were taken during R.R.S. *Discovery* cruises 7, 29, 33 and 47, M.V. *Surveyor* cruise 1/71 and D.V. *Glomar Challenger* Leg 12. The profiling system used from *Discovery* and *Surveyor* consisted of an airgun, a 73 m (200 ft) NIO hydrophone array with 100 elements and a triggered 45 cm (18 in) Mufax wet paper recorder with variable delays and time varied gain. The filtered data were displayed at 2 or 4 s sweeps. A summary of the system used during each cruise is listed below.

	sound source	filter settings	display
R.R.S. *Discovery*, cruise 7	60 kJ arcer	30–300 Hz	45 cm (18 in) dry paper recorder
R.R.S. *Discovery*, cruise 29	490 cm^3 (30 in^3) Lamont airgun	30–300 Hz 20–80 Hz	45 cm (18 in) wet paper recorder
R.R.S. *Discovery*, cruise 33	490 cm^3 (30 in^3) Lamont airgun	30–300 Hz	45 cm (18 in) wet paper recorder
M.V. *Surveyor*, cruise 1/71	490 cm^3 (30 in^3) 80 kJ sparker	30–300 Hz	45 cm (18 in) wet paper recorder
R.R.S. *Discovery*, cruise 47	650 cm^3 (40 in^3) Bolt airgun	30–80 Hz	45 cm (18 in) wet paper recorder
D.V. *Glomar Challenger*	650 cm^3 (40 in^3) Bolt airgun	various	45 cm (18 in) dry paper recorder

All the *Discovery* and *Surveyor* data were tape recorded and have been replayed at various filter settings and sweep speeds to clarify ambiguous reflexions. Two short sections of *Discovery* 29 data acquired by using matched gun and array depths (Ziolkowski 1971) have been digitally processed by Geophysical Service International using both minimum and non-minimum phase time varied deconvolution.

Objectivity in the illustration of seismic profiles is difficult, for to fully and convincingly present the data both the original record and interpretation should be shown. The profiles are thus presented here as mainly line overlays on photographic reductions of the original records and secondly as a structural and stratigraphic interpretation. Detailed interpretation of the age and stratigraphy of a particular profile is not possible because the thick Neogene cover prevents sampling of the deeper reflectors by conventional methods. The geological interpretation is thus generalized and is based mainly on deep sea drilling data, refraction velocities, magnetics, gravity and the limited available geological data. Continuity of the major reflectors has been established by crossovers between the profiles given here and those reported by Jones *et al.* (1970); Le Pichon *et al.* (1970) and Ruddiman (1972 and personal communication).

The maps of isopachs on reflector R4 and basement were drawn in the following way. Sediment thicknesses in two way time above R4 and basement were measured at 10 min intervals, plotted and contoured manually. Crossovers afforded an internal check on consistent identification and measurement. The greatest element of error and interpretation exists in the isopachs on basement. This is because not all seismic profiles show 'basement' as evidenced by deeper reflectors observed on intersecting profiles. Further the definition of basement is itself geologically ambiguous. For example, basement west of Rockall Plateau is the top of layer 2, Precambrian to Tertiary igneous rocks on Rockall Plateau and Lower Palaeozoic rocks on the shelf west of Ireland. In this paper, basement is defined as oceanic, i.e. the top of oceanic layer 2 or continental; i.e. Precambrian to Palaeozoic igneous, sedimentary and metamorphic rocks with the local inclusion of Tertiary and Cretaceous igneous rocks. Basement has been

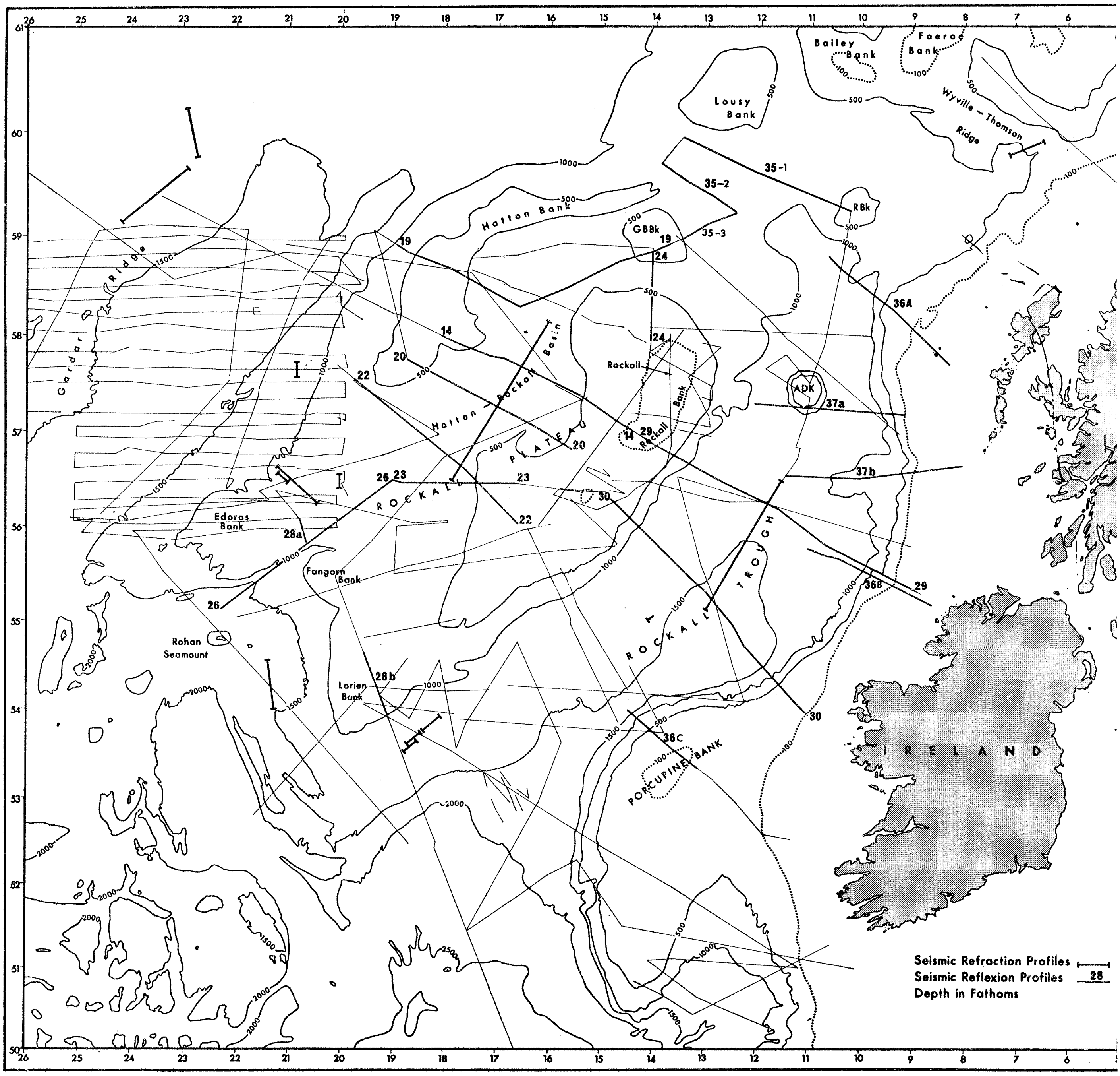


FIGURE 12. Location of seismic profiles (by figure number).

identified by its seismic refraction velocity, by bottom sampling and its presence on a particular profile by a comparison of reflexion and refraction depths. It should be noted that acoustic continuity of continental and oceanic basement does not imply geological continuity.

Thicknesses have also been measured on profiles reported by Jones *et al.* (1970), Le Pichon *et al.* (1970), Sichler, Mascle & Cressard (1972). J. Ewing kindly allowed me to use unpublished Vema-27 records on the Rockall Plateau. An isopach map of the east flank of the Reykjanes Ridge has been incorporated in the total sediment thickness map. Isopachs on acoustic basement in the Porcupine Bank area (Clarke, Bailey & Taylor-Smith 1971) have been modified to accord with the greater basement depths shown by the other data. A total of 30000 km of seismic profile has been used in the isopach maps. Track control is shown on the maps. Profiles illustrated in this report are keyed in figure 12.

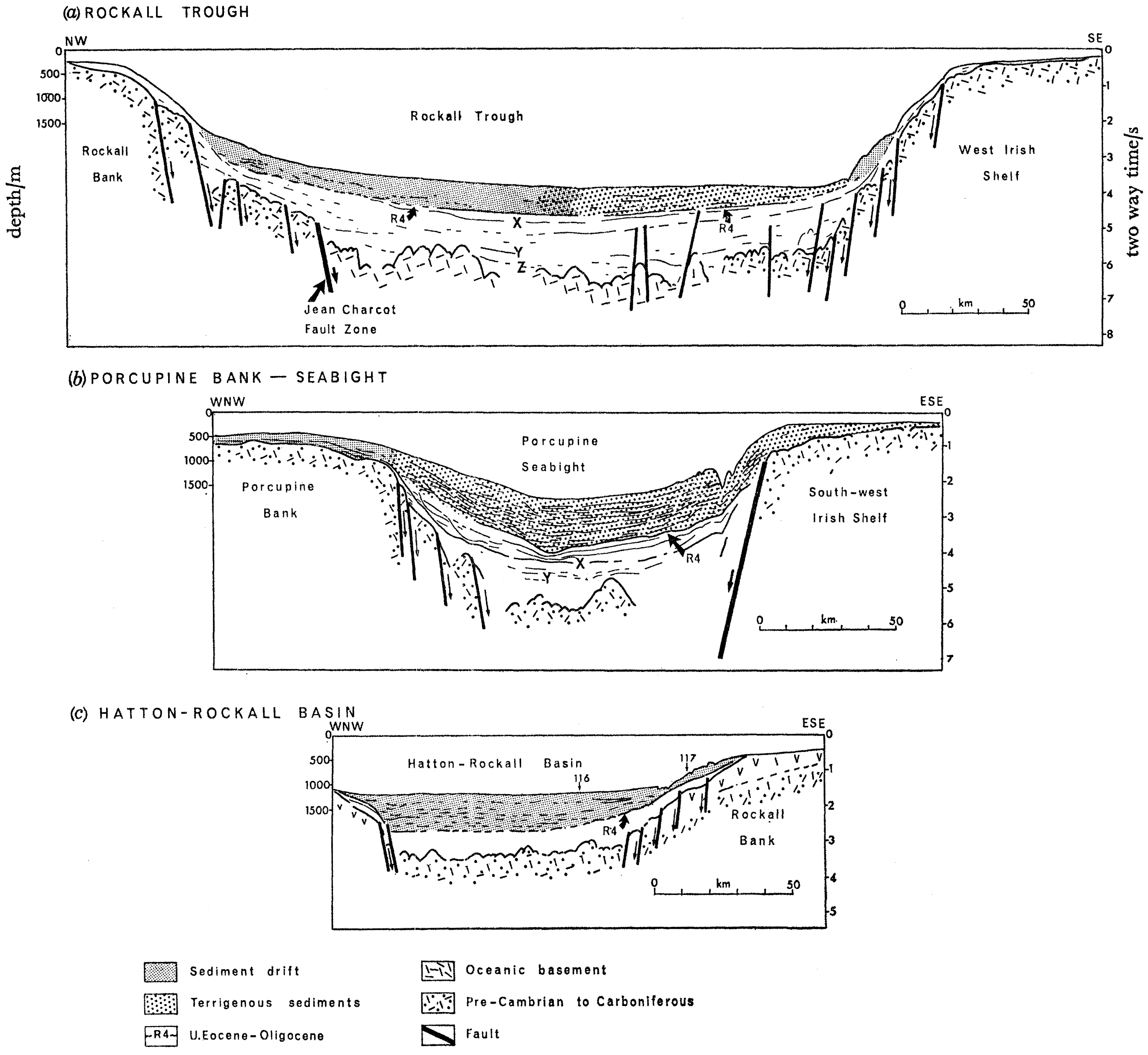


FIGURE 13. Comparative seismic reflexion sections across the Rockall Trough, Hatton–Rockall Basin and Porcupine Seabight.

(b) *The sedimentary basins*

Four large sedimentary basins occur in the area. Of these, the Rockall Trough and East flank of the Reykjanes Ridge are developed on oceanic crust, in contrast to the Hatton–

Rockall Basin and Porcupine Seabight which are developed on thinned continental crust (Gray & Stacey 1970; Scrutton, Stacey & Gray 1971; Scrutton 1972).

On the youngest, post-60 Ma basement of the Reykjanes Ridge there are only 0.6 km of sediment compared to more than 5 km beneath Rockall Trough; comparable differences in thickness clearly exist between the Hatton–Rockall Basin and Porcupine Seabight (figure 13). In spite of these differences, the basins have a broadly similar structural framework and pattern of subsequent deformation and sedimentation. Reflector R4 for example is common to the east flank of the Reykjanes Ridge, Hatton–Rockall Basin, Rockall Trough and the Porcupine Seabight, and marks a period of downwarping coupled with a major change in sedimentation regime.

Evidence of widespread continuity of R4 is given by intersecting seismic profiles on the southwest margin of Rockall Plateau and to the northwest of Edoras Bank (figure 12) which demonstrate that R4 and the overlying cherts are equivalent to the zone of diffuse reflectors called R by Jones *et al.* (1970) and Ruddiman (1972). R4 can thus be followed southward beneath the Feni Ridge into the Rockall Trough, thereby establishing a wider correlation. Ruddiman (1972) has shown that R4 continues westward to the Reykjanes Ridge where it pinches out on 37 Ma oceanic basement.

Recognition of the widespread occurrence of R4 permits discussion of basin geology in terms of three units: a post-R4 series, a pre-R4 series and basement without prejudice to the variations in the stratigraphy of the post-R4 series and the age of the pre-R4 series that are discussed in each section.

(*c*) *The Rockall Plateau*

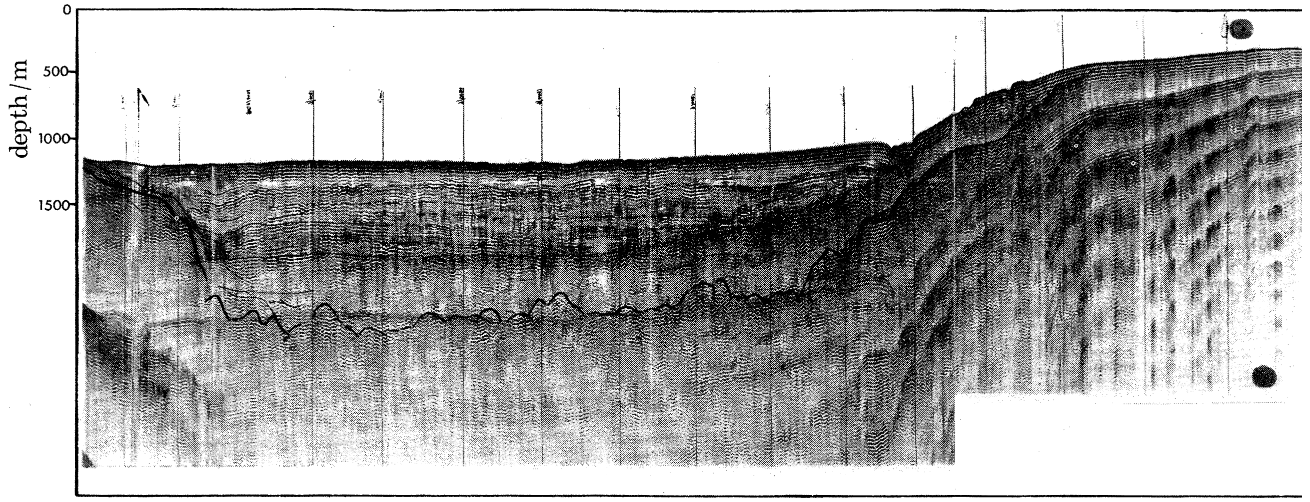
On the Rockall Plateau, the thickest sediments underlie the Hatton–Rockall Basin and are partly contained by the basement highs forming the shallow banks (Roberts *et al.* 1970; Scrutton & Roberts 1971). Within the Basin, a major reflector originally called 4 underlies the basin at about 1.0 s depth and comprises an unconformity at the basin margins (figure 14). This reflector has been renamed R4 because of its correlation with reflector R of the Rockall Trough and Reykjanes Ridge (Jones *et al.* 1970; Ruddiman 1972). A steeper and irregular reflector called reflector 5 at about 2.0 s depth marked the acoustic basement. Lateral continuity of reflectors 1, 2 and 3 previously identified in the post-R4 series has not been proven by subsequent seismic profiling.

(i) *Age and post-basement stratigraphy of the Hatton–Rockall Basin*

Evidence for the age and stratigraphy of the succession west of Rockall Plateau is given from JOIDES Holes 116 and 117 (Laughton, Berggren *et al.* 1972) in the Hatton–Rockall Basin (figures 14 and 15) and cores reported by Ruddiman (1972).

JOIDES Hole 116 (figure 15) bottomed in Late Eocene limestones at 854 m. Much of the sequence consisted of Quaternary and Neogene oozes passing into Lower Miocene cherts at 670 m. The cherts became abundant at 700 m and continued to the base of the Oligocene at 808 m. Underlying Late Eocene hard chalks passed into soft oozes at 840 m. An intra Early–Late Oligocene unconformity was inferred between 730 and 750 m.

In Hole 117 (figure 15) Quaternary to Neogene oozes overlay Oligocene cherts resting unconformably on Early Eocene clays. The clays passed down into conglomeratic Upper Palaeocene overlying subaerial basalt.



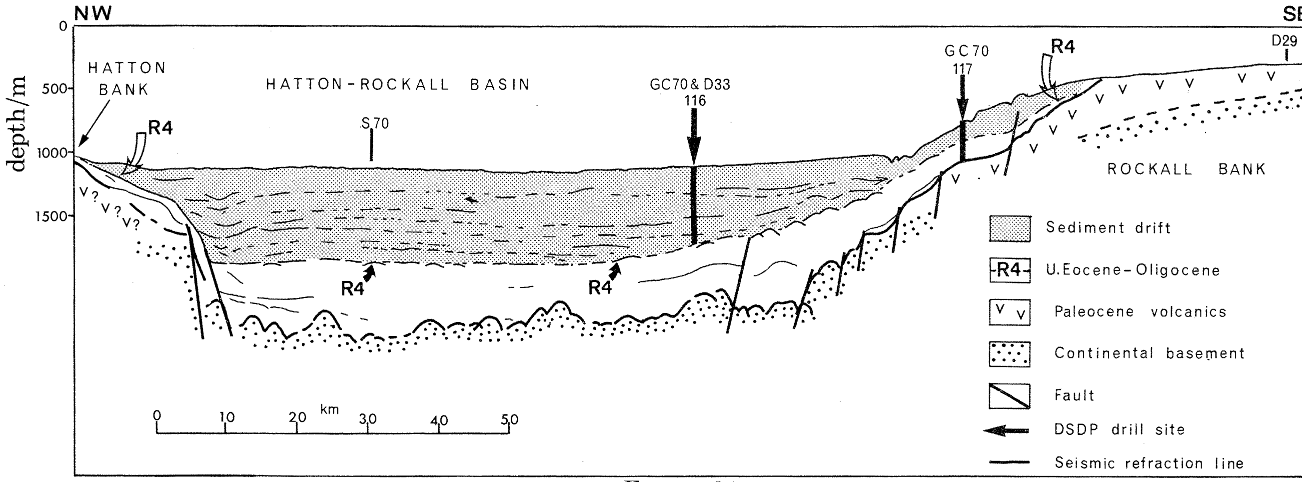
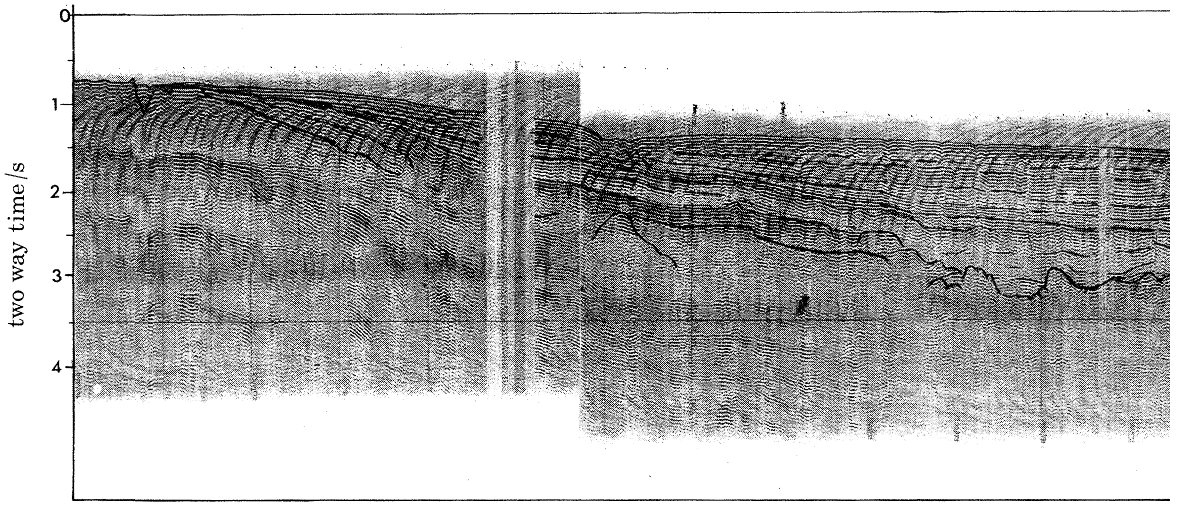
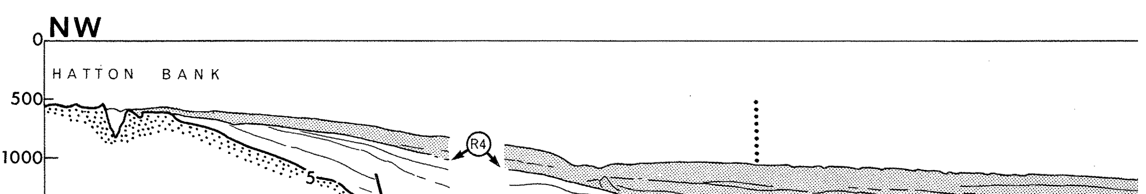
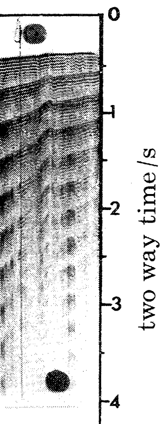
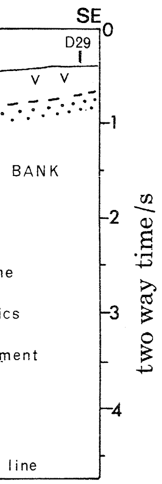


FIGURE 14.
Seismic reflexion profile (14) across the Hatton-Rockall Basin through JOIDES sites 116 and 117. Profile is located in fig









ted in figure 12.

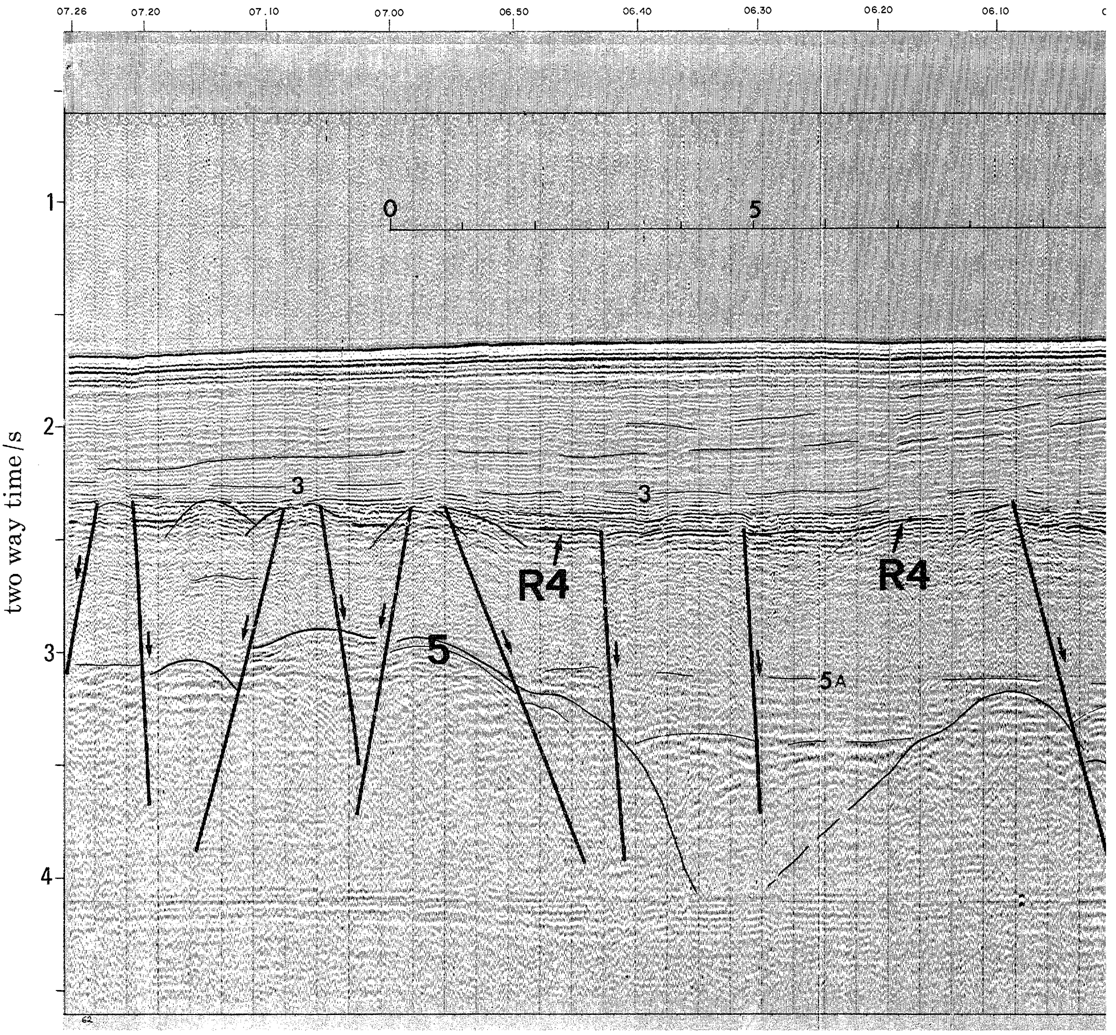
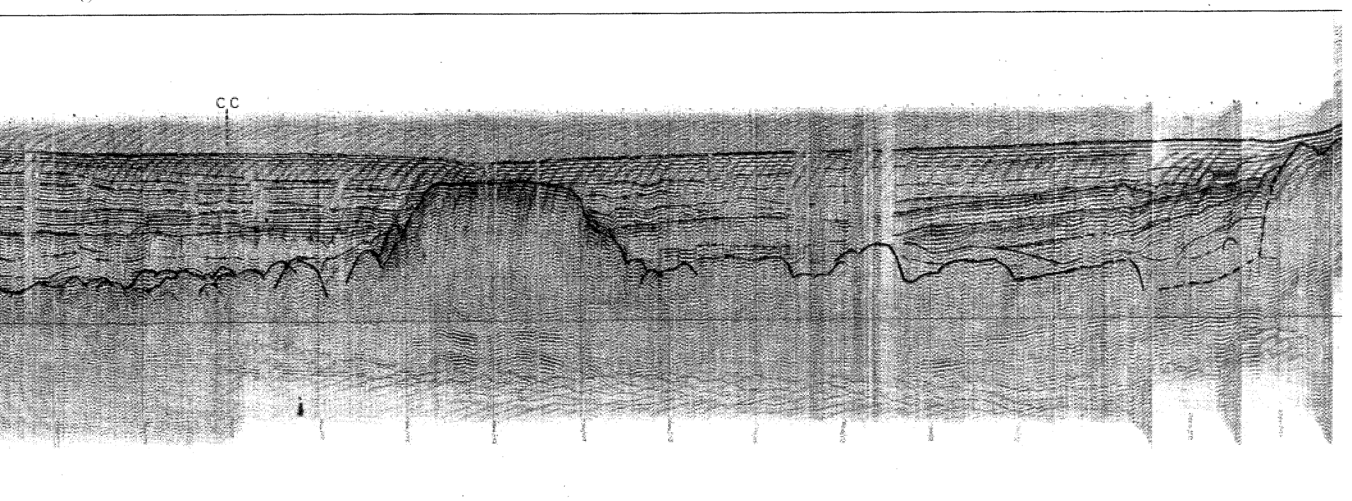
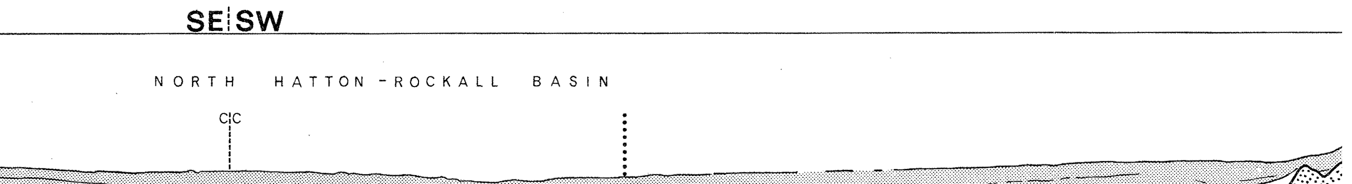
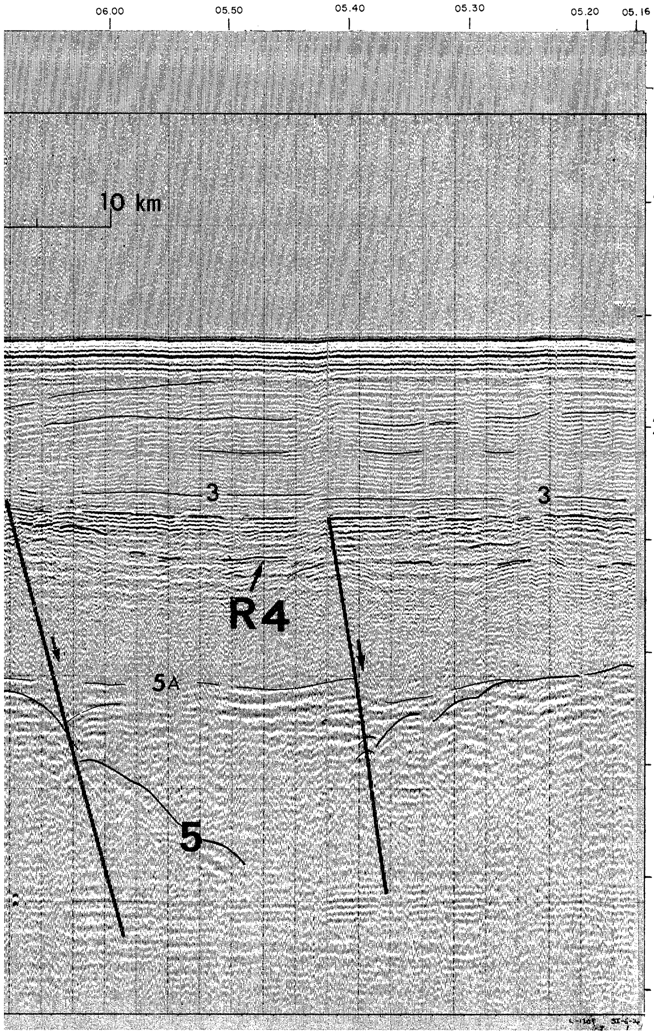


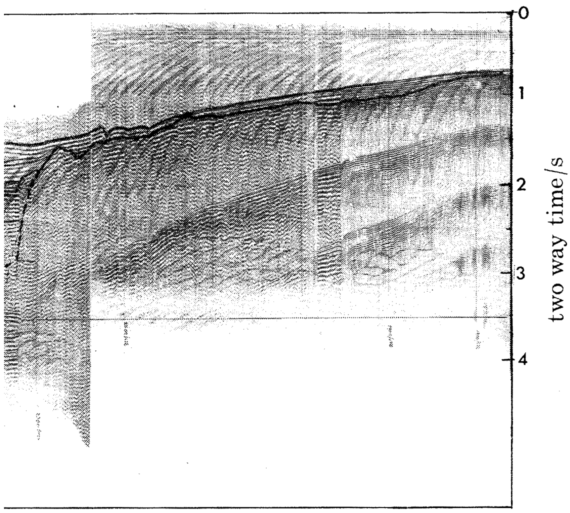
FIGURE 16. Deconvolved seismic reflexion profile in the Hatton-Rockall Basi

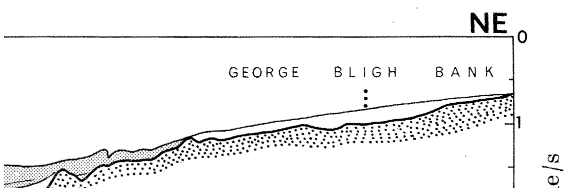






kall Basin. Section is located in figure 22.





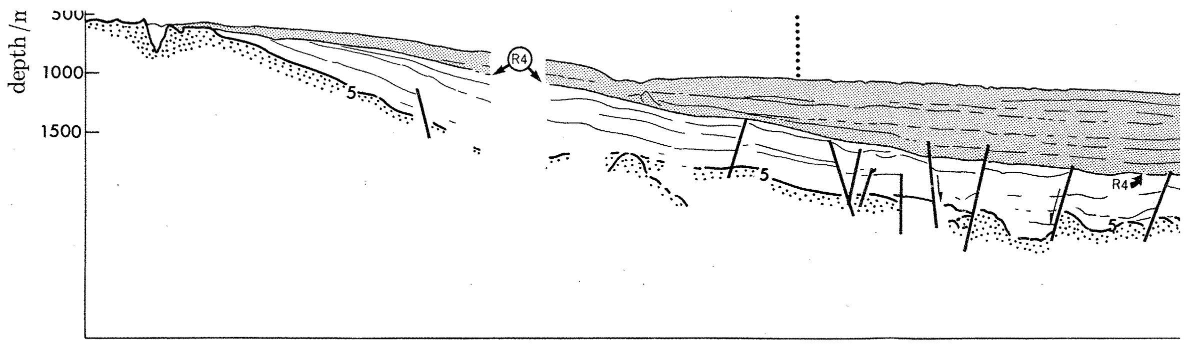
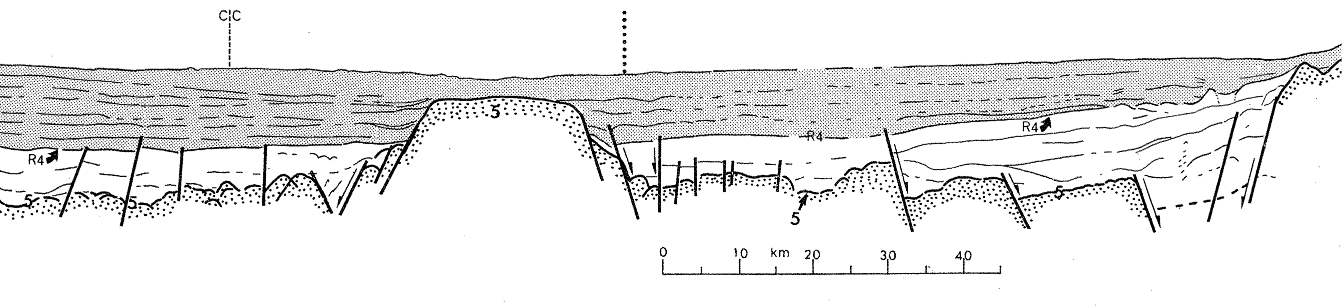
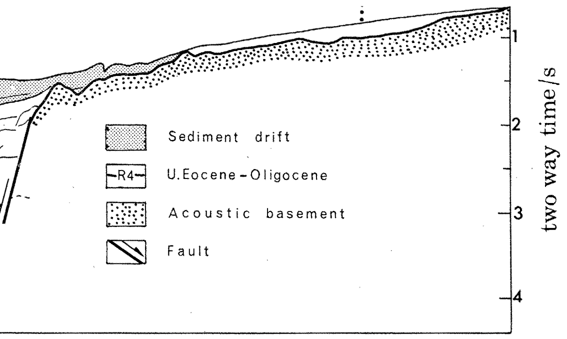


FIGURE 19. Seismic reflexion pr



lexion profile (19) across the northern Hatton-Rockall Basin. Profile is located in figure 12.



Sediment drift
R4
U.Eocene-Oligocene
Acoustic basement
Fault
1
2
3
4
two way time/s

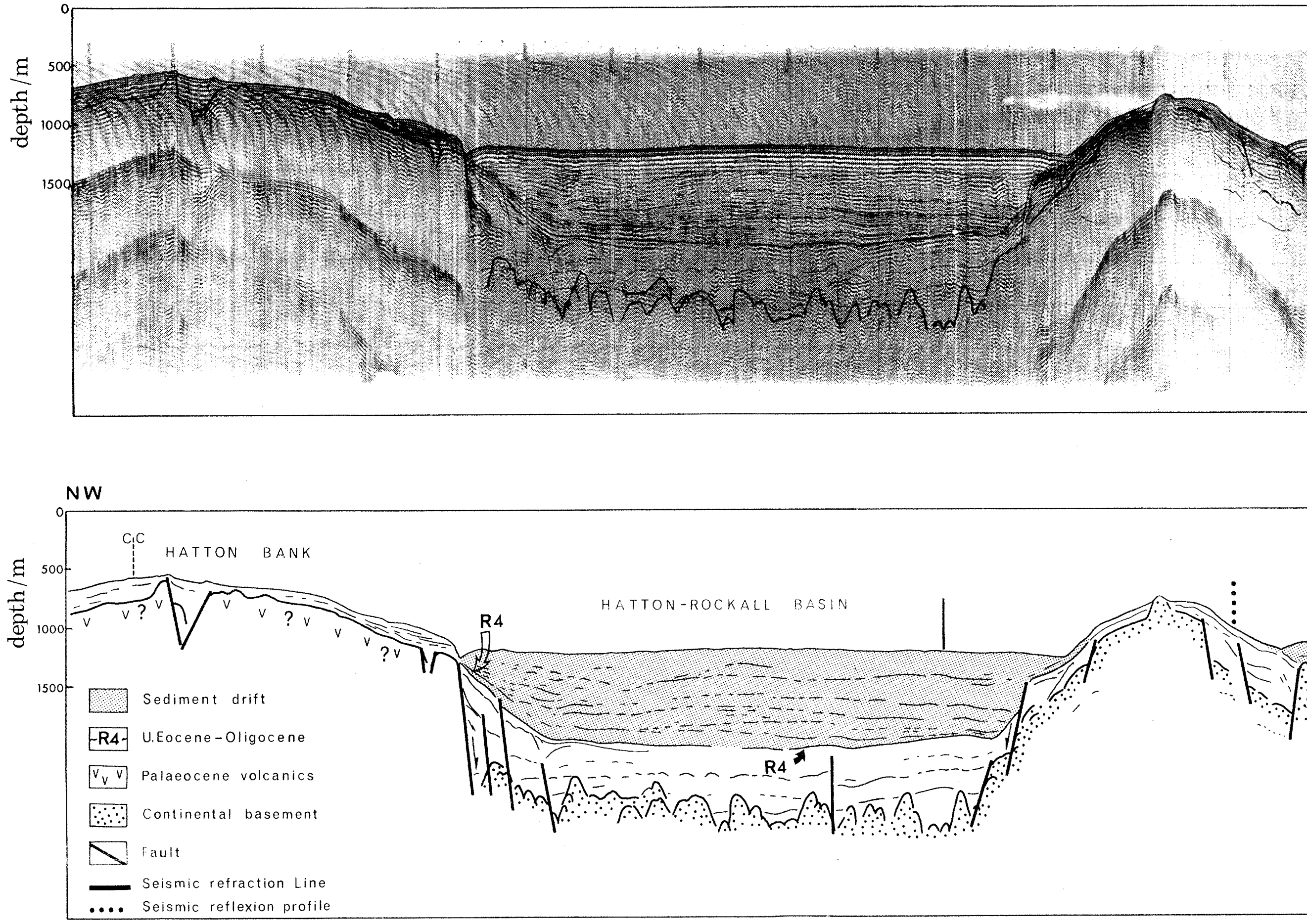
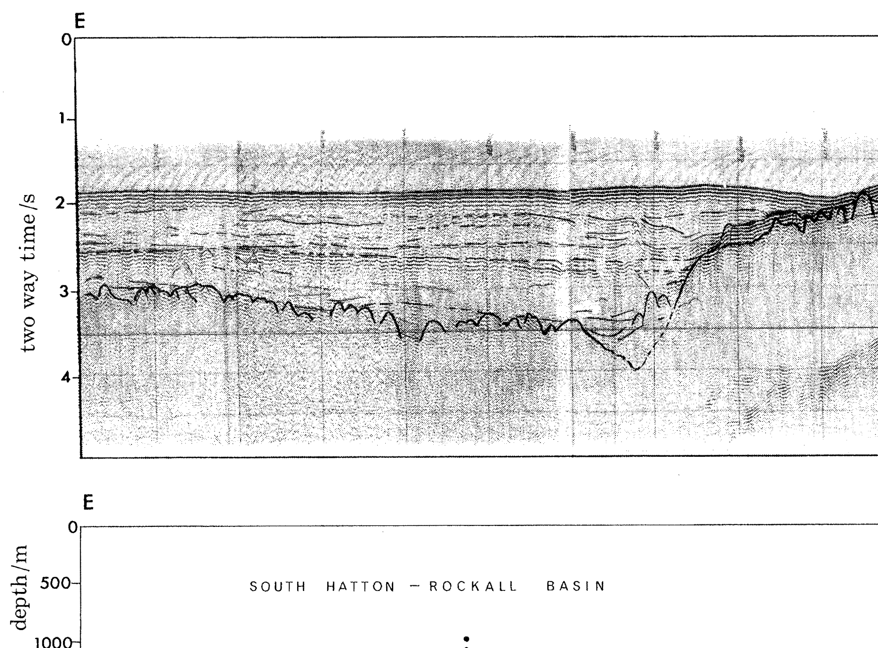
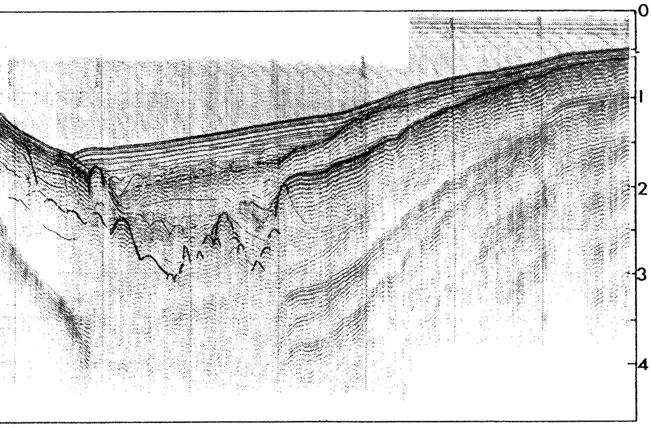
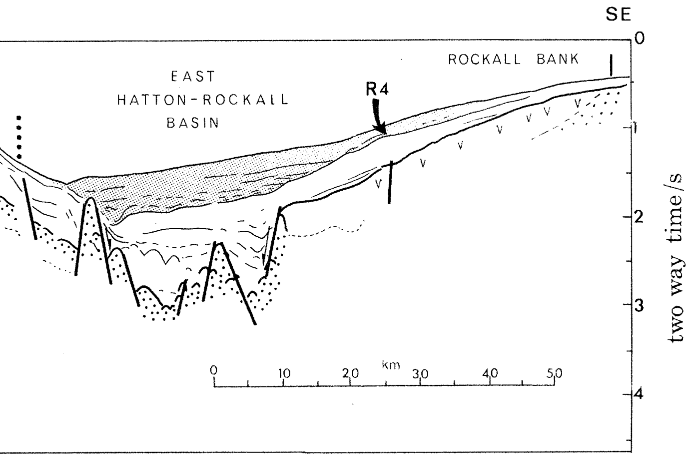


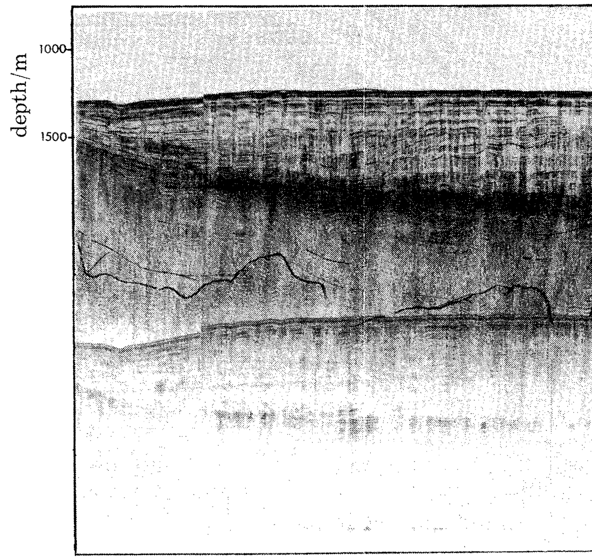
FIGURE 20. Seismic reflexion profile (20) across the central Hatton-Rockall Basin. Profile







n. Profile is located in figure 12.



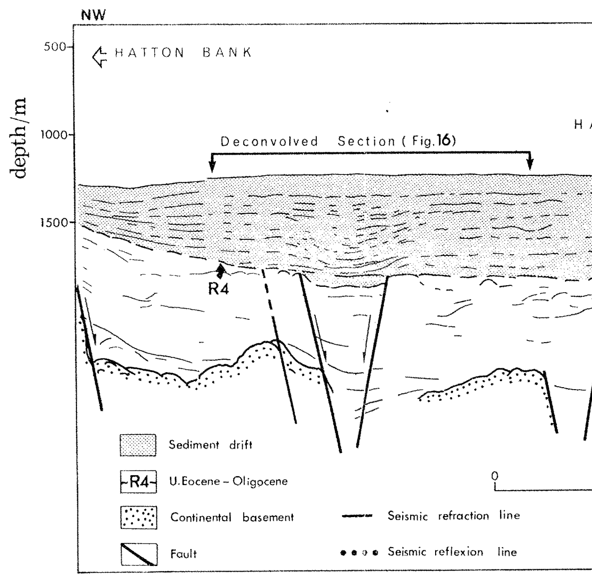
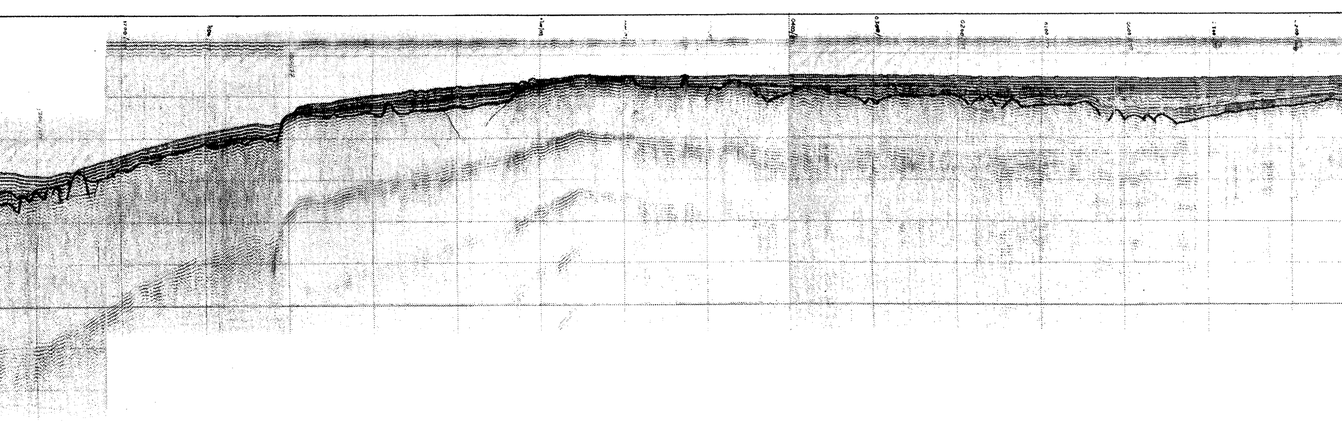
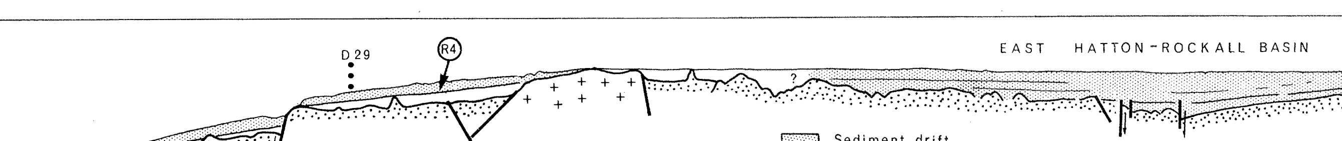
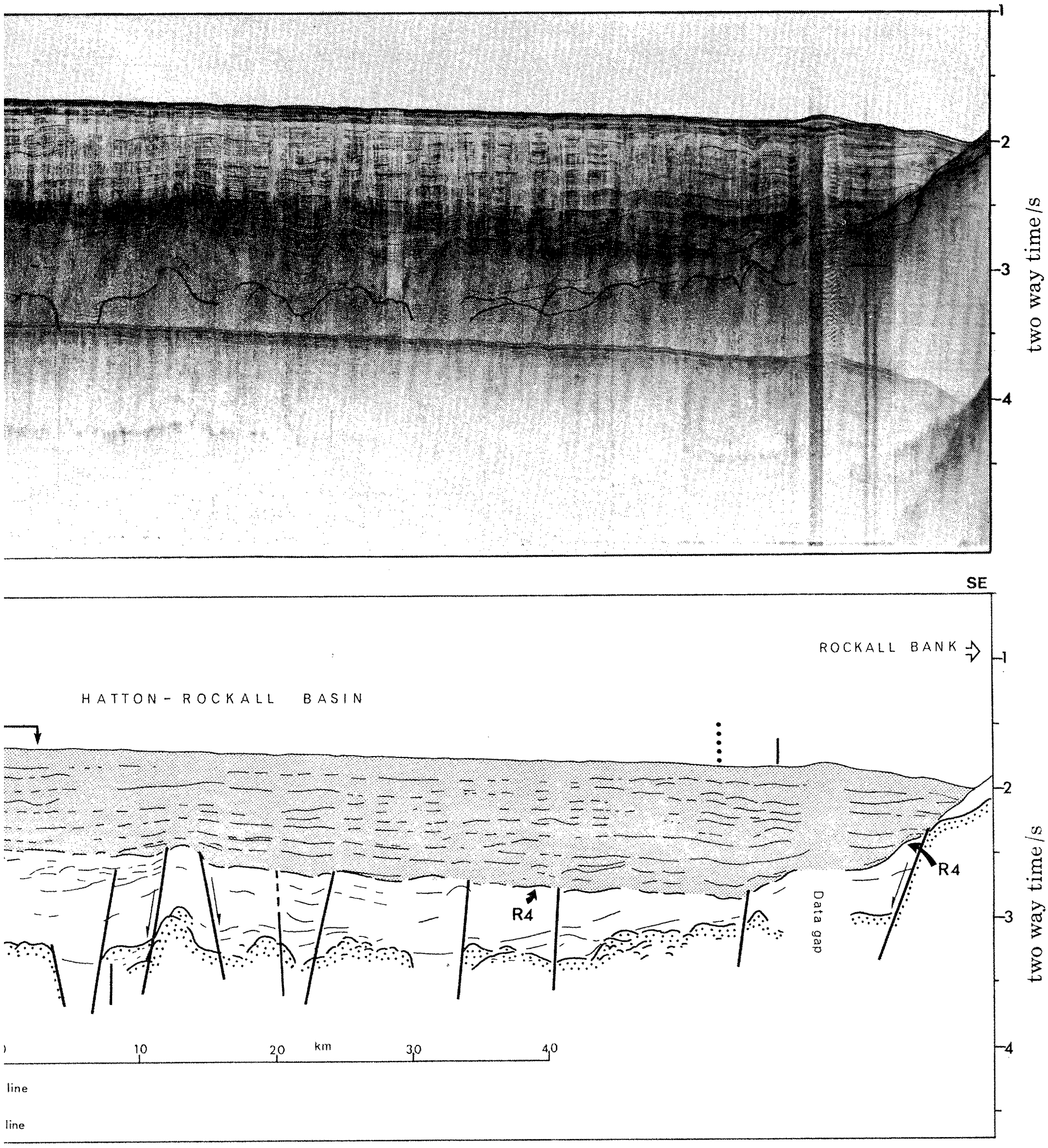


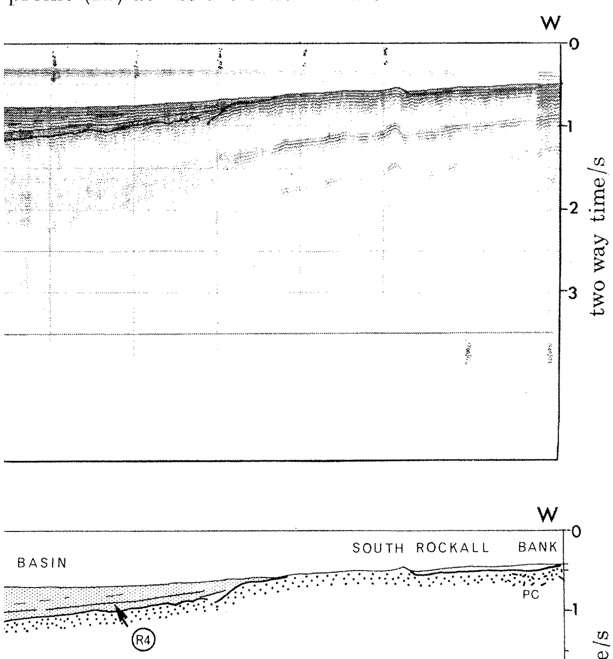
FIGURE 22. Seismic reflexion profile (2







profile (22) across the central Hatton-Rockall Basin. Profile is located in figure 12.



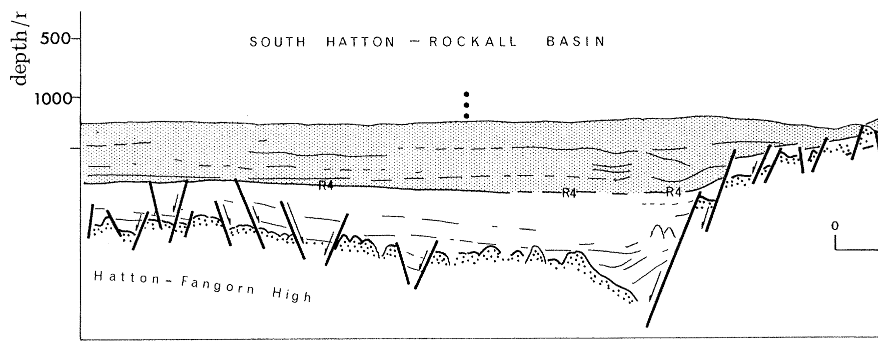
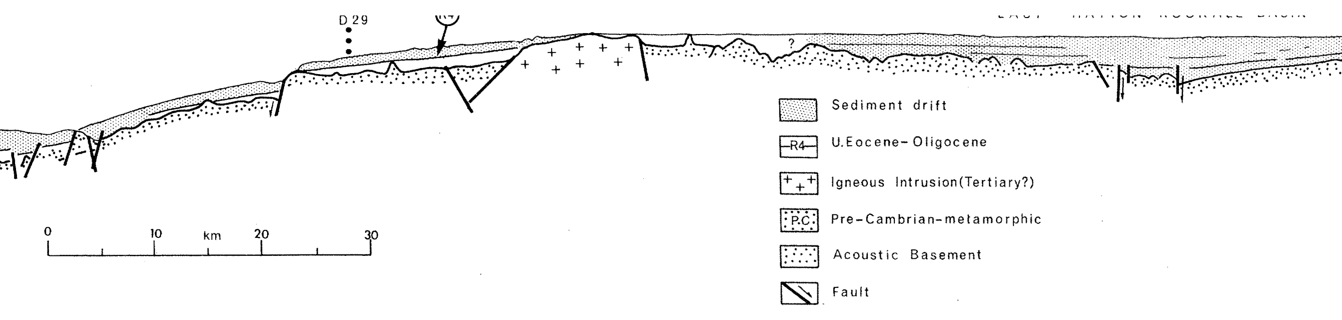


FIGURE 23. Seismic reflexion p



eflexion profile (23) between the south Hatton-Rockall Basin, Fangorn High and Rockall Bank. Profile is located in figure 1

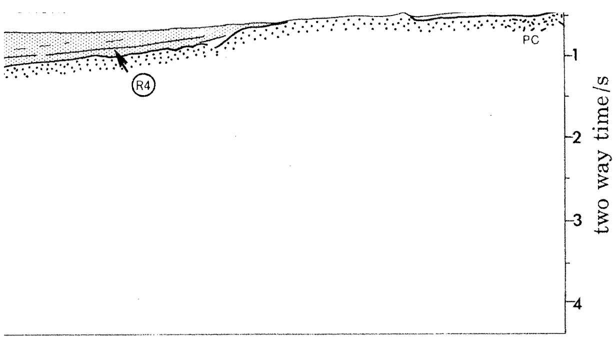


figure 12.

An unambiguous correlation between reflectors and lithologies could not be established because of the poor quality of the original seismic data and lack of interval velocity data. The ambiguities can be reduced by using the better quality deconvolved section (figure 16).

On this section, R4 is a discrete sub-horizontal reflector at about 0.9 s depth overlain with slight unconformity by a nearly horizontal sequence of strong reflectors of about 0.2 s thickness. These beds are laterally and vertically impersistent not occurring above the weak reflector '3'. Hyperbolae originating at the ends of these reflectors cause the 'crinkled' and diffuse aspect of R4 observed on unprocessed and highly exaggerated seismic records (cf. figure 22). The weak and discontinuous reflectors within the overlying reflector '3' sequence exhibit internal unconformities and are also unconformable with past and present depositional surfaces represented by '3', R4 and the seabed. These unconformities are due to differential deposition since

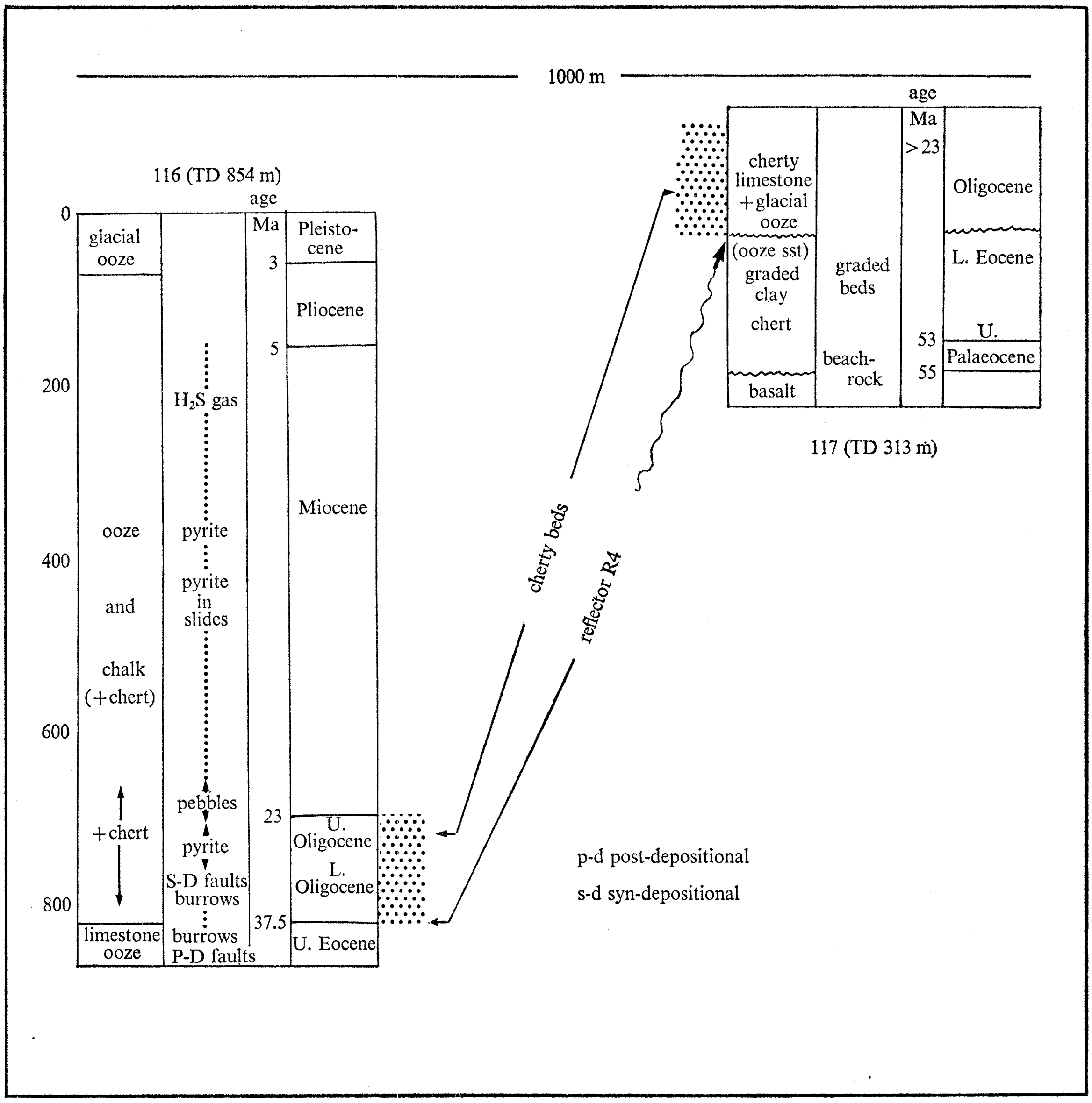


FIGURE 15. Summary stratigraphy of JOIDES Holes 116 and 117. Redrawn from Laughton, Berggren *et al.* (1972).

they occur in a sequence deposited on a sub-horizontal and undeformed surface. Note that the impersistent cuspate reflectors observed on the unprocessed record are the sub-surface continuation of the undulations observed in the seabed (cf. figures 16 and 22). In the post-R4 sequence, the contrast between the strong reflexions and the weak, discontinuous reflectors at reflector 3, indicates a substantial lithological change that can be plausibly correlated with the transition from ooze to chert in hole 116. If the boundary between the 140 m thick cherts and the underlying lithified Upper Eocene oozes is assumed to be R4, the interval velocity for the chert sequence is 2.8 km s^{-1} in close agreement with measured and pressure corrected velocities of 2.9 km s^{-1}. Reflector R4 may therefore be the top of the Upper Eocene lithified oozes.

This correlation is at variance with the previous identification of R4 as an intra-Oligocene unconformity (Laughton, Berggren *et al.* 1972). In hole 116, the intra-Oligocene unconformity was not cored but estimated to be a hiatus of *ca.* 10–15 Ma by using sedimentation rates derived from a graph of biostratigraphic age against depth. However, there are discrepancies in the ages and depths of the key foraminiferal and nannoplankton zones used to construct the graph, e.g. the Miocene/Oligocene boundary and the *Sphenolithus ciproensis* zone of the Late Oligocene. The foraminiferal and nannoplankton zonation of the Early Oligocene is also inconsistent. For example Perch-Nielsen (in Laughton, Berggren *et al.* 1972) assigns cores 23, 24 to 25 to the *Ericsonia obruta* zone while Bukry assigns core 25 to the overlying *Helicopontosphacra reticulata* zone. Identification of *Globigerina ampliapertura* in core 24 is stratigraphically consistent with the *Helicopontosphacra reticulata* zone but not the *Ericsonia obruta* zone. In part these inconsistencies reflect the problems inherent in high latitude biostratigraphy due to low diversity of the foraminifera and coccoliths (K. Perch-Nielsen, private communication). A corrected age against depth curve passes through all the Oligocene dates indicating reduced sedimentation rather than a major hiatus.

A major Oligocene hiatus is thus not supported by the data and the lithological correlation suggests R4 lies at the base of the Oligocene. At site 117, the unconformity between lower Eocene dark green clays and the cherts undoubtedly corresponds to R4. Equally, there can be little doubt that reflector R4 and the overlying cherts are continuous between sites 116 and 117 (see figure 14). However, the absence of the Late Eocene at site 117 must be reconciled with the seismic data to define the nature of R4 within the Hatton–Rockall Basin. The seismic profiles show the cherts within the post-R4 sequence are largely conformable with R4 but the angular discordance of R4 increases towards the basin margins. At the basin margins, R4 is thus an unconformity that decreases basinward and passes into a conformable sequence. A comparable onlapping relation may exist between the Late Eocene and Early Eocene. Two other unconformities can be inferred in the post-R4 series. Between sites 116 and 117; the Quaternary and Neogene oozes overstep the Oligocene cherts (within which overstep also occurs) to rest on the Lower Tertiary igneous or Precambrian basement of Rockall Bank and the Late Eocene may pinch out against the basement. The onlap and unconformities are consistent with deposition in an intermittently subsiding basin (cf. Laughton, Berggren *et al.* 1972).

Supporting evidence for the 37.5 Ma age of R4 is also given by its pinch out on oceanic basement dated at 37 Ma and by the presence of Upper Oligocene soft chalk (26–30 Ma) at 11 m in a core taken 0.09 s above R4 (Ruddiman 1972). Drilling of a similar reflector in the Labrador Sea has shown 'the centre of the energy of the reflexion' to be 35 Ma.

From these data a tentative stratigraphy for the Hatton–Rockall Basin, east flank of the

Reykjanes Ridge and southwest margin of Rockall Plateau is given in figure 17. The greatest uncertainties are the ages of reflectors observed in the pre-R4 series of the Hatton–Rockall Basin, Rockall Trough and southwest margin of Rockall Plateau. In the latter case, a reflector called 'X' is absent above oceanic basement younger than 60 Ma (Scrutton & Roberts 1971; Ruddiman 1972) but is present in mid-section above 60–76 Ma basement (figure 26). Sedimentation rates of *ca.* 3.1 cm per 1000 years and a V_p of 2.3 km s^{-1} (site 117; Ewing & Ewing 1958) indicate an age for 'X' of *ca.* 63 Ma that suggests a relation to the 60 Ma formation of oceanic crust between Greenland and Rockall Plateau (Vogt & Avery 1974). A comparable reflector (5*a* of figure 16) in the Hatton–Rockall Basin may have a related origin.

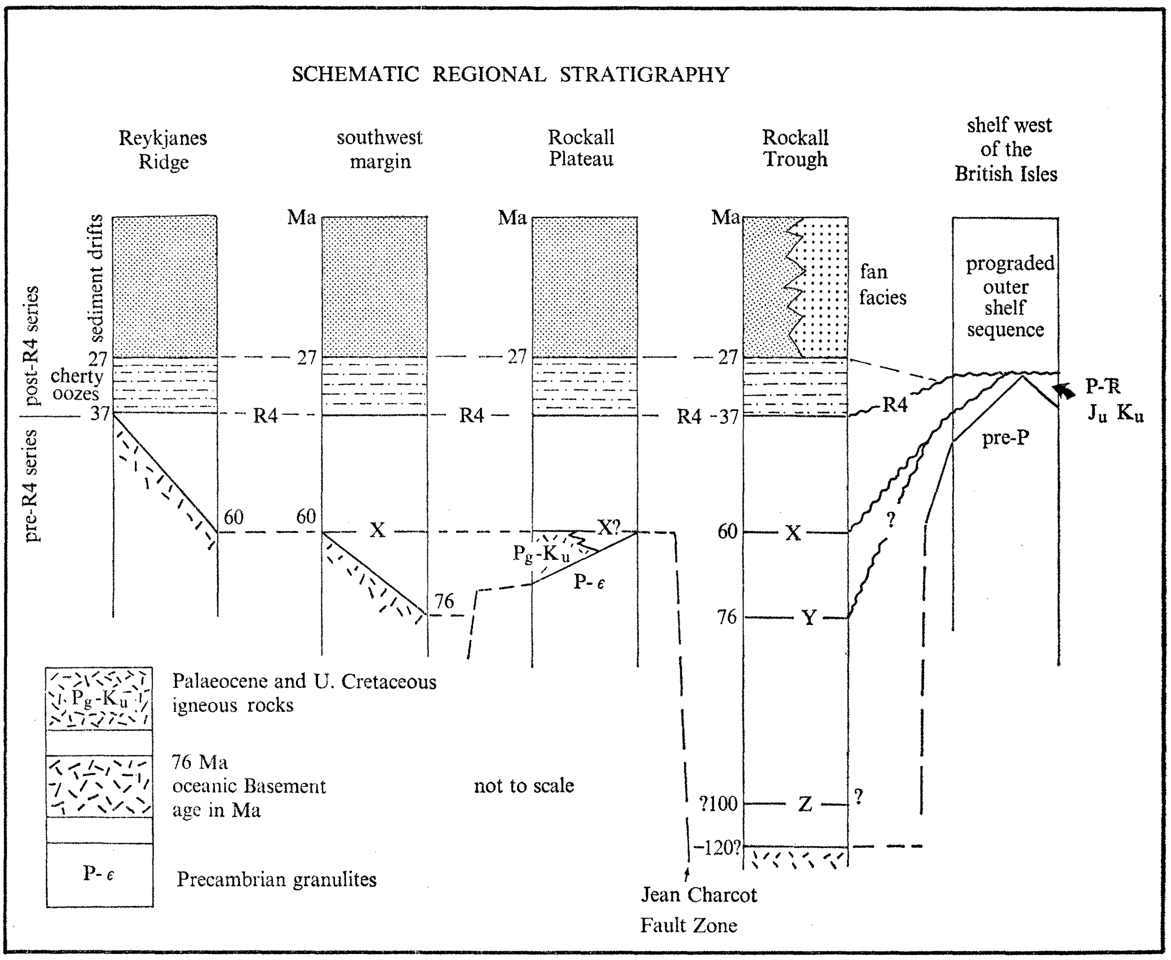


FIGURE 17. Schematic regional stratigraphy of Rockall Plateau and adjacent areas.

(ii) *The basement*

Rockall Bank, George Bligh Bank, Hatton Bank, Lorien Bank, Fangorn Bank and Edoras Bank are thinly covered in sediments that rarely exceed 0.2 s in thickness and rest on a flat-lying acoustic basement characterized by many hyperbolic echoes.

Dredging and drilling of outcrops south of 57° 00′ N on Rockall Bank has yielded Laxfordian (*ca.* 1700 Ma) and Grenvillian (*ca.* 1000 Ma) granulites (Roberts *et al.* 1972; Roberts *et al.* 1973*a*; Miller *et al.* 1973). These localities are separated by an ENE–WSW trending magnetic

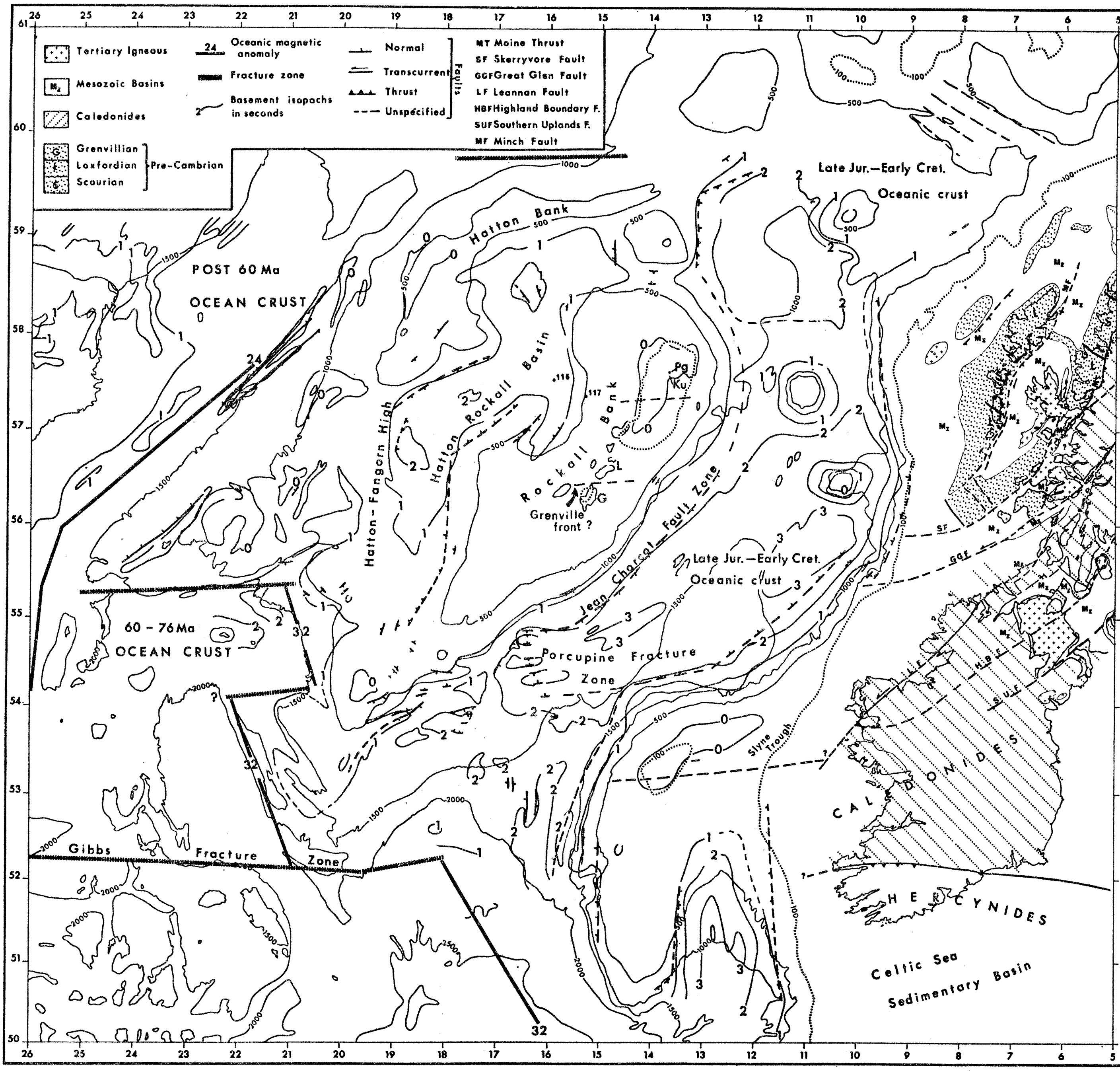
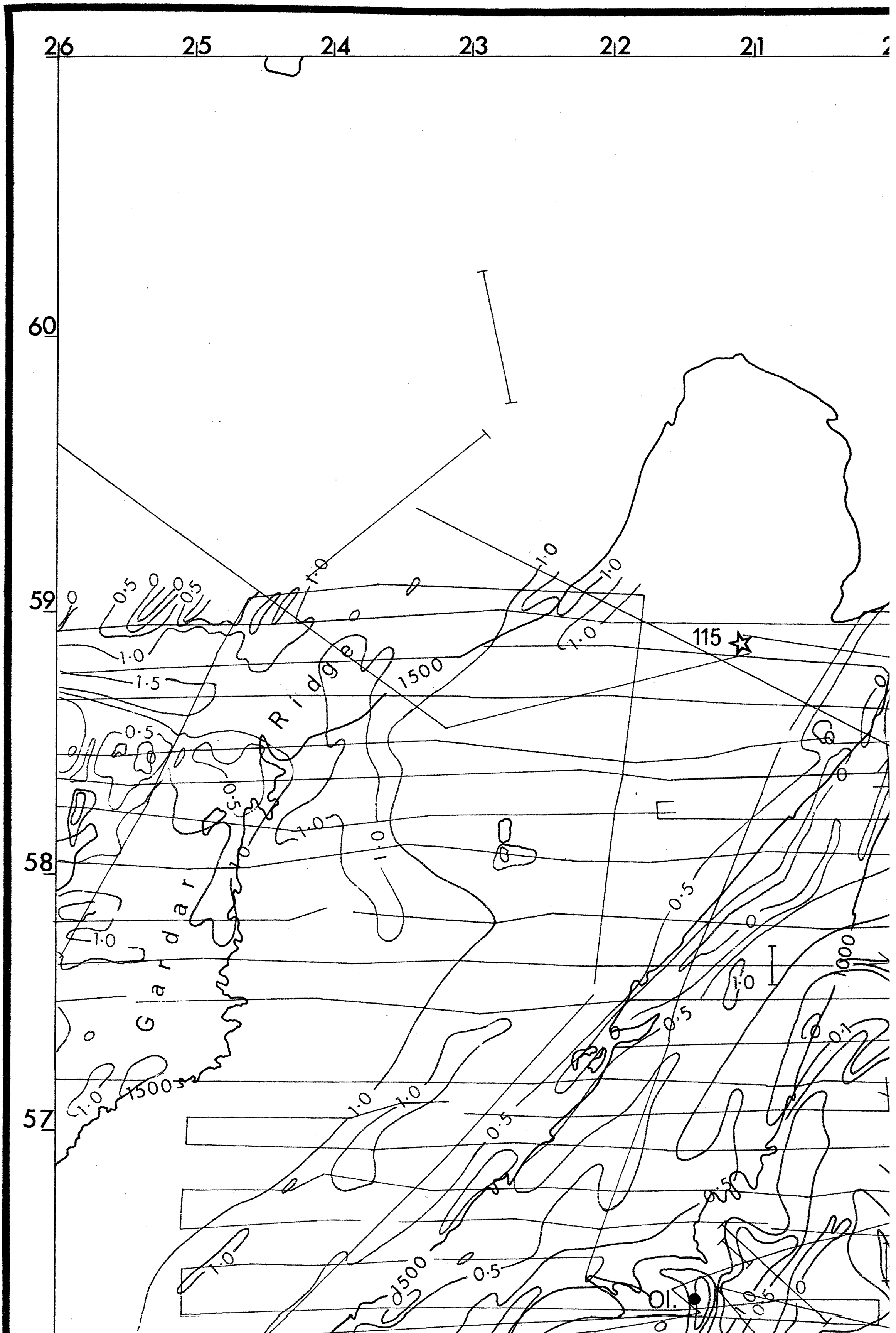
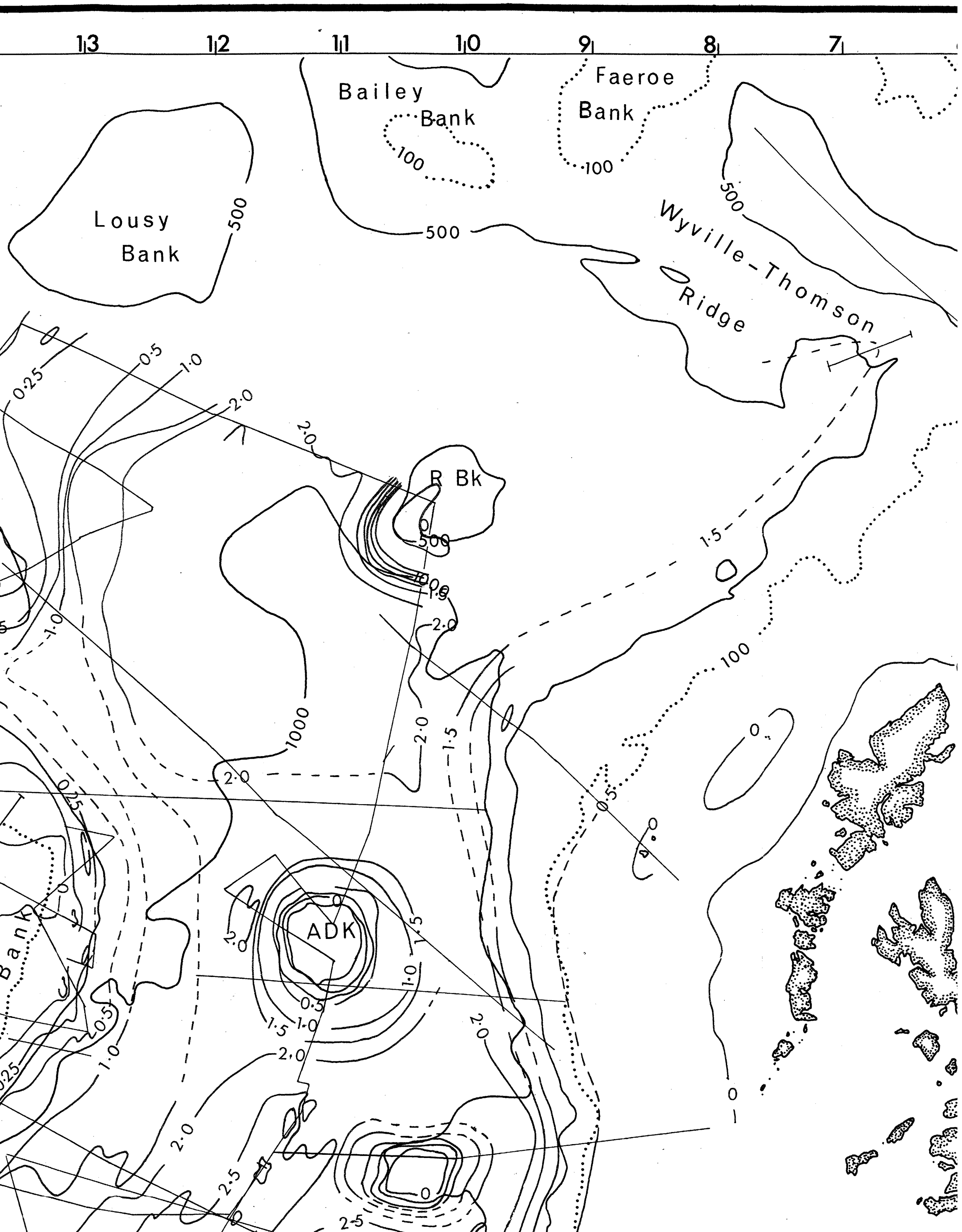


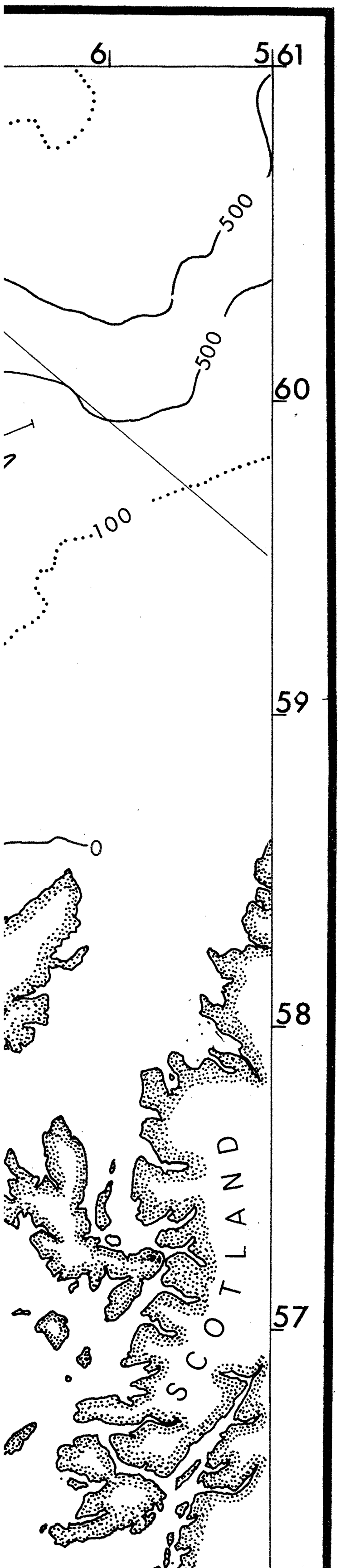
FIGURE 21. Main structural elements of the Rockall Plateau, Rockall Trough and continental margin west of the British Isles (after Dalziel 1969; Eden, Wright & Bullerwell 1971; Roberts 1974). Depths in fathoms.

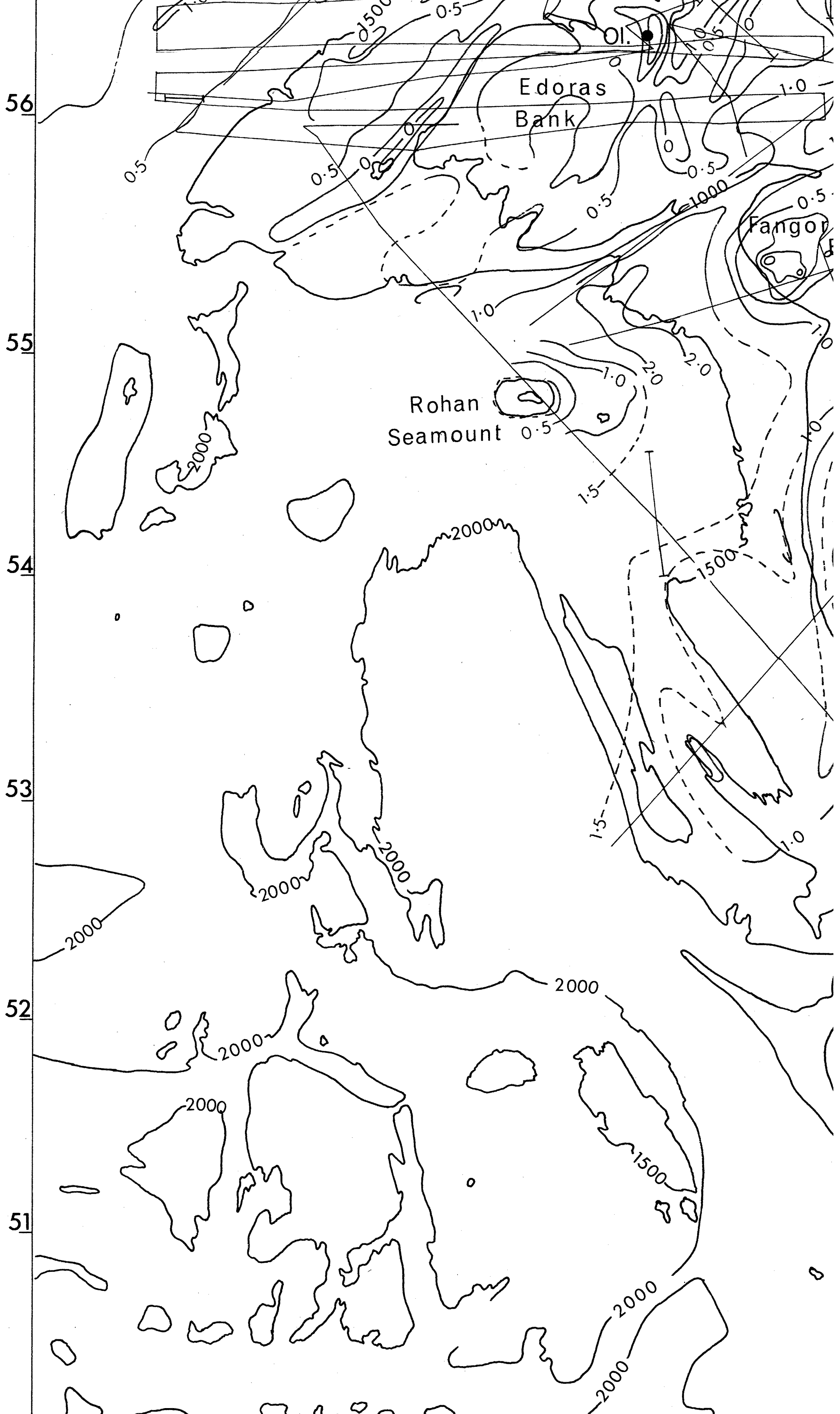
lineament that may represent the eastward extension of the Grenville front (figure 21). Rockall Island consists of 52 ± 9 Ma aegirine-granite situated within an intrusive complex composed of Late Tertiary and Upper Cretaceous acid and basic igneous rocks (Sabine 1965; Miller & Mohr 1965; Roberts 1969; Jones, Mitchell, Shido & Phillips 1972; Hawkes, Merriman, Harding & Harrison 1973; Roberts *et al.* 1974). Widespread extrusive activity is indicated by the pre-Upper Palaeocene subaerial basalt penetrated in hole 117 on the west side of Rockall Bank (Laughton, Berggren *et al.* 1972).


26
25
24
23
22
21
60
59
58
57
115
Gardar Ridge
1500
1000
01.

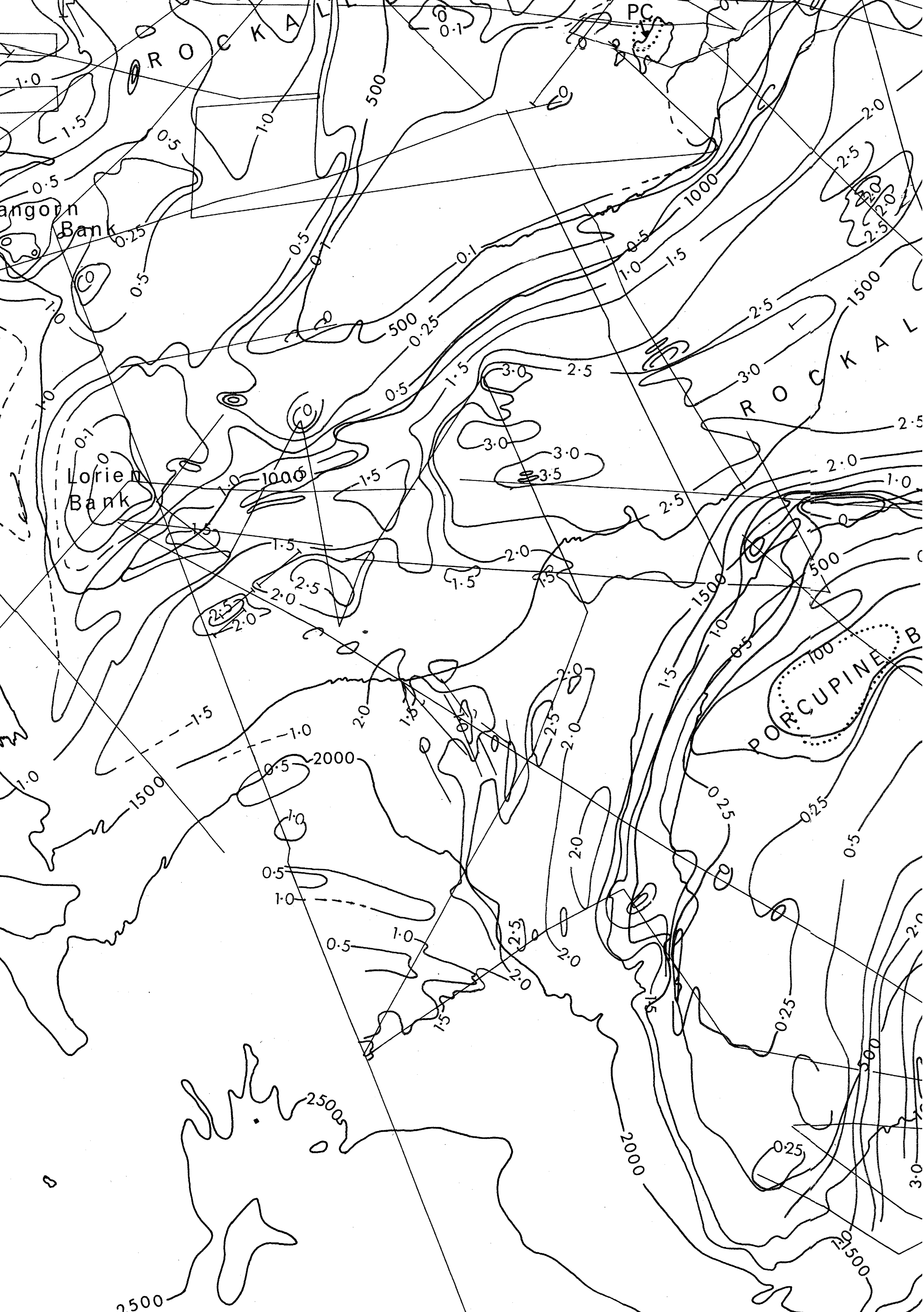
20
19
18
17
16
15
14
1000
1500
500
Hatton Bank
G B Bk
Rockall
Paleo.
116
117
Basin
Hatton - Rockall
PLATEAU
Rockall Bank
PC
ROCKALL

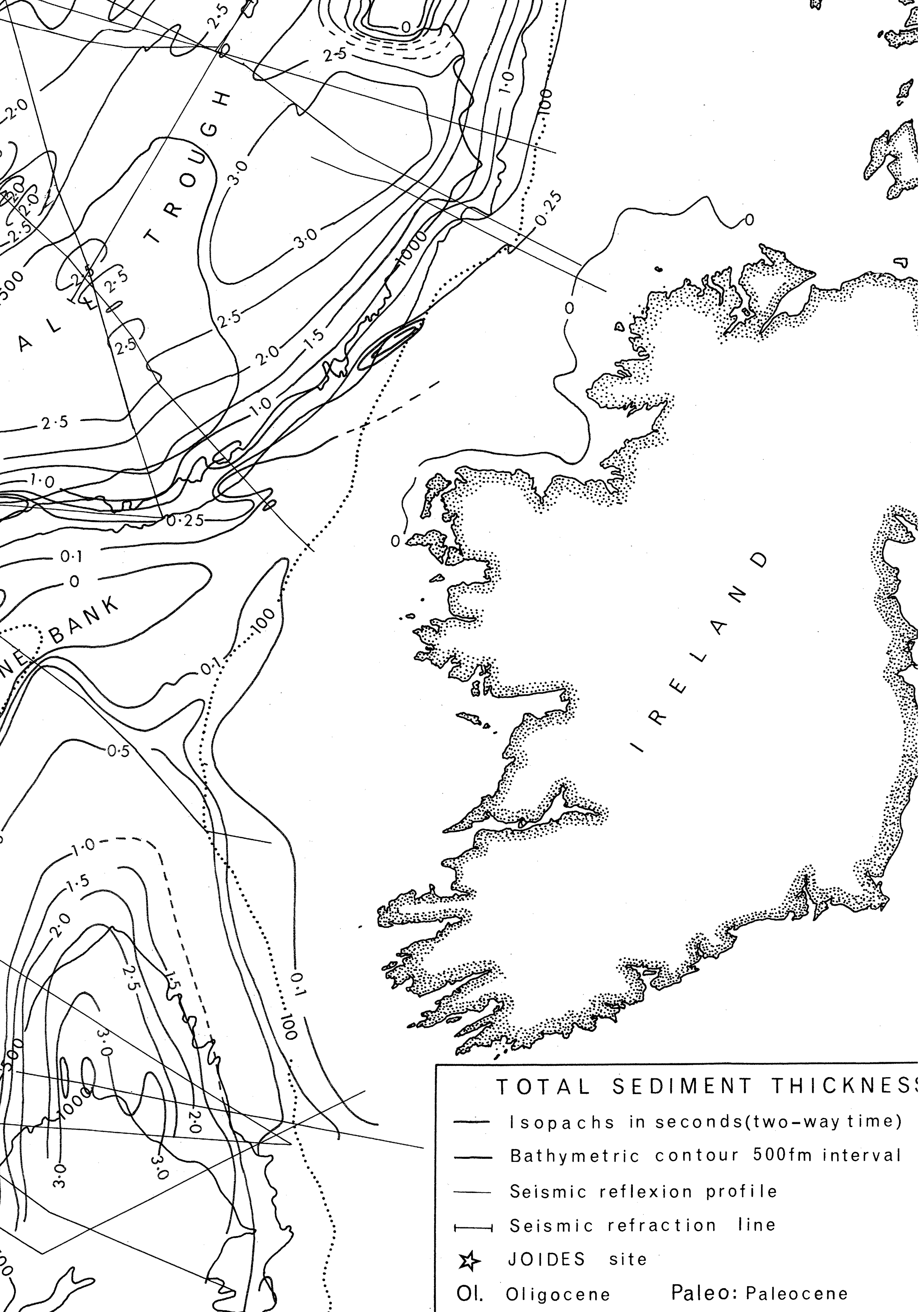

13
12
11
10
9
8
7
Bailey
Bank
100
Faeroe
Bank
100
Lousy
Bank
500
500
500
Wyville-Thomson
Ridge
0.5
1.0
0.25
2.0
2.0
R Bk
0
500
1000
1.5
2.0
1.5
100
1000
2.0
1.5
0
0.5
0
0.25
-1.0
2.0
ADK
0
1.5
1.0
0.5
1.5
1.0
2.0
2.0
0.5
1.0
0.25
2.0
2.5
0
0
2.5
Bank
0

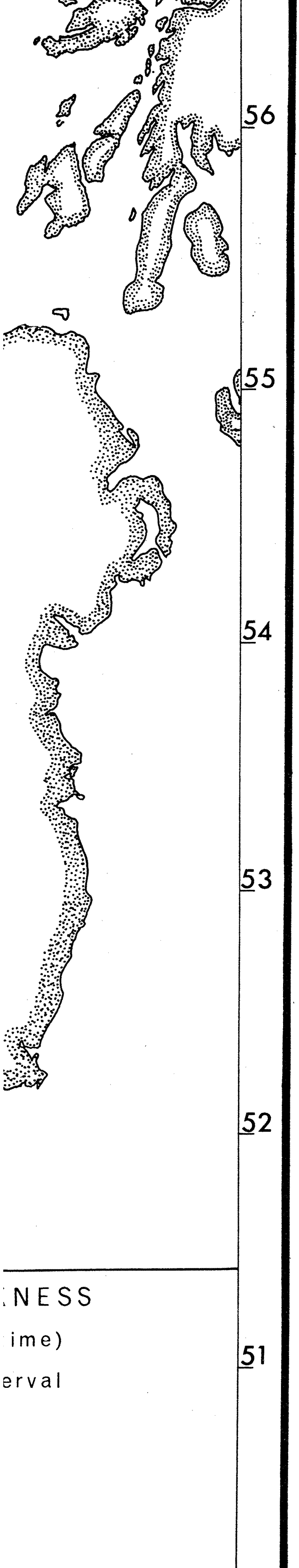

6
5
61
500
500
60
100
59
0
58
SCOTLAND
57



56
55
54
53
52
51
Edoras
Bank
Rohan
Seamount
Fangor
OI.
1500
1000
2000
0·5
0
1·0
1·5
2·0


ROCKALL
PC
Lorien Bank
ROCKAL
PORCUPINE B


TROUGH
BANK
IRELAND
TOTAL SEDIMENT THICKNESS
Isopachs in seconds(two-way time)
Bathymetric contour 500fm interval
Seismic reflexion profile
Seismic refraction line
JOIDES site
Ol. Oligocene
Paleo: Paleocene


56
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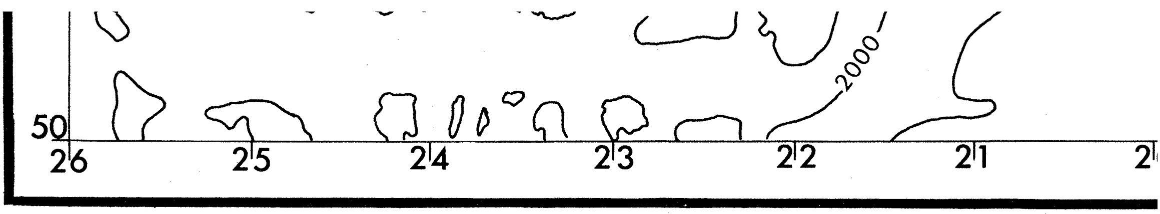
52
.NESS
ime)
erval
51


2000
50
26
25
24
23
22
21

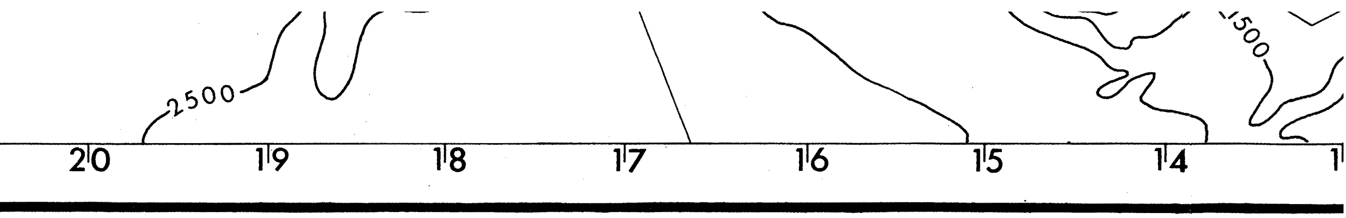
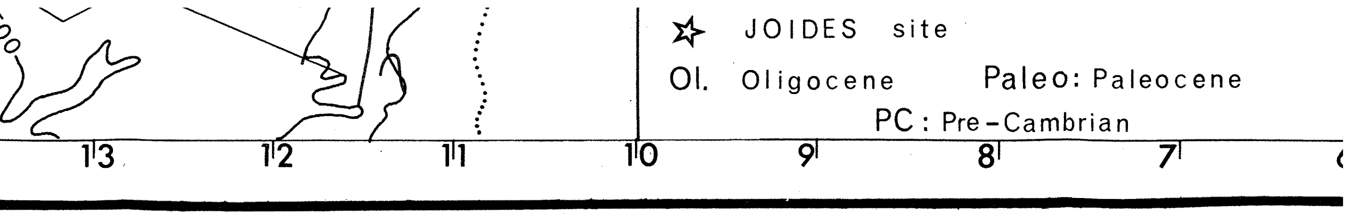
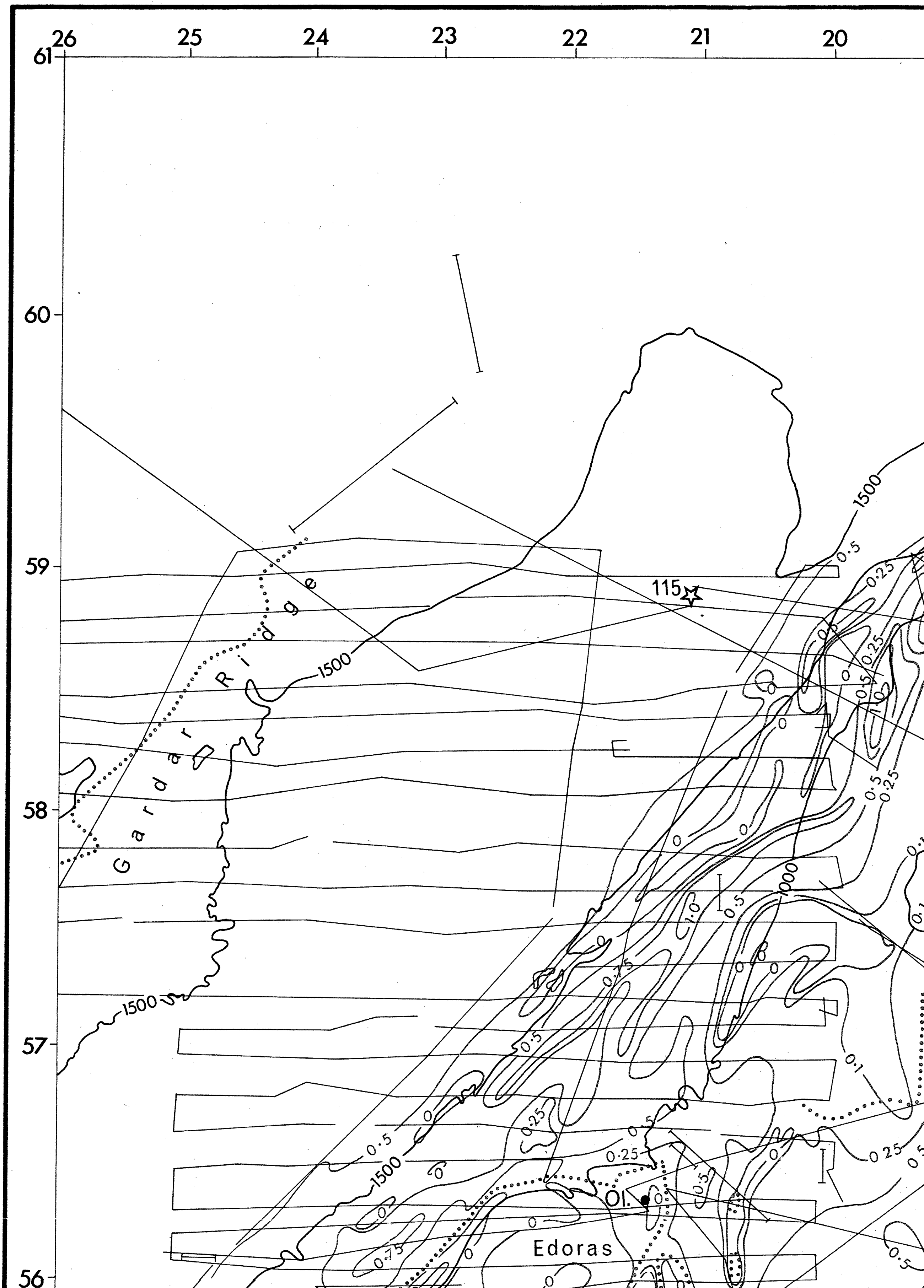


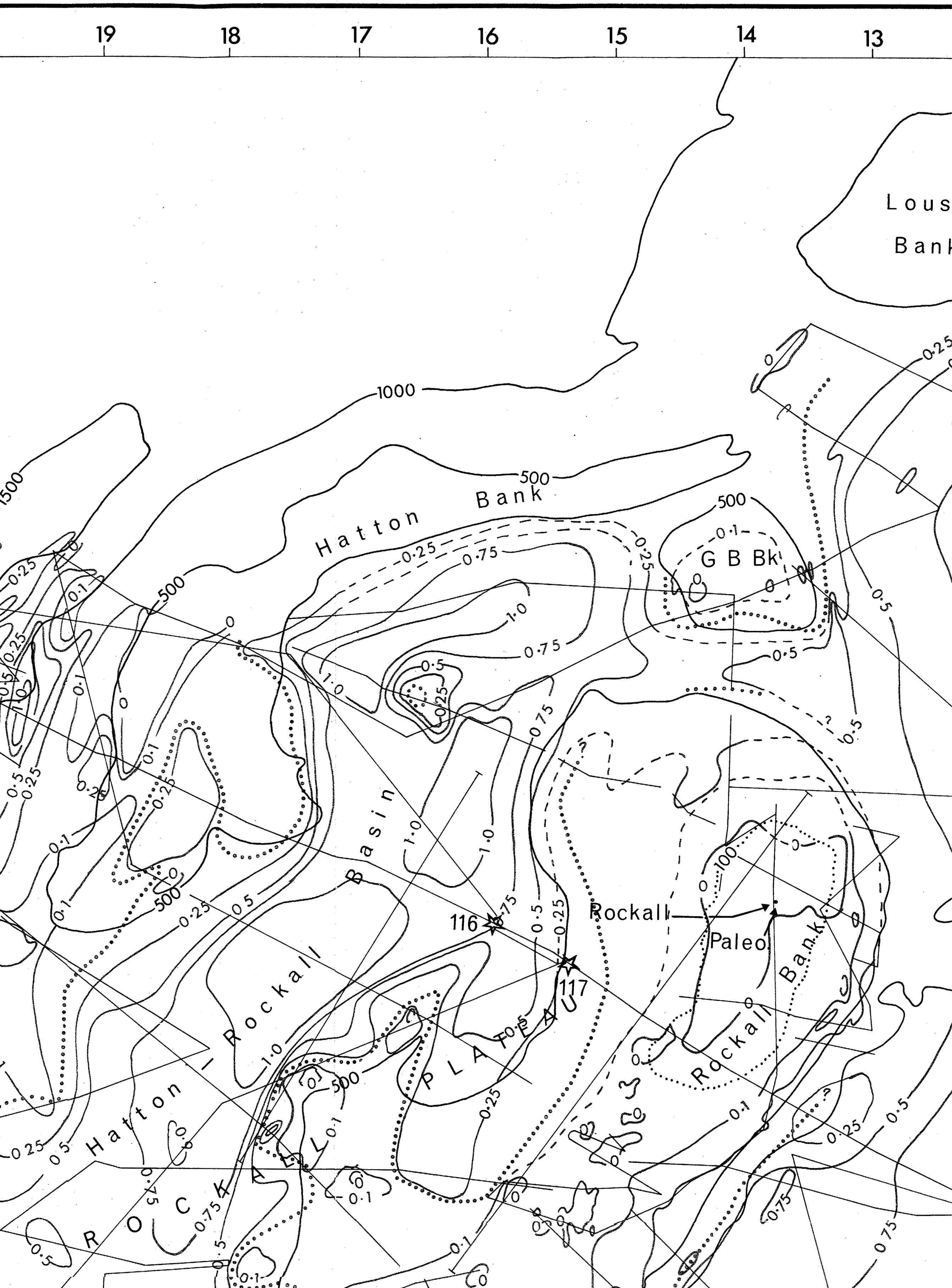
FIGURE 18. Isopachs on basement in two way time below seabed.

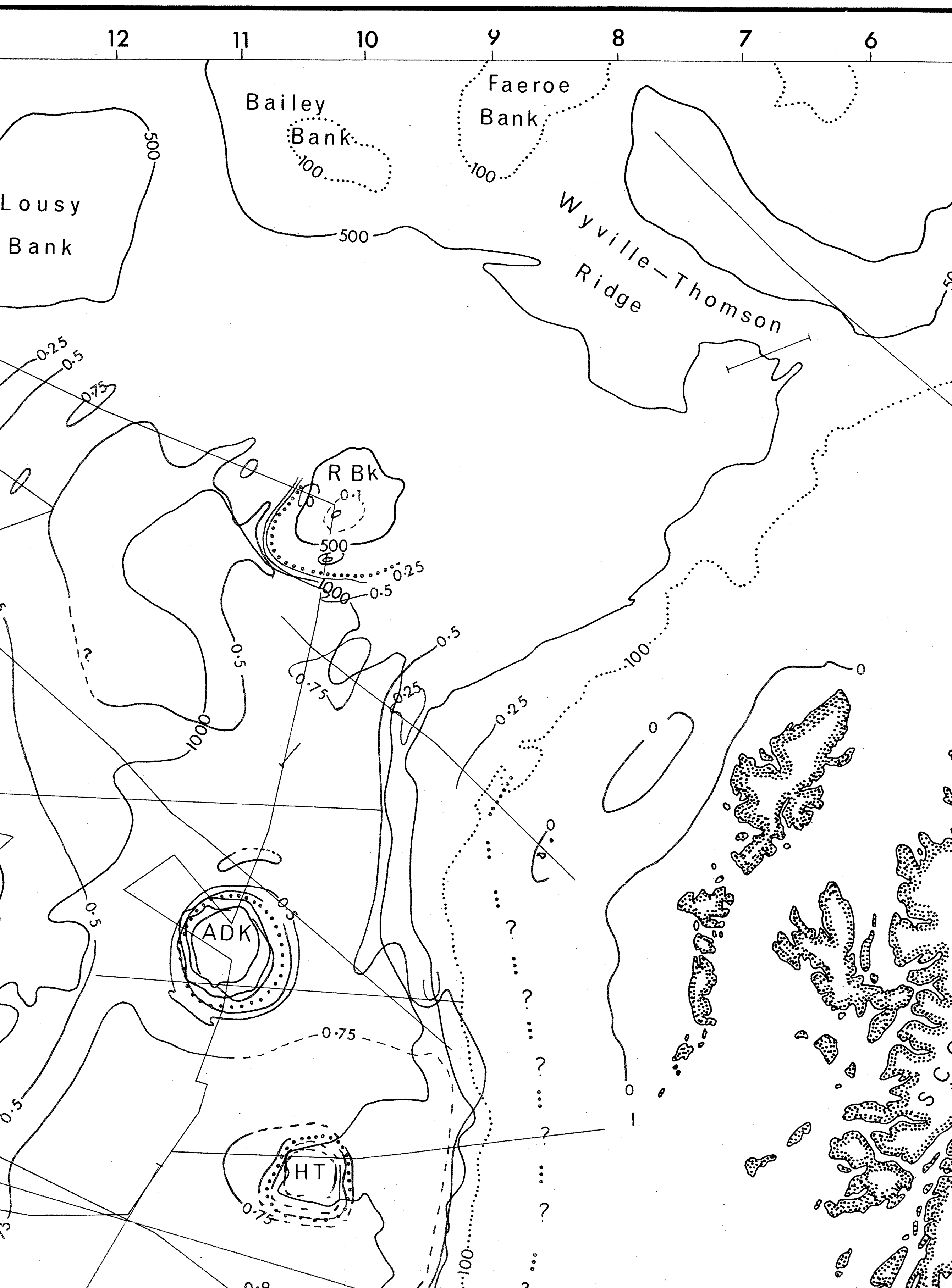

JOIDES site
Ol. Oligocene
Paleo: Paleocene
PC: Pre-Cambrian
13
12
11
10
9
8
7

50

6 5


26
25
24
23
22
21
20
61
60
59
58
57
56
G a r d a r R i d g e
1500
1000
115
Edoras
0.5
0.25
0.75
0.1
0


19
18
17
16
15
14
13
Lous
Bank
1000
500
Hatton Bank
G B Bk
Rockall
Paleo
Basin
Rockall Bank
116
117
PLATEAU
Hatton Rockall
ROCKALL


12
11
10
9
8
7
6
Bailey Bank
Faeroe Bank
Lousy Bank
Wyville–Thomson Ridge
R Bk
ADK
HT
500
100
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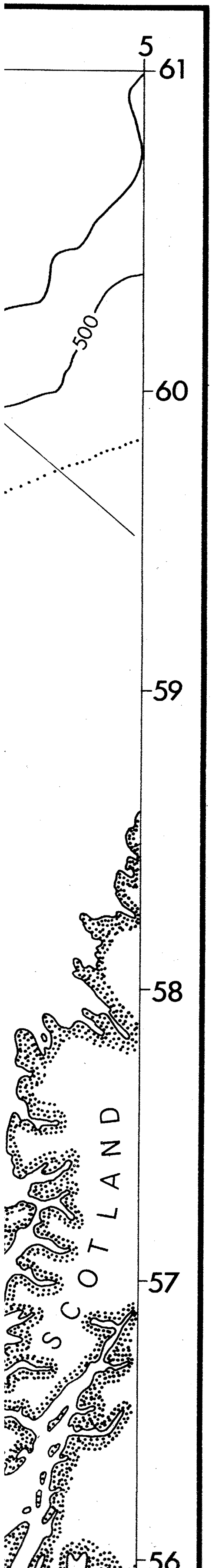
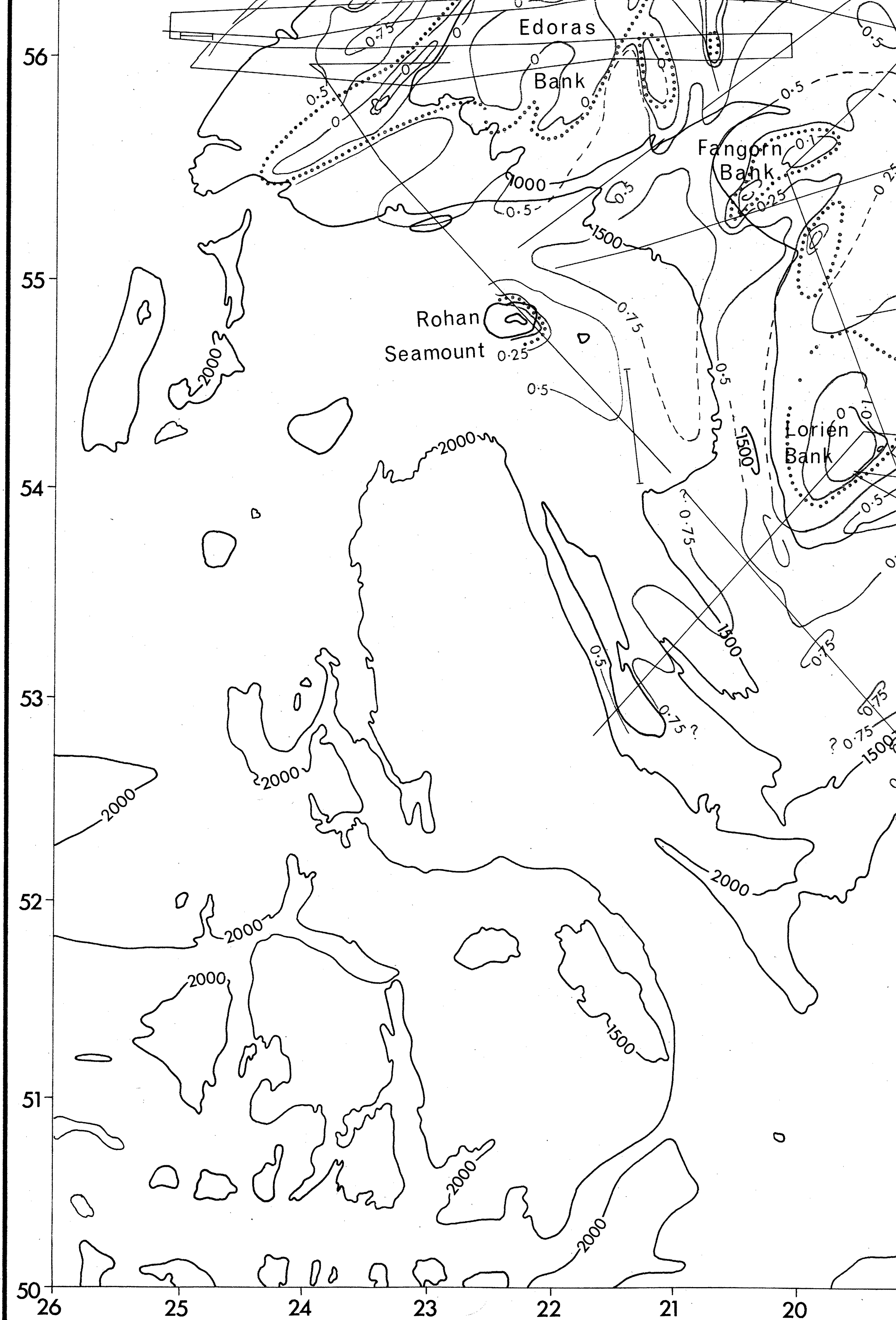
0·1
0·25
0·5
0·75
0
?


5
61
500
60
59
58
SCOTLAND
57
56



Edoras
Bank
Fangorn
Bank
Rohan
Seamount
Lorien
Bank
1000
1500
2000
0.5
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50
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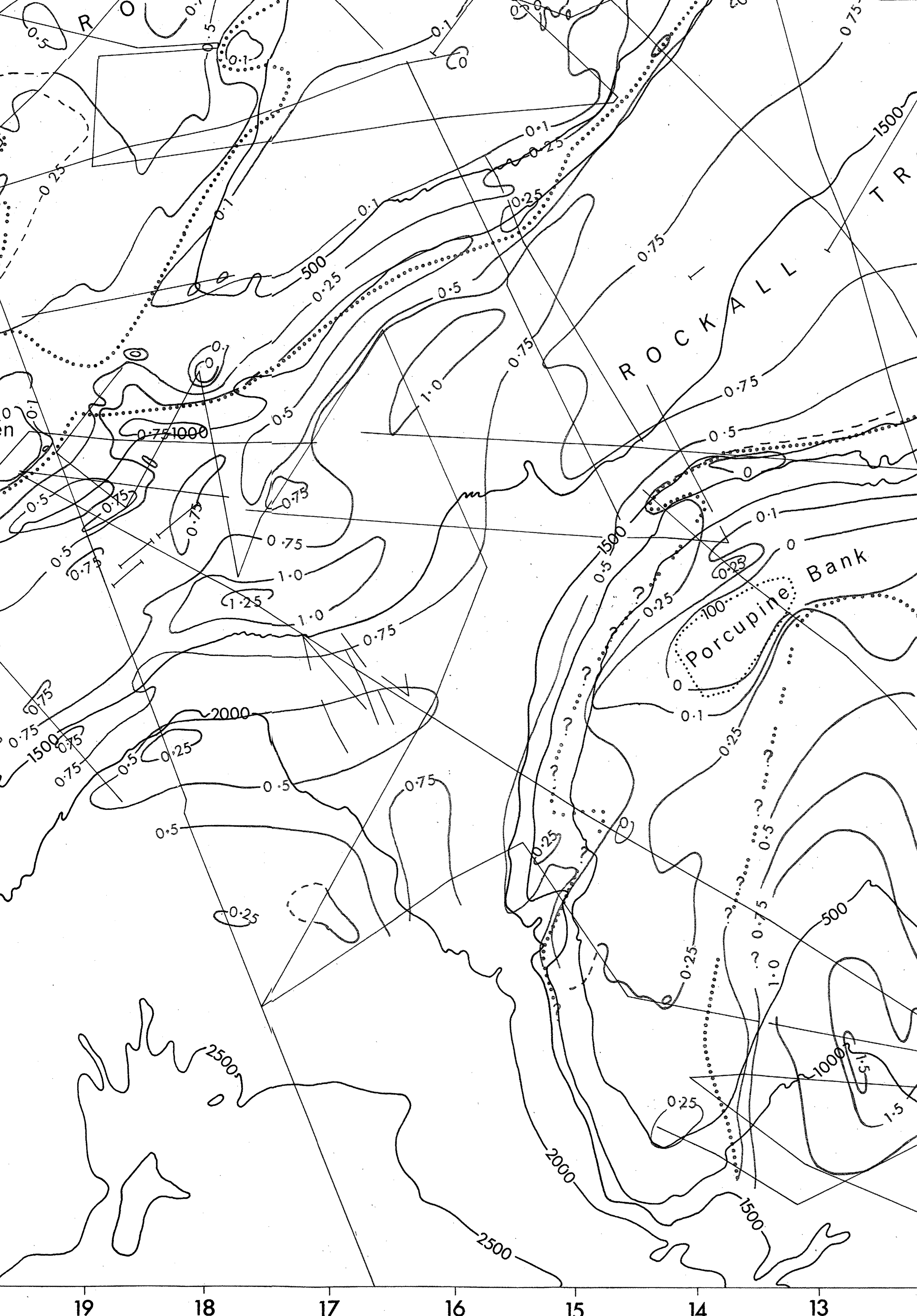

R O
ROCKALL TR
Porcupine Bank
Porcupine
Bank
en
100
500
1000
1500
2000
2500
0·1
0·25
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0·75
1·0
1·25
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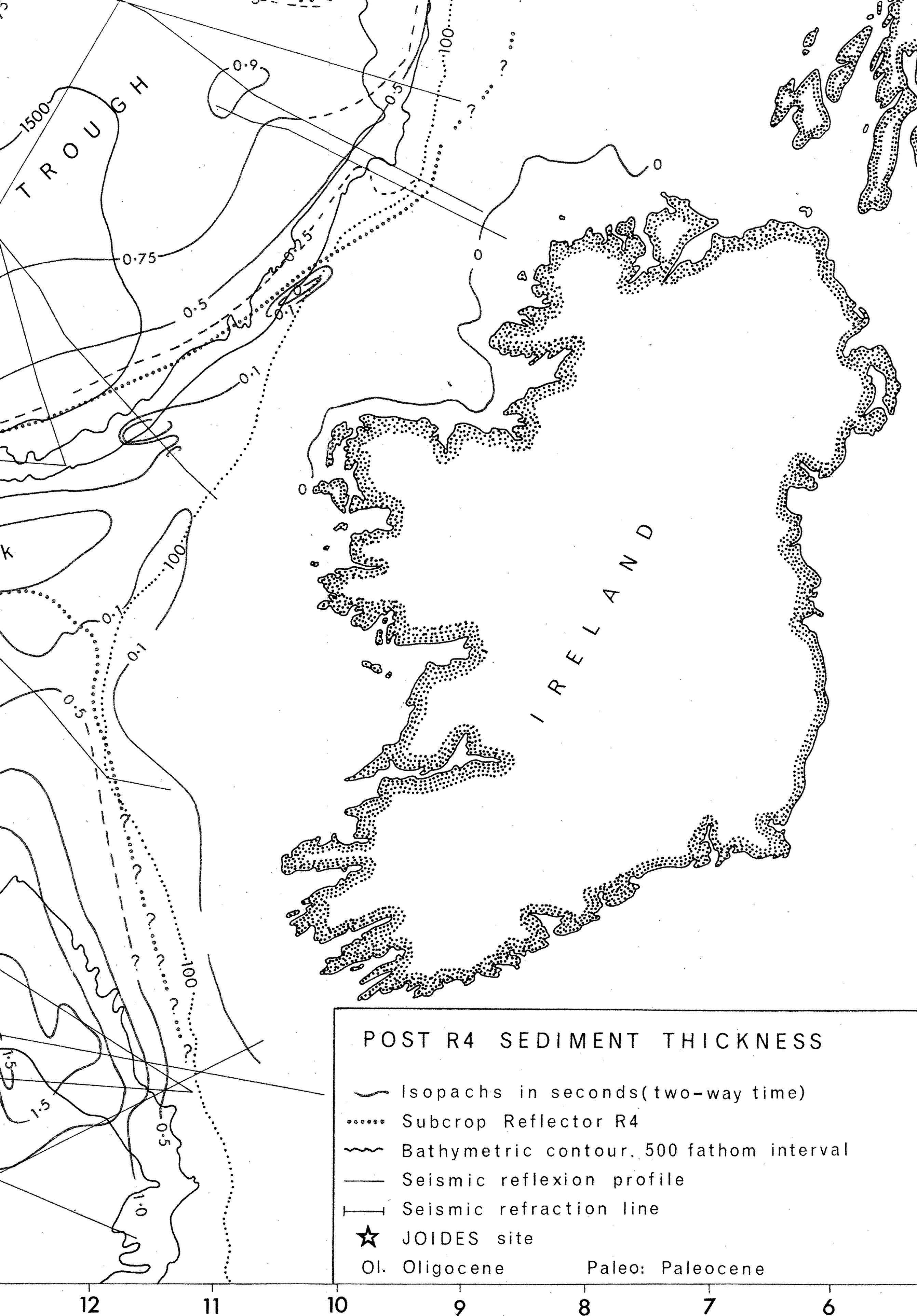
13


TROUGH
IRELAND
1500
0.9
0.75
0.5
0.25
0.1
0
100
1.5
1.0
POST R4 SEDIMENT THICKNESS
Isopachs in seconds (two-way time)
Subcrop Reflector R4
Bathymetric contour, 500 fathom interval
Seismic reflexion profile
Seismic refraction line
JOIDES site
Ol. Oligocene
Paleo: Paleocene
12
11
10
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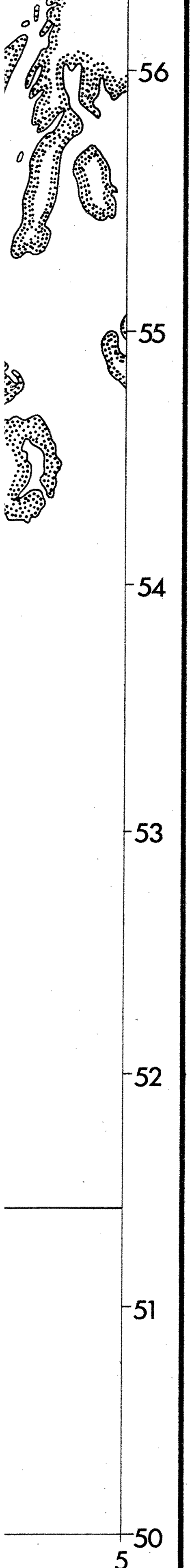
6


56
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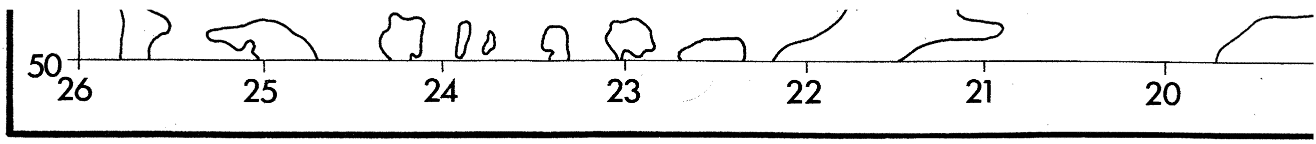
53
52
51
50
5


50
26
25
24
23
22
21
20

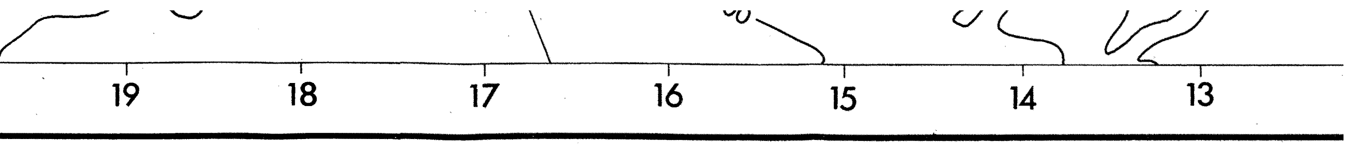
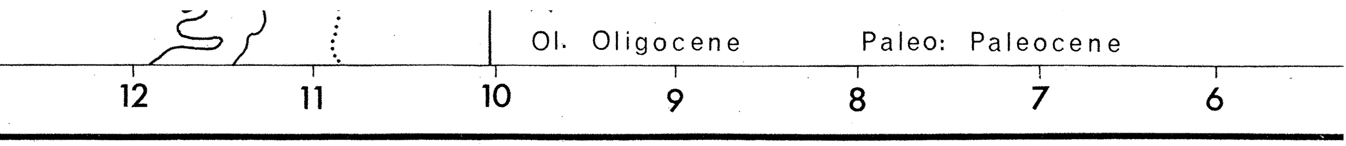
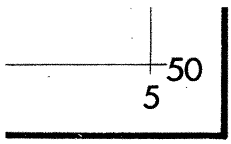


FIGURE 27. Isopachs on R4 in two way time below the seabed.


Ol. Oligocene
Paleo: Paleocene
12
11
10
9
8
7
6


50
5

Apart from these data, there is no direct evidence for the age and composition of the basement. Seismic refraction stations marked near Rockall Island show a 0.4 km sub-bottom layer with a V_p of 3.80 km s^{-1} in contrast to a 2.3 km s^{-1} layer with a V_p of 4.17 km s^{-1} layer further south (Scrutton & Roberts 1971). Comparable velocities of 3.9 km s^{-1} observed in the Faeroes and Iceland (Palmason 1967), the geological data and magnetic surveys suggest basic lavas may cover the northern Rockall Bank and that the underlying granulites form inliers although comprising much of the basement to the south. Scrutton (1972) has interpreted a +80 mGal free-air gravity anomaly at 58° 10′ N 14° 30′ W as another intrusive centre. A pre-lava fault at 57° 10′ N (figure 21) is shown by a step in the basement and also by E–W trending magnetic and gravity lineations (Scrutton & Roberts 1971; Roberts & Jones 1974).

The nature of the basement beneath George Bligh Bank is not known. 'Blackstones' reported from Hatton Bank may be derived from igneous centres shown by +100 mGal free-air gravity anomalies (Donovan 1968; Scrutton 1972). A 3.0 km s^{-1} refractor on the southern Hatton Bank (station CR45, Gaskell *et al.* 1958) may also indicate basic extrusives or Mesozoic sediments although it is absent 60 km further west (cf. stations CR44/44A, Gaskell *et al.* 1958). Fangorn Bank is composed of highly irregular outcropping basement associated with a 1200 nT magnetic anomaly and may be composed of basic igneous rocks (Laughton, Berggren *et al.* 1972). Beneath the Hatton–Rockall Basin the acoustic basement is the top of a 4.54 km s^{-1} refractor (Scrutton 1972). This velocity is higher than 3.80 km s^{-1} velocity associated with the basic igneous rocks of Rockall Bank and may indicate their absence beneath the Basin.

The isopachs on basement (figure 18) define an irregular basin oriented NE–SW and measuring 400 km × 100 km at the 1.25 s isopach. In general, the structure is either a broad downflexure with superposed faulting or a broad graben (figures 14, 19, 20 and 22). Much of the irregularity reflects the intersection of several different marginal fault trends. Correlation of individual faults from profile to profile is not possible but the isopachs and individual profiles offer some insight into their variable development.

In the northern Basin, ENE–WSW trending isopachs parallel the axis of Hatton Bank. Little faulting is present and the basin margin can be regarded as a dipslope of this escarpment. West of George Bligh Bank, a N–S fault zone downthrows antithetically faulted basement by 1.25 s (figure 19). This fault zone may be intersected by the E–W graben (figure 24) separating George Bligh Bank and Rockall Bank. In the northwestern basin, faulting is present around a small horst but dies out towards Hatton Bank. However, impersistent reflectors below acoustic basement may indicate a thicker sedimentary sequence (figure 19).

At 59° N 18° W, the Hatton–Rockall Basin trends NE–SW and parallel to the west margin of Rockall Plateau. The west margin of the Basin is steeply faulted in contrast to the downflexed and locally faulted east margin (figure 14). Further south, the main basin is underlain by a 60 km wide graben divided from the 30 km wide asymmetric graben underlying the East Hatton–Rockall Basin (figure 40) by a NE–SW trending horst (figure 20). The horst must plunge northward or be terminated by a cross-fault since it is absent north of 57° 20′ N (figure 18). Perhaps significantly this point lies along the westward projection of the pre-lava fault on Rockall Bank referred to above. The basement underlying both grabens has an irregular relief of *ca.* 0.5 s that may be the expression of a block faulted topography (figures 14, 20 and 22).

At 57° N, there is a major change in the trend, structure and depth of both the Hatton–Rockall Basin and the East Hatton–Rockall Basin. In the latter case, the basement shoals

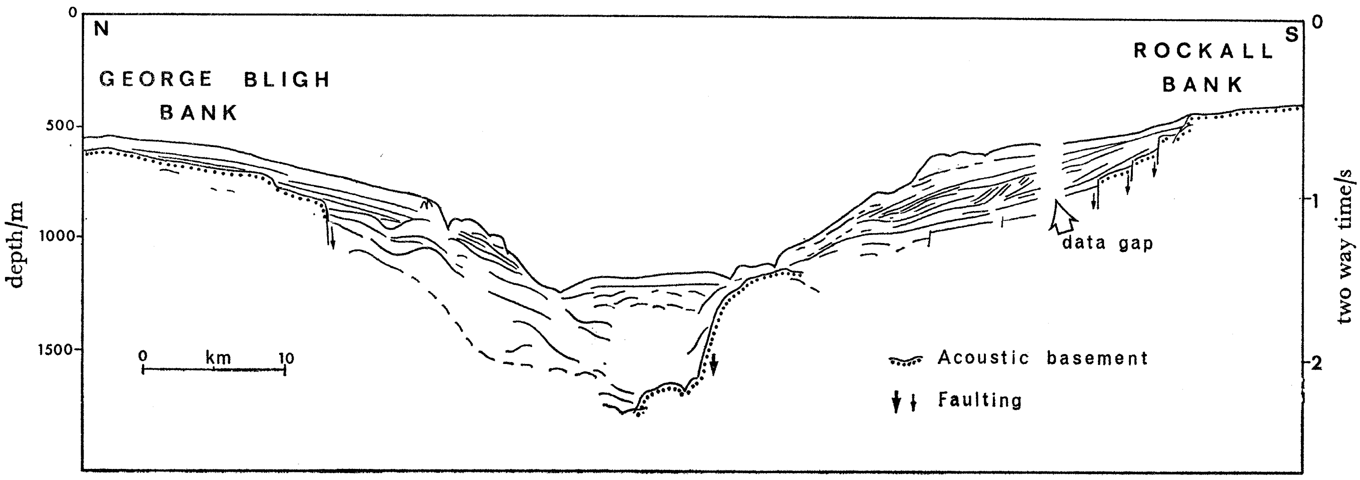


FIGURE 24. Seismic reflexion profile (24) between George Bligh Bank and Rockall Bank (9 kJ sparker). Profile is located in figure 12.

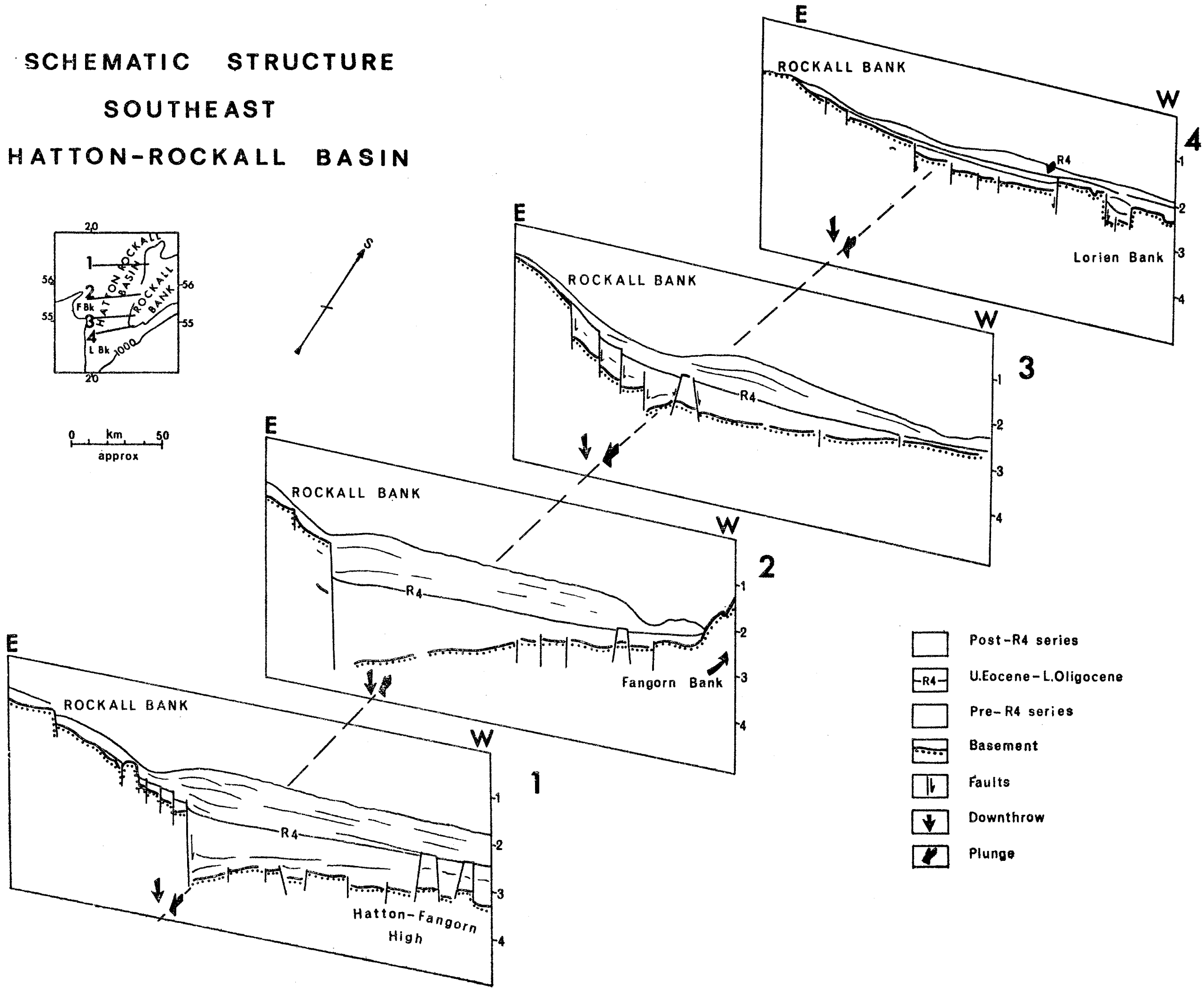


FIGURE 25. Schematic structure of the southeast Hatton–Rockall Basin.

southwestward to form a broad basement low associated with only minor faulting on the southwestern part of Rockall Bank (cf. figures 20 and 23). In the main basin, the eastern boundary faults change trend to N–S and progressively decrease in throw from 1.75 s at 57° N to about 0.3 s in the irregular basement low between Rockall Bank and Lorien Bank (figures 23 and 25). A minor change in trend of the isopachs from N–S to NNE–SSW at 56° N may indicate cross faulting by the fault system controlling the East Hatton–Rockall Basin (figures 18 and 21). The western boundary fault plexus merges southward with easterly dipping basement that is faulted antithetically to the eastern boundary fault (figure 23). The western culmination of this basement surface is a broad sub-surface high, here called the Hatton–Fangorn High (figure 23) that is cut by numerous faults and extends from Hatton Bank to Fangorn Bank and perhaps to Lorien Bank (figures 21 and 26). Both the pre- and post-R4 series thin over the high indicating subsequent syn-depositional uplift matched by subsidence along the boundary faults of the Hatton–Rockall Basin. The northward plunge of the isopachs indicate cumulative subsidence was greatest near 57° N.

Comparatively little is known of the basement structure of the southwest margin of the Rockall Plateau. A steep basement scarp extending along the slope south of Edoras Bank (Jones *et al.* 1970; Ruddiman 1972) may represent the fracture zone previously inferred from magnetic data. Seismic profiles across the Hatton–Fangorn High show the underlying basement is progressively down thrown by 3.0 s toward the adjoining oceanic basement (figure 26). At 54° N, anomaly 32, the oldest anomaly abutting the margin is offset 60 km westward (figures 3*b* and 21). To the south of the offset, the shallower basement is characterized by irregular non-linear magnetic anomalies (Vogt & Avery 1974). The offset may indicate a smaller fracture zone separating oceanic crust to the north from deeply subsided continental crust that may continue southward to 52° 30′ N (figure 21).

Seismic reflexion profiles across the west margin of Rockall Plateau show a strongly reflective flat-lying basement. Other than a locally developed basement scarp, there is no major change in character over the transition to flat-lying oceanic basement defined by the presence of anomaly 24 (Scrutton & Roberts 1971; Ruddiman 1972). Seismic refraction velocities observed on the continental Hatton Bank and on the adjacent oceanic crust have comparable values of 4.3 and 4.5 km s^{-1}. These results may be reconciled if the west margin comprises a flexure, comparable with East Greenland, covered in basic lavas and thus contiguous with the oceanic basement. The complex isopachs on the slope west of Hatton Bank do not reflect true basement structure but sediment drifts. A linear NE–SW trending horst is present at the base of the slope and a NE–SW trending scarp may control the west side of Edoras Bank (Scrutton & Roberts 1971; Ruddiman 1972).

Beneath the Hatton–Rockall Basin margins, the transition from the flat-lying to faulted basement occurs over a narrow depth range of 1100–1400 m that is also comparable to the depth of the flat-topped basement highs within the basin (figures 19 and 20). The flat-lying basement is overlain by Upper Palaeocene neritic sediments that indicate at least 1850 m of subsidence (Laughton, Berggren *et al.* 1972). These data and the restricted depth range of the transition suggest the flat-lying basement is a deeply down-warped shelf. The hole 117 data suggest the down-warping was Palaeocene–Eocene. Conceivably, rifting associated with the development of the Reykjanes Ridge may have structured the graben with associated volcanism. Subsequently, the uplifted graben margins were eroded and down-warped to form the buried shelf.

There is clearly a close relation between changes in basement trend, structure and depth to concomitant changes in trend of the west margins of Rockall Plateau. In the northern basin, the ENE–WSW trends may reflect the influence of the fracture zone north of Hatton Bank. The development of the NE–SW trending graben closely follows the change in trend of the west margin at 59° 00′ N. The N–S trends cut across the Grenville front but parallel the 76 Ma rifted margin between Fangorn and Lorien Bank. These faults may reflect the older rift fabric of the southwest margin.

(iii) *The pre-R4 series*

A. *The Hatton–Rockall Basin.* The stratigraphy of the pre-R4 series is obscured by the absence of reflectors that can be correlated basinwide and the further complication of basin margin faulting. As mentioned earlier, the JOIDES data indicate the basal part of the series is probably base Tertiary to Upper Eocene in age.

Within the central part of the Hatton–Rockall Basin, impersistent flat-lying reflectors infill the basement relief suggesting deposition by basin fill (figures 16, 20 and 22). The upper part of the series is largely conformable with R4 (figures 16, 19, 20 and 21). Beneath the basin margins, the series thins and is pinched out against the basement by the overlapping and unconformable post-R4 series so that pre-R4 sediments are absent beneath the Banks. Although stratigraphic relations are commonly obscure at the basin margins, the pre-R4 series is prograded to the east of Hatton Bank (figure 19) and overlain unconformably by the post-R4 series. West of George Bligh Bank, overlap may be present (figure 19), although the series has been much reduced in thickness by faulting.

At the basin margins, rapid variations in thickness are related to the two types of basement faulting (figures 19, 20, 22 and 23). The greatest changes in thickness are associated with the steep marginal plexus, e.g. by 1.6 s to the west of George Bligh Bank and the smallest over the flexured, locally faulted basement (cf. figures 19 and 14). Thinning, pinch out of the pre-R4 series and faulting over the Hatton–Fangorn High indicate syn-depositional uplift and subsidence (figure 23). West of George Bligh Bank, a close relation between syn-depositional faulting and basement faulting is shown by an easterly dipping and antithetically faulted reflector that is draped over the easterly dipping basement and unconformably overlain by the upper, westerly dipping part of the post-R4 series (figure 19).

The age of faulting is not entirely clear. The seismic data suggest deposition in a subsiding basin with syn-depositional faulting in early pre-R4 time. In support, palaeoenvironment data from hole 117 indicate progressive deepening from a neritic Upper Palaeocene environment to about 600 m in the Early Eocene (51 Ma). Minor faulting of R4 compared to the pre-R4 series (for example figure 19) suggests this phase of subsidence was largely completed by the Upper Eocene. The Middle Eocene of hole 117 is tectonically disturbed indicating faulting ended in post-Middle Eocene–pre-Upper Eocene time. The early basin fill and faulting may be related to the downwarping of the graben margins discussed in the previous section.

B. *The west and southwest margins of Rockall Plateau and the Reykjanes Ridge.* On the west margin, the pre-R4 series infills irregularities in the basement and may be unconformably overlain by R4 (Scrutton & Roberts 1971; Ruddiman 1972, Fig. 17). It is thinner (*ca.* 0.5 s) than in the Hatton–Rockall Basin, but thickens westward to about 1.0 s adjacent to basement scarps developed at the 42 Ma (anomaly 18) spreading discontinuity. West of these scarps, it is pinched out on 37 Ma (anomaly 13) oceanic basement by R4. The basal part of the series

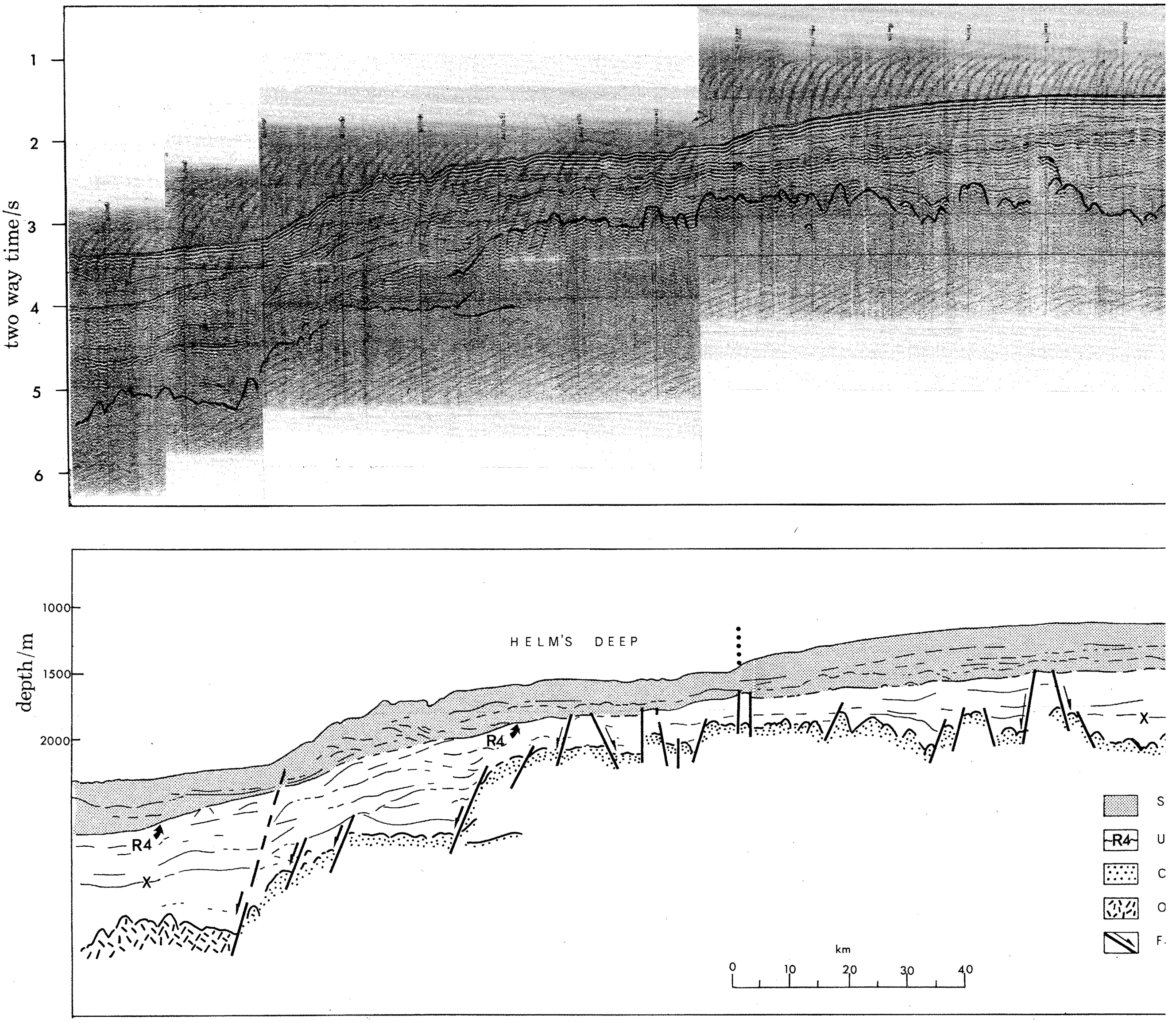
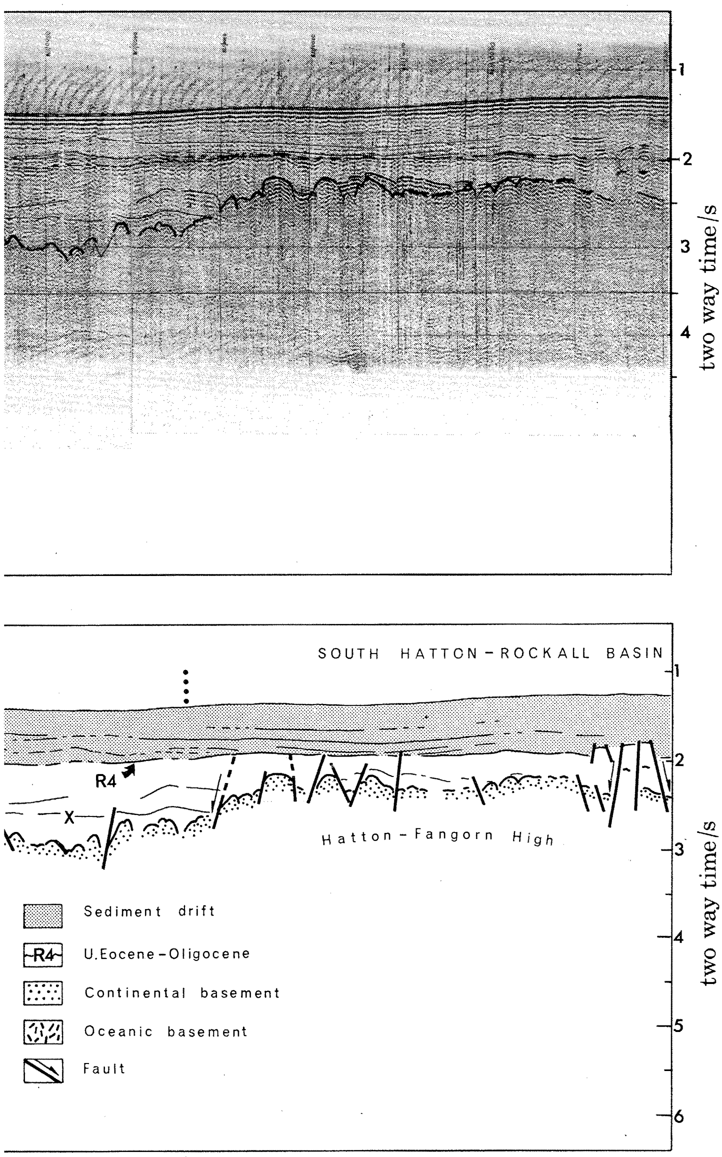


FIGURE 26. Seismic reflexion profile (26) across the southwest margin of the south Hatton-Rockall



-Rockall Basin. Profile is located in figure 12.

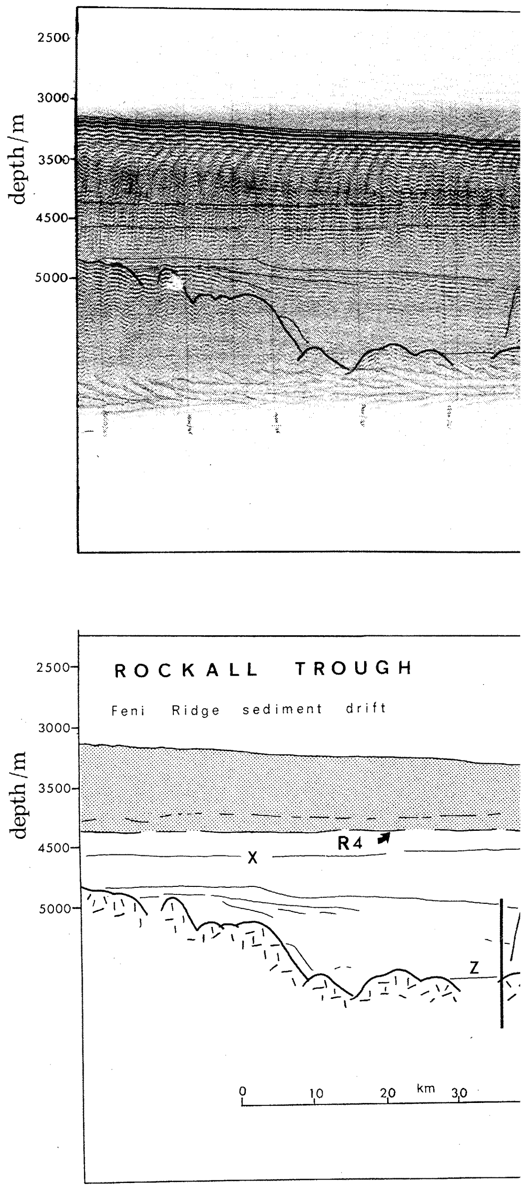
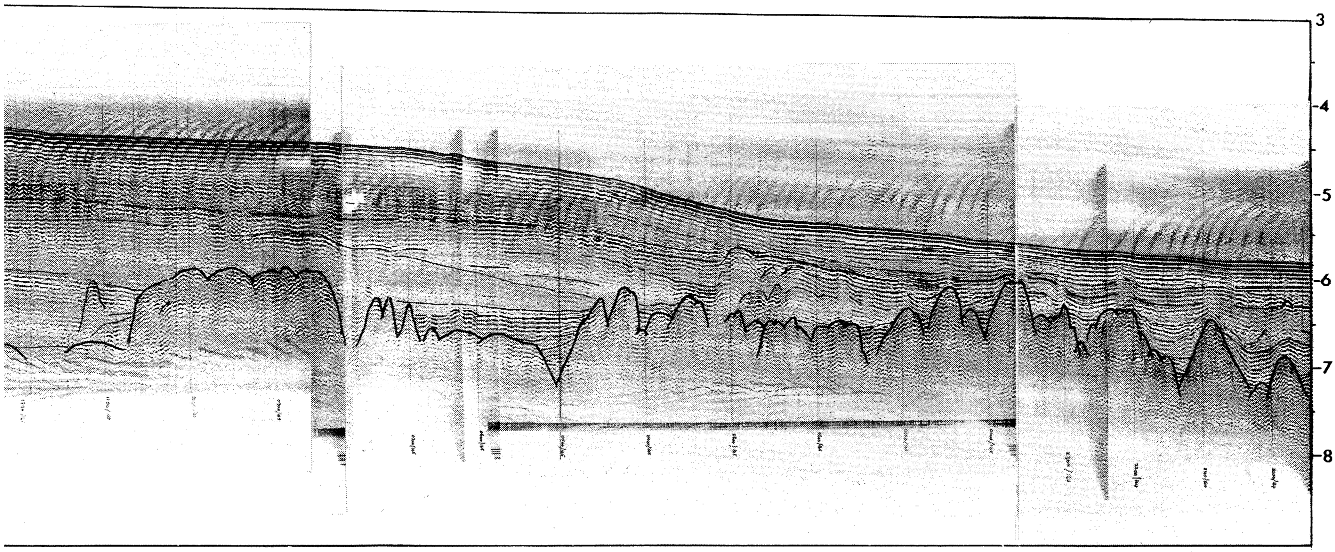
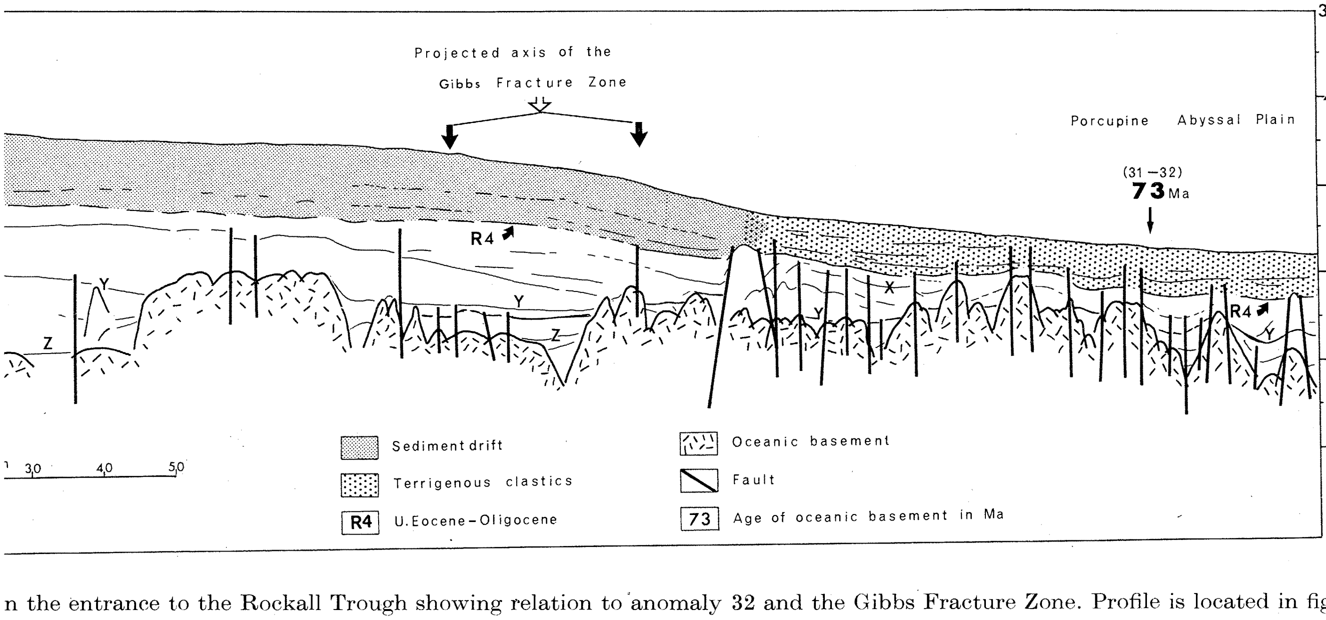
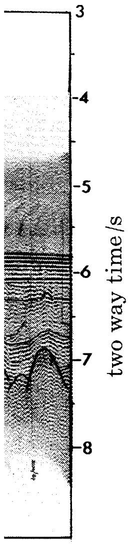


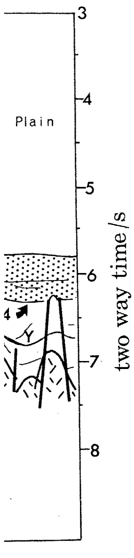
FIGURE 34. Longitudinal seismic profile in the en



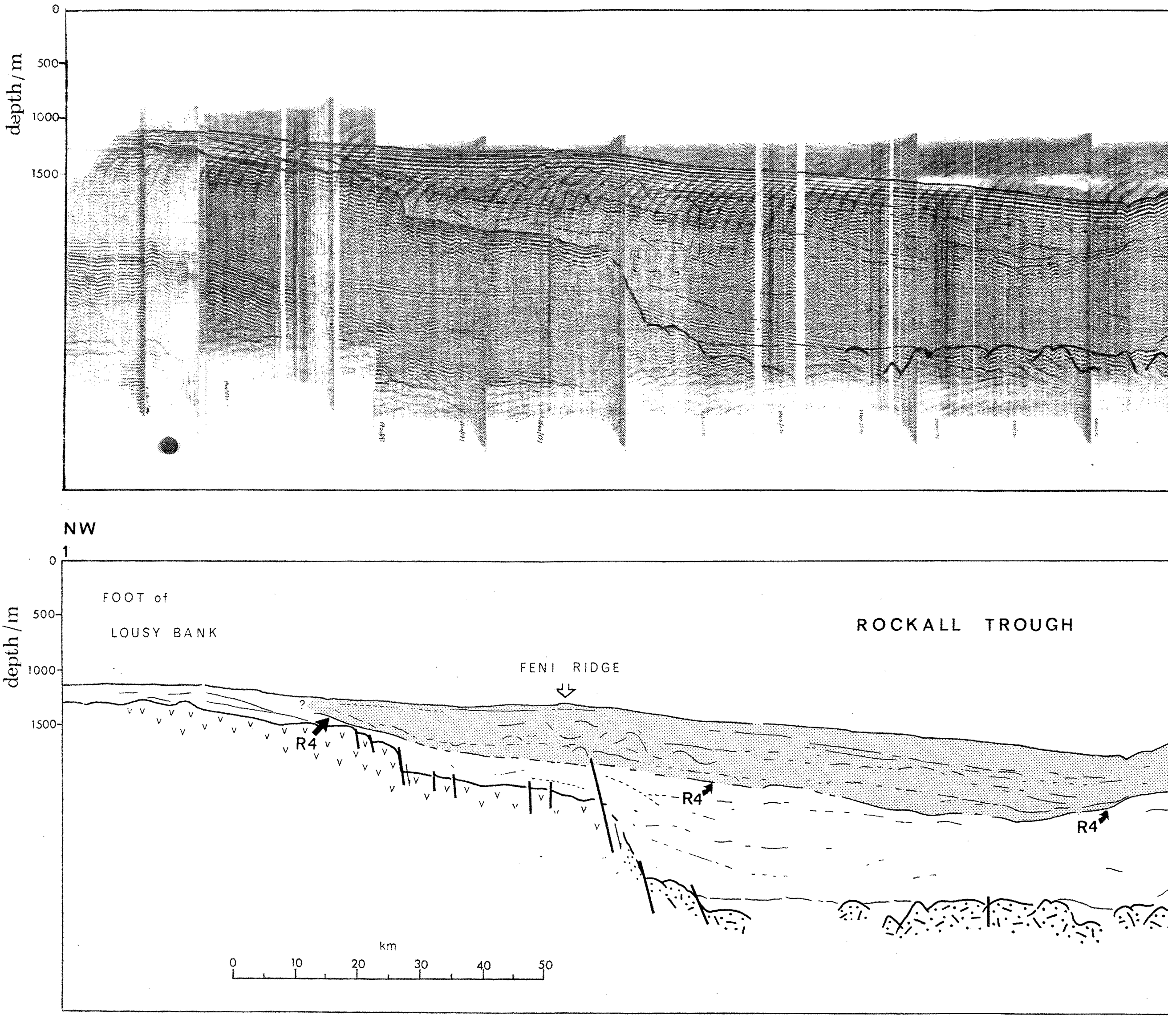


n the entrance to the Rockall Trough showing relation to anomaly 32 and the Gibbs Fracture Zone. Profile is located in figu





ed in figure 31.



0
500
1000
1500
depth/m
NW
1
FOOT of
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km
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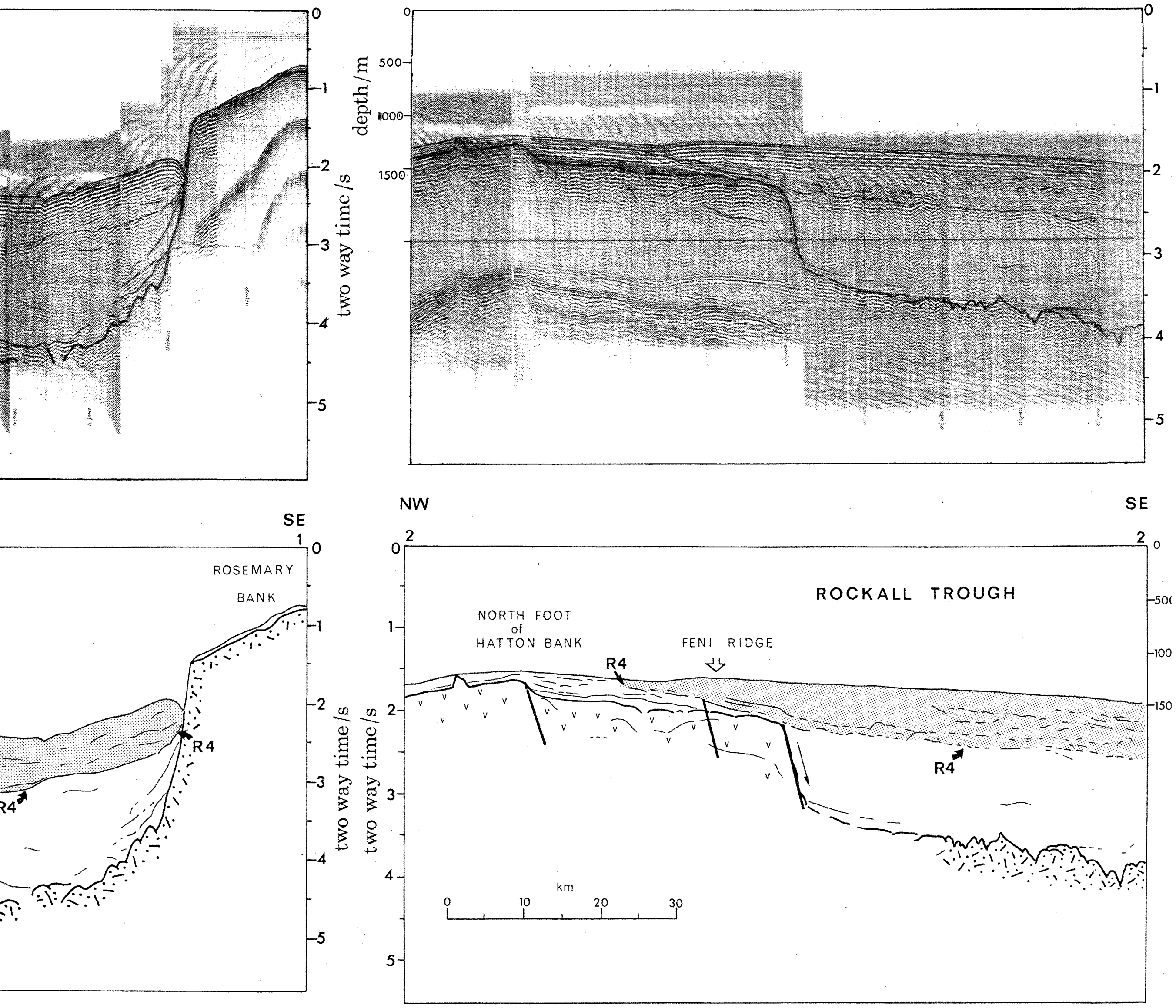
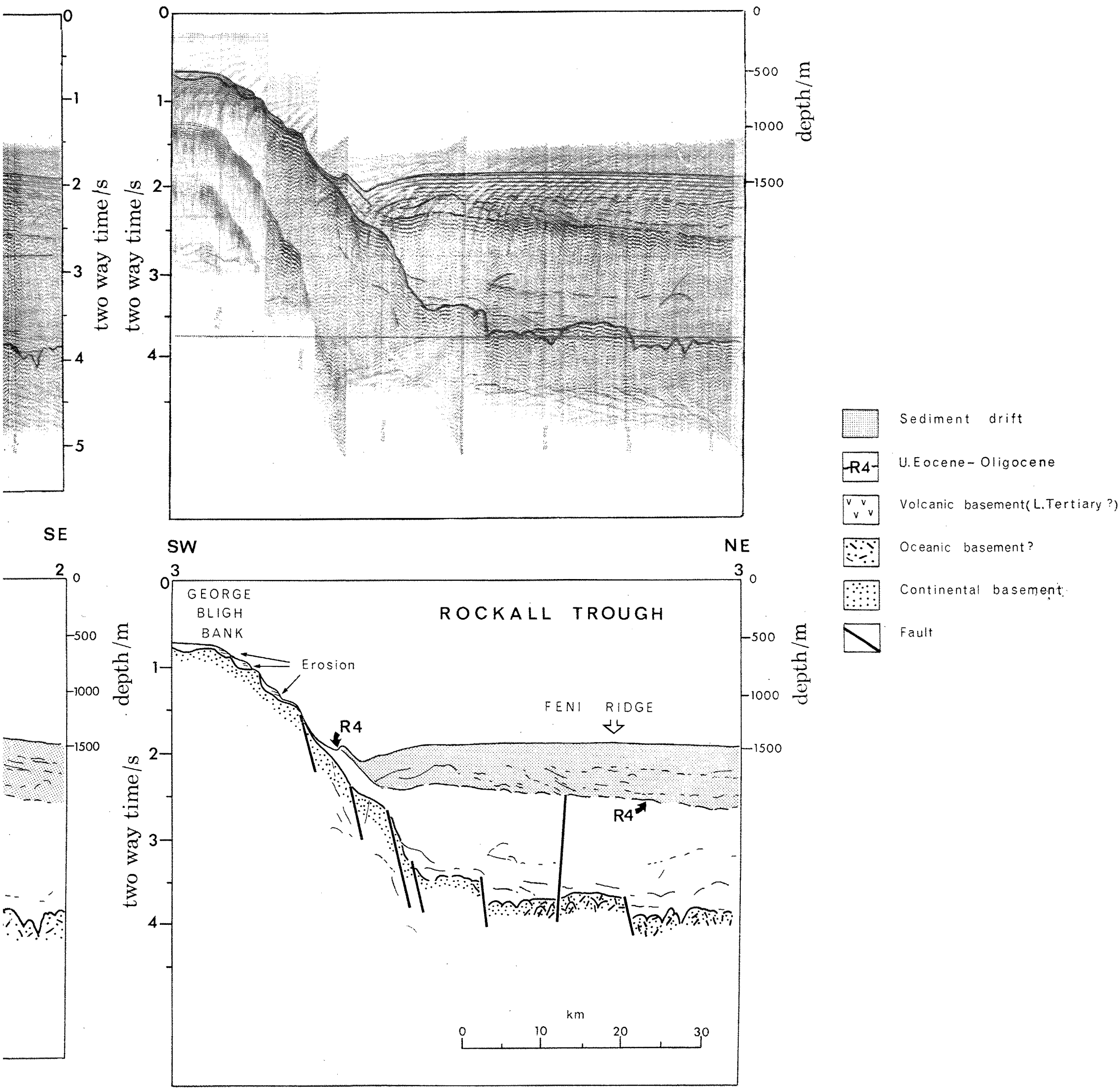


FIGURE 35. Seismic reflexion profiles (35) in the northern Rockall Trough. Profiles located in figure 12.


two way time/s
depth/m
SE
SW
NE
GEORGE
BLIGH
BANK
ROCKALL TROUGH
Erosion
R4
FENI RIDGE
km
0 10 20 30
Sediment drift
U. Eocene – Oligocene
Volcanic basement (L. Tertiary ?)
Oceanic basement ?
Continental basement
Fault

is time transgressive since it rests on flat-flying oceanic basement ranging in age from 37 to 60 Ma (anomaly 24). Note that the westward increase in thickness coupled with decreasing age is the converse of the age–thickness relation normally observed in the ocean basins and must indicate a westward increase in sedimentation rate (Ewing & Ewing 1967; Ruddiman 1972).

The stratigraphy of the pre-R4 series developed on the southwest margin is poorly known. The series adjacent to the margin is about 1.7 s thick (cf. Hatton–Rockall Basin) and rests on 76 Ma oceanic basement. Within the series, there is a strong persistent reflector called 'X' whose age has been previously estimated at *ca.* 60 Ma. The reflector is draped over the oceanic basement and is unconformable with R4 (figure 26). The seismic data suggest reflectors above X are flat-lying perhaps indicating increased sedimentation associated with erosion of the then subaerial Rockall Plateau (cf. hole 117).

(iv) *Reflector R and the post-R4 series on and west of the Rockall Plateau*

Within the Hatton–Rockall Basin, R4 is a discrete sub-horizontal reflector lying at about 0.75 and 1.0 s depth (figure 27). It is typically associated with overlapping hyperbolae that originate within the cherts and commonly subdue or mask R4. The pre- and post-R4 series are conformable within the basin though an important disconformity is widely developed at the top of the cherts .There is also a marked acoustic contrast between the transparency of the pre-R4 series, the strongly reflective chert sequence and the impersistent cuspate reflexions that characterize the post-chert sequence (figure 16). In general R4 is remarkably undeformed although diffractions originating from R4 may indicate minor faulting or differential compaction (figure 16). Towards the basin margins R4 is unconformably overlapped by the post-R4 series and is affected by faulting. As mentioned earlier, the unconformity decreases basinward (figures 16, 19, 20, 22 and 23). The seismic profiles and JOIDES data show the cherts thin and pinch out against R4, and both are progressively overstepped by the post-chert sequence as it thins toward the Banks. On the Hatton–Fangorn High, R4 is broadly domed and the lower parts of the post-R4 series are cut out against it (figure 23).

On the southwest margin, R4 can be followed from the Hatton–Fangorn High into the adjoining ocean basin. A sequence of strong reflectors about 0.2 s in thickness and closely comparable to the cherts overlies R4 but pinches out against it on the west side of the Hatton–Fangorn High (figure 26).

A common feature of all three areas (and also of the Rockall Trough) is the widespread presence of cherts followed by a sequence that is primarily the product of differential deposition. As in many other well documented and comparable instances, there is a marked unconformity between the cherts, the post-chert sediments and the seabed. Marginal channels and sediment ridges are present in both the seabed and sub-surface (for example figures 22 and 25). Although the post-chert sequence can be largely attributed to this mechanism, there are also important differences in the time and space development of differential deposition both within and between the Hatton–Rockall Basin and Reykjanes Ridge.

In the northern Hatton–Rockall Basin, the post-chert sequence is thinner and individual reflectors are more persistent exhibiting onlap toward the basin margins (figure 19). Differential deposition is indicated by the ridges and channels cutting George Bligh Bank (figure 19), although the basin floor is not significantly arched compared to the south. On sections south of 58° N, the more important role of differential deposition is shown by the gentle arching of the basin floor that is present throughout the whole post-chert sequence and may also affect the cherts (figure 22). On the west slope of Rockall Bank, irregular ridges and channels are indicative of more localized erosion and/or non-deposition compared to the Basin (figure 14). The arching is best developed around 57° N although the maximum thickness does not underlie the crest and is developed towards the east side of the Basin (cf. figures 27 and 22). To the south of 56° 40′ N, the arch does not continue southward and the greatest thickness is no longer axial (cf. figure 22 with figures 23 and 25). The change reflects the sub-surface influence

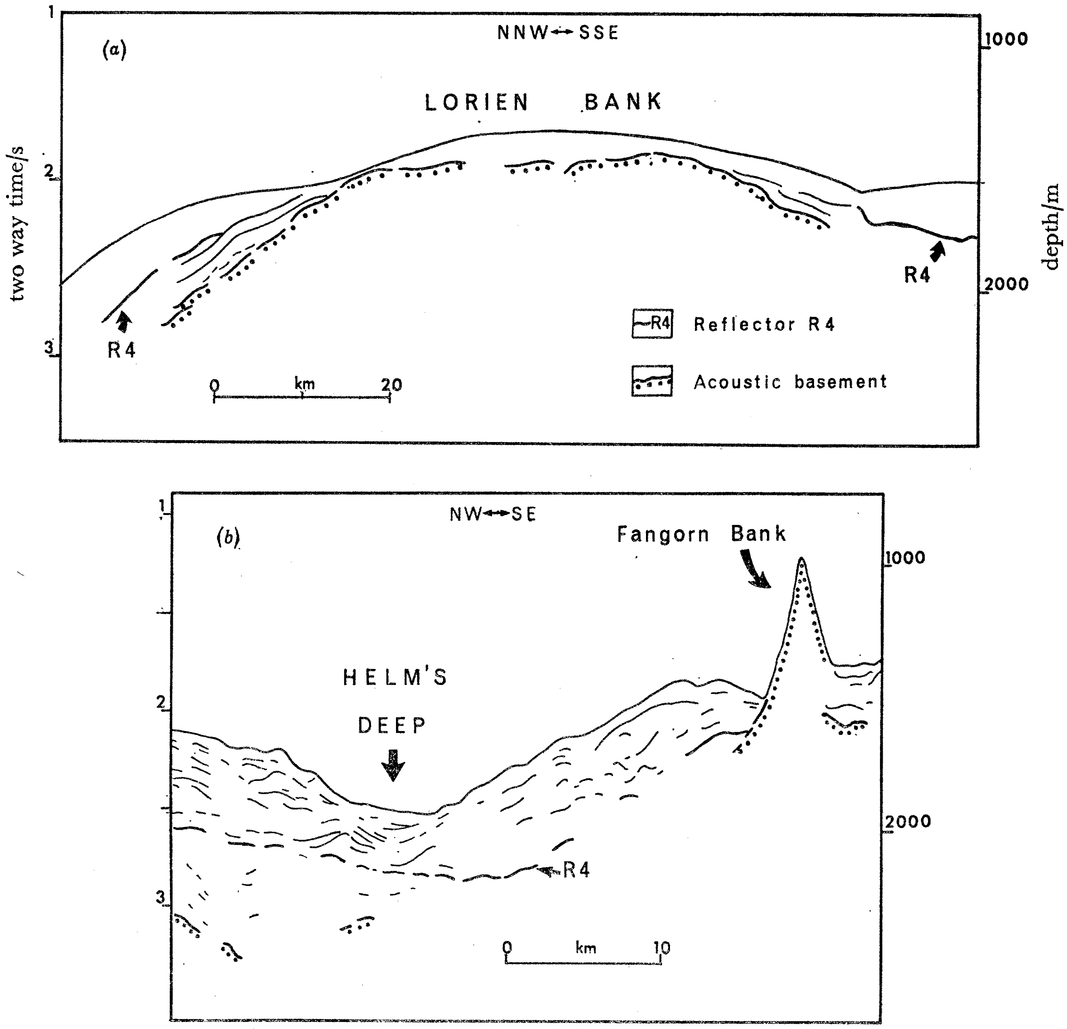


FIGURE 28 (*a*). Seismic reflexion profile across Helm's Deep. (*b*) Seismic reflexion profile across Lorien Bank. Profiles located in figure 12.

of the Hatton–Fangorn High and also the changing trend and slope of the east margin of the Basin. Over the Hatton–Fangorn High, the attitude of the reflectors and thinning of the post-R4 sequence indicate tilting and uplift of the High during deposition of the lower part of the series. Extreme differential deposition around Fangorn Bank has locally enhanced the thinning. The change in trend of the east margin to NNE–SSW coincides with the disappearance of the arch and the appearance of a linear sediment drift extending the length of the east margin (figures 26, 28 and 40). Both drift and the arch are clearly closely related since the flanking marginal channels form part of a continuous system (figure 9). However, the sediments underlying both ridge and channel show a southward decrease in thickness that may perhaps be related to the concomitant southward decrease in relief and dip of both R4 and basement (figure 26).

On the west margin of Rockall Plateau, the Hatton sediment drift extends from 58° 30′ N to 55° 30′ N over a depth range of 1000–1500 fathoms (1818–2743 m) and has caused the smooth appearance of the slope. The drift is *ca.* 1.0 s thick in contrast to minimal thicknesses on either side (Scrutton & Roberts 1971; Ruddiman 1972). At 56° 30′ N, the drift is cut by the sub-parallel systems of drifts and non-depositional channels that partly encircle Edoras Bank and comprise the western edge of the southeastern part of the Hatton–Rockall Basin (figure 25). This relation suggests the sedimentary process responsible for their formation predominates over the long-slope deposition responsible for the Hatton Drift.

On the southwest margin, the widely spaced seismic profiles show the post-R4 stratigraphy has been largely controlled by differential deposition evidenced by moating around Rohan seamount (Jones *et al.* 1970), the presence of sediment drifts and impersistent cuspate reflectors that mirror undulations in the seabed. The seismic profiles show a continuum of differential deposition between the Hatton–Rockall Basin and the ocean basin floor. For example Helm's Deep (figure 28*a*) underlain by cross-bedded sediments showing non-deposition and local erosion, is a non-depositional channel that begins in the non-depositional channels east of Fangorn Bank and extends over a depth range of 1700 m.

Sediment distribution and deposition of the post-R4 series on the east flank of the Reykjanes Ridge have been largely governed by the effects of topography on the southwestward flow of North East Atlantic Deep Water (Johnson & Schneider 1969; Johnson *et al.* 1971; Davies & Laughton 1972; Ruddiman 1972). The Gardar sediment drift is *ca.* 1.80 s thick and initially developed adjacent to a 45 Ma basement escarpment (Ruddiman 1972). Gardar drift sediments extend almost to the base of Rockall Plateau but beneath Maury Channel pass laterally into volcanogenic turbidites extending to a depth of 0.3 s. The greater, 1.80 s thickness of the Gardar Drift compared to the Hatton–Rockall Basin suggests a greater sediment supply to the former in post-Oligocene time.

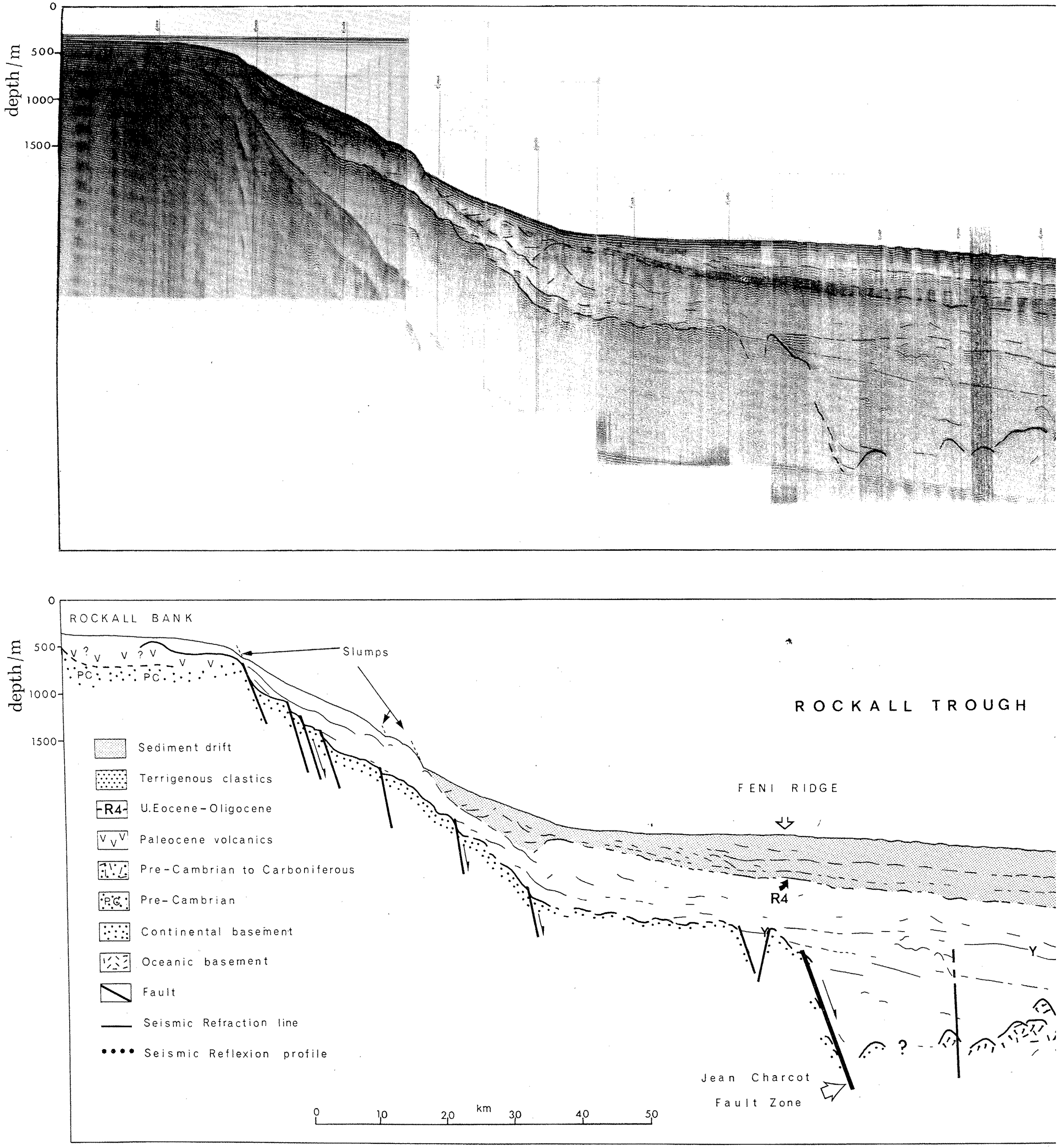
Evidence of a change in differential deposition on the Reykjanes Ridge at 10–18 Ma has been suggested by Ruddiman (1972). A possibly contemporaneous change may also be present in the Hatton–Rockall Basin. For example, east of the Hatton Bank, a sub-surface marginal channel broadens and decreases in relief at 0.5 s depth with migration of the arch (figure 20). At 57° N, a sub-surface channel underlies the dome crest and continues to within 0.5 s of the seabed where it is overlain unconformably by the rest of the series (figure 22). The present day marginal channel can be traced to 0.5 s depth where it rests on the earliest, marginal and onlapping part of the series. These changes may be partly related to the decreasing dip of R4. For example, on figure 20, the channel increases in width at the decrease in dip of R4 in contrast to the migration up dip of the channel against the uniformily dipping R4. Apart from these topographic relations, the migration of depositional and non-depositional axes suggests a change in bottom current regime that may correspond to the change in sedimentation rate from 3 to 6 cm per 1000 years at 8 Ma recorded in JOIDES hole 116. The similar (but) less precisely defined change (10–18 Ma) on the Reykjanes Ridge thus suggests a regional change in North Atlantic bottom water circulation. Evidence of a Pleistocene change in sedimentation rate is also indirectly shown by the presence beneath Gardar drift sediments of the Pleistocene volcanogenic turbidites of the Maury fan (Ruddiman 1972). The uppermost Pleistocene fan sediments are between 0.1 and 0.2 s depth (*ca.* 90–180 m at a V_p of 1.8 km s^{-1}) and indicate a post-Pleistocene sedimentation rate of 9–18 cm per 1000 years compared to a mean rate of *ca.* 5 cm per 1000 years for the whole drift at its thickest point.

Deformation of R4 and the post-R4 series is slight and evidenced by a gentle regional warping accompanied by minor faulting (figure 22). On Rockall Plateau, a useful assessment of the often variable influence of basement tectonics is given by a comparison of isopachs on R4 and basement though care is necessary to distinguish subsidence from thickening by differential deposition. In the Hatton—Rockall Basin, the conformity between these isopachs suggests the downwarping has been largely conditioned by the basement structure. In the northern part of the basin, where differential deposition has been least, the R4 isopachs outline a closed basin elongated ENE–WSW parallel to Hatton Bank in contrast to the basement isopachs which show the greatest pre-R4 subsidence to the west of George Bligh Bank. Towards Hatton Bank, faults cut R4 and the cherts but not the overlying sequence (figure 19). Note that the major fault west of George Bligh Bank does not displace R4. The distribution of faults and the isopachs suggest a gentle northward tilt toward Hatton Bank. Further south, R4 is downfaulted by as much as 0.5 s along rejuvenated basement faults (figure 20). In the East Hatton–Rockall Basin, R4 is tilted westward against the boundary faults of the horst (figure 20). In the axial parts of the basin, however, there is no major faulting (cf. figures 16, 20, 22, 23 and 24). Between 58° and 57° N, the downwarp of R4 has been closely controlled by the basement graben.

To the south of 57° N, the basement and R4 isopachs change trend to NNE–SSW and show R4 plunging northward (figures 25 and 26). The seismic profiles show the northward plunge is accompanied by eastward tilt indicating down faulting and northward rotation against the NNE trending boundary faults (figure 25). The absence of deformation between Rockall Bank and Lorien Bank suggests the hinge may have lain in that area (figure 25). These movements were matched by relative uplift of the Hatton–Fangorn High evidenced by the arching and faulting of R4 (figures 23 and 26). As in the northern part of the basin, the faulting appears to affect only R4 and the lowermost part of the post-R4 series.

The seismic profiles suggest the faulting was completed by the end of chert deposition, i.e. end Oligocene–Early Miocene. In support, syn-depositional faults were detected in the Lower Oligocene of hole 116 and Laughton, Berggren *et al.* (1972) have suggested on palaeoenvironmental grounds that about 1200 m of subsidence occurred between the base of the Oligocene and the Middle Miocene.

To the west of the Hatton–Fangorn High and on the west margin of the Hatton–Rockall Basin, minor faulting of R4 is present but dies out toward the base of the slope. On the east flank of the Reykjanes Ridge and adjacent to the west margin of the Rockall Plateau, R4 is flat-lying and late faulting of the type observed on the equatorial Mid-Atlantic Ridge is absent (Van Andel & Heath 1970; Ruddiman 1972). Minor faulting may be present adjacent to the 45 Ma basement scarp (Scrutton & Roberts 1971). The absence of faulting is surprising in view of the gentle but significant deformation observed on the Rockall Plateau. The difference may imply decoupling between the continental and oceanic lithosphere though profiles figured by Ruddiman do not show any evidence of deformation in support. More plausibly, its absence may reflect the fundamental difference in tectonic fabric and geology of oceanic and continental lithosphere. The origin and implications of the R4 deformation are reviewed later in terms of the regional development of the Rockall Plateau and Trough.



0
500
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depth/m
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Slumps
ROCKALL TROUGH
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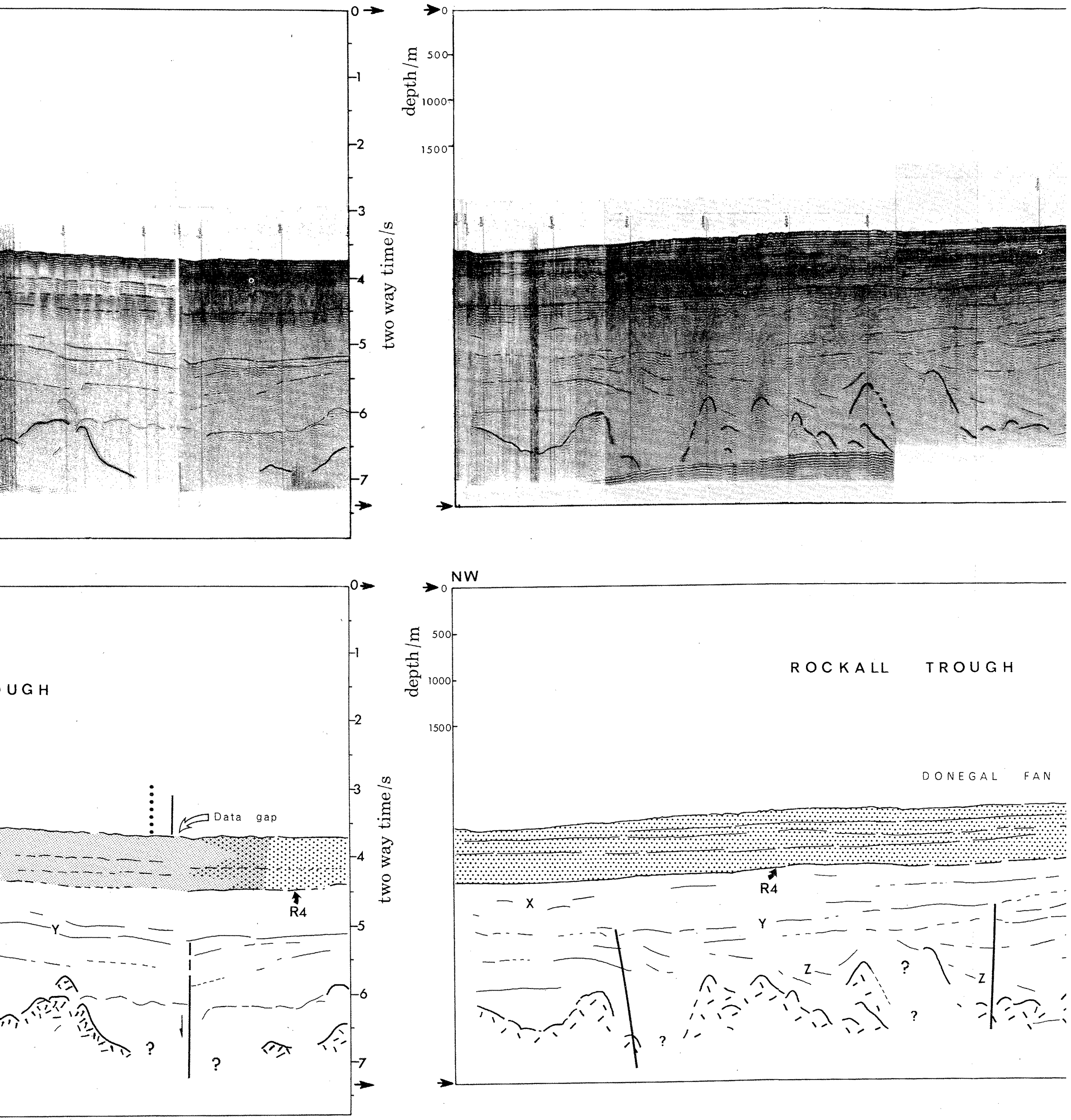
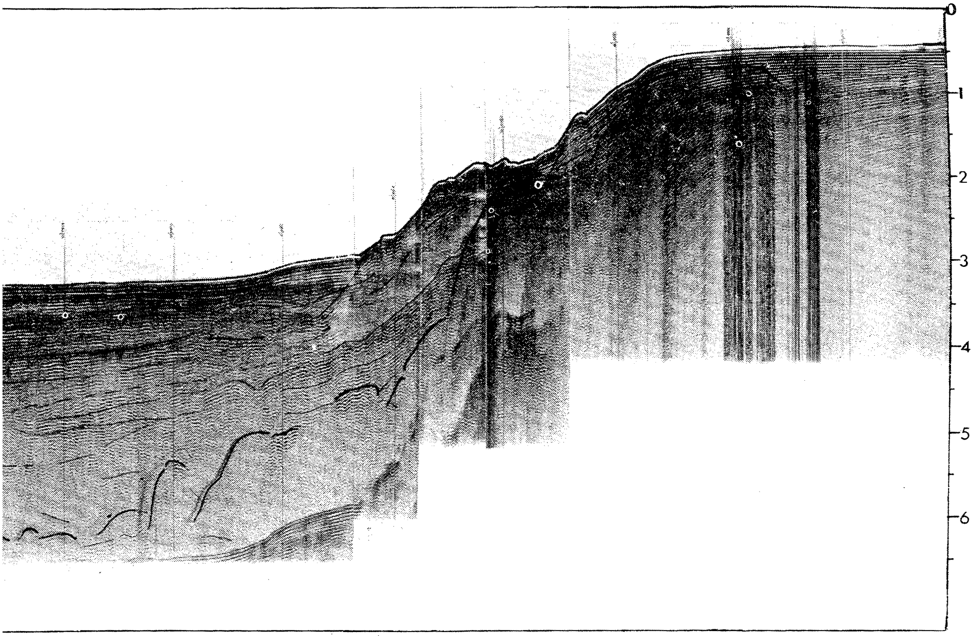
Sediment drift
Terrigenous clastics
R4
U. Eocene–Oligocene
Paleocene volcanics
Pre-Cambrian to Carboniferous
Pre-Cambrian
Continental basement
Oceanic basement
Fault
Seismic Refraction line
Seismic Reflexion profile
Jean Charcot
Fault Zone
0
10
20
km
30
40
50

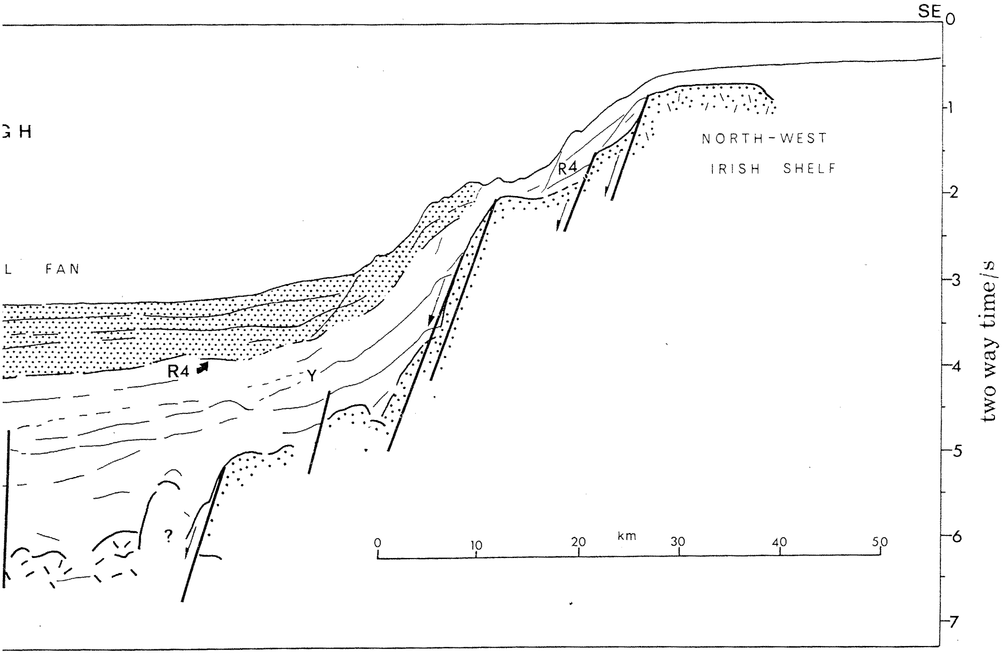
FIGURE 29. S



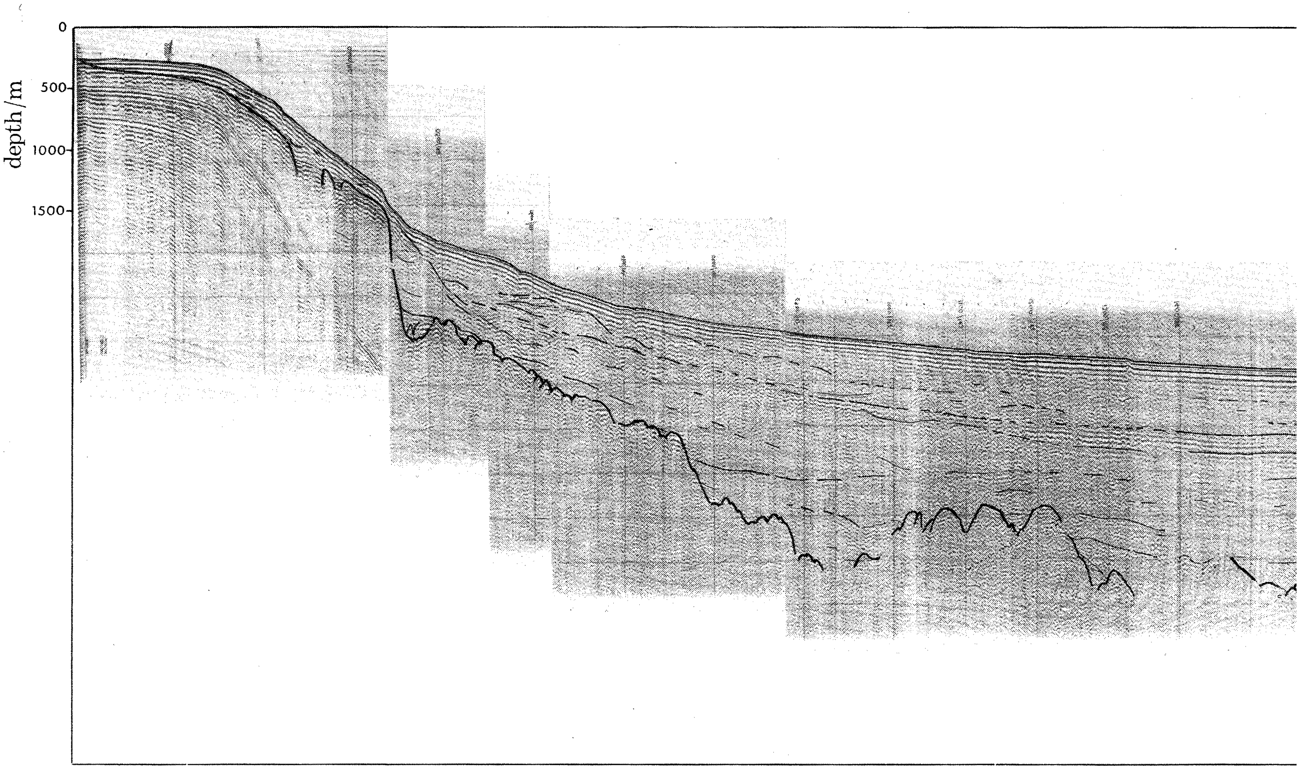
RE 29. Seismic reflexion profile (29) between Rockall Bank and Donegal. Profile is located in figure 12.

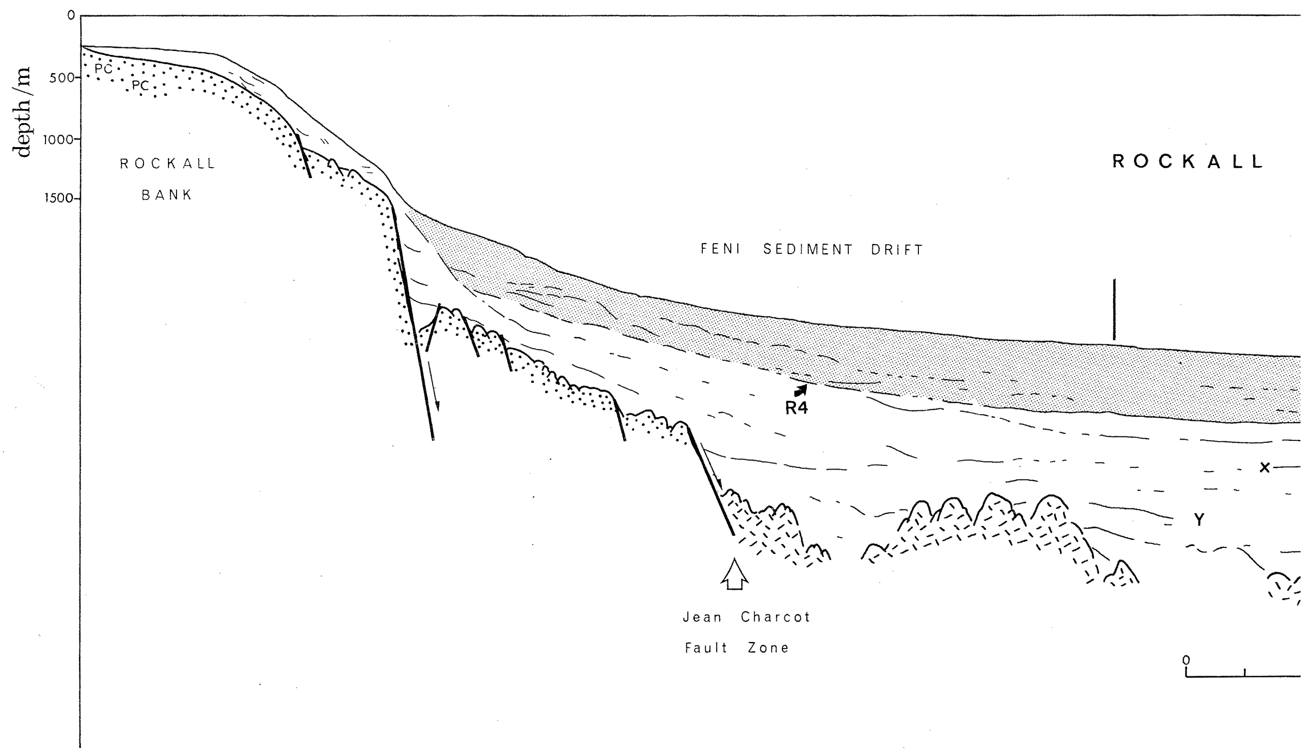


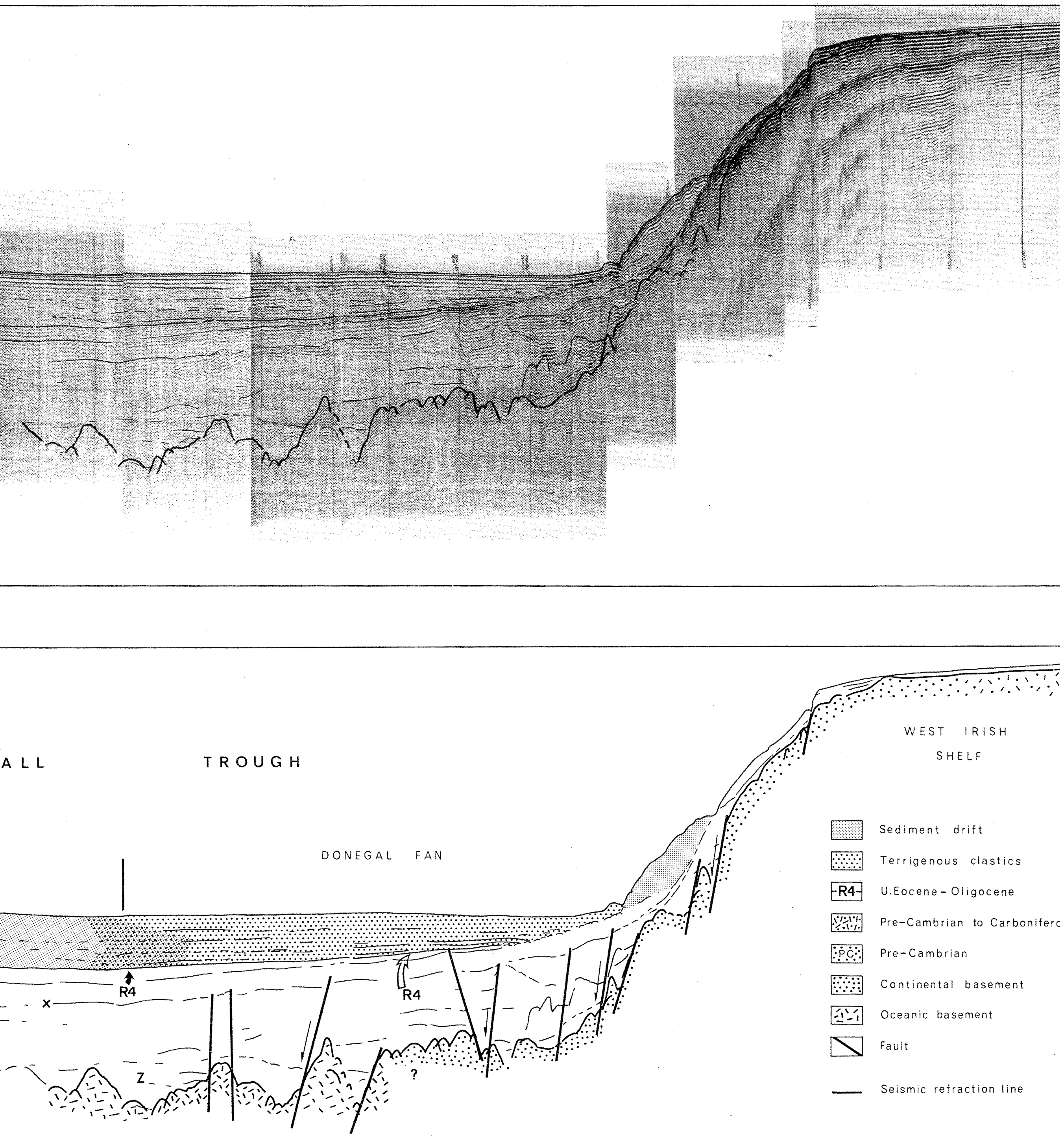
0
1
2
3
4
5
6
two way time/s

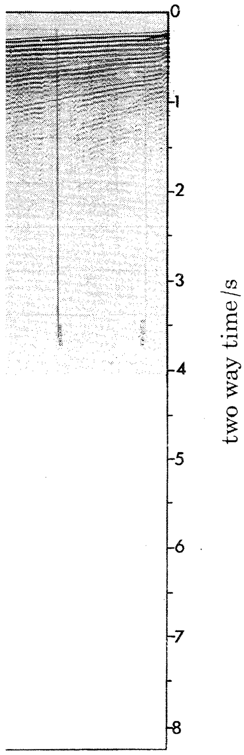


SE
0
1
2
3
4
5
6
7
two way time/s
G H
L FAN
R4
Y
R4
NORTH-WEST
IRISH SHELF
?
0
10
20
km
30
40
50


0
500
1000
1500
depth/m


0
500
1000
1500
depth/m
PC
PC
ROCKALL
BANK
ROCKALL
FENI SEDIMENT DRIFT
R4
X
Y
Jean Charcot
Fault Zone
0


ALL
TROUGH
DONEGAL FAN
WEST IRISH
SHELF
R4
R4
X
Z
?
50
100 km
Sediment drift
Terrigenous clastics
R4
U. Eocene – Oligocene
Pre-Cambrian to Carbonifero
PC
Pre-Cambrian
Continental basement
Oceanic basement
Fault
Seismic refraction line


0
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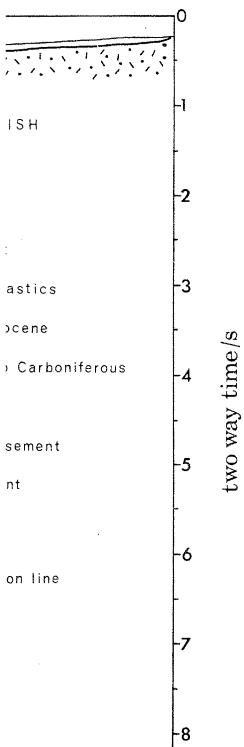
6
7
8
two way time/s


ISH
astics
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two way time/s

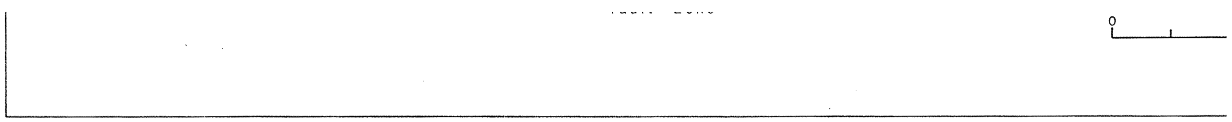
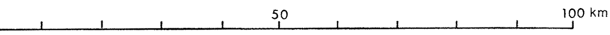
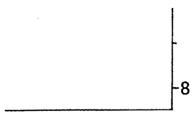


FIGURE 30. Seismic reflexion profile (30) between I



between Rockall Bank and Achill Island. Profile is located in figure 12.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(*d*) *The Rockall Trough*

(i) *Basement structure*

There is a marked asymmetry in basement depth and structure that is best developed between Anton–Dohrn Seamount and 53° 30′ N. On the west side of the Trough, the basement forms a 30 km wide platform lying at about 1.3 s depth (figure 30). The platform is bounded by a prominent NE–SW trending escarpment with a relief of *ca.* 2.0 s and dip of about 20° that is here called the Jean Charcot Fault Zone (figures 21, 29 and 30). No comparable feature is present along the east margin although there is a change in basement relief at about 2.0 s depth (figures 29 and 30). To the east of the scarp, the basement has a typical relief of *ca.* 1.0 s and wavelength of 20 km, that may on the limited evidence of intersecting profiles trend NE–SW. The seismic profiles suggest that this type of basement relief is confined within the 2.0 s isopachs (figure 30). A gravity model suggests the Jean Charcot Fault Zone (figure 3*a*) marks the boundary between the thicker continental crust of Rockall Plateau and the thinner crust beneath Rockall Trough. The nature of the thinner crust has yet to be defined precisely. However, the basement relief and refraction velocities of 4.97 km s^{-1} (Scrutton & Roberts 1971; Ewing & Ewing 1958) are comparable to those typical of oceanic basement. Further, the basement within the 2.0 s isopachs is characterized by NE trending linear magnetic anomalies (figure 31) that contrast with the incoherent anomalies associated with the platform (Vogt & Avery 1974). This evidence suggests that the thinner crust is oceanic. The Jean Charcot Fault Zone may therefore mark the continent-ocean boundary and the 2.0 s isopach may mark it beneath the eastern part of the Trough.

South of 54° 30′ N, the Jean Charcot Fault Zone radically changes trend to E–W and is nearly perpendicular to the east margin of Rockall Plateau. North of Porcupine Bank the 2.0 s isopach congruently trends westward across the whole width of the Trough outlining a 100 km wide east–west trending basement low (figure 31). The sides of the low are steepest adjacent to Rockall Plateau and are matched by congruent steepening north of Porcupine Bank (figures 18 and 31). Within the low, narrow troughs are infilled by flat-lying sediments (figure 32). Basement south of the low is shallowest in the centre of the Trough compared to the sides (cf. *ca.* 1.5 and 2.5 s). Linear magnetic anomalies to the north and south are truncated against the minimum (figure 31). Comparable features on large fracture zones suggests the minimum may be a buried fracture zone that is here called the Porcupine Fracture Zone (figure 21). The congruent trend changes in the Jean Charcot Fault Zone and north margin of Porcupine Bank may mark offsets in the margin associated with the fracture zone. Note that an E–W direction of opening is thus implied for the Rockall Trough.

South of the Porcupine Fracture Zone, the position of the continent ocean boundary is not known in detail. Le Pichon *et al.* (1970) observed a prominent basement scarp near 53° N 16° W which they called the Jean Charcot Fracture Zone postulating a northerly continuation with the scarp described above. However, the isopachs and detailed surveys (figure 18) show it does not continue north to the Porcupine Fracture Zone and suggest it is a buried seamount. West of Porcupine Bank, at 2.0 s depth, a change to an irregular hummocky relief typical of oceanic basement may indicate the approximate continent/ocean boundary (figure 33). On the west side of the Rockall Trough (west of 15° W), the Jean Charcot Fault Zone has not been proven. However, a scarp, defined by the 1.5 s isopach, has comparable relief and is abutted by basement with refraction velocities of 4.9 km s^{-1} (Hill 1952). Further, this scarp (figure 31)

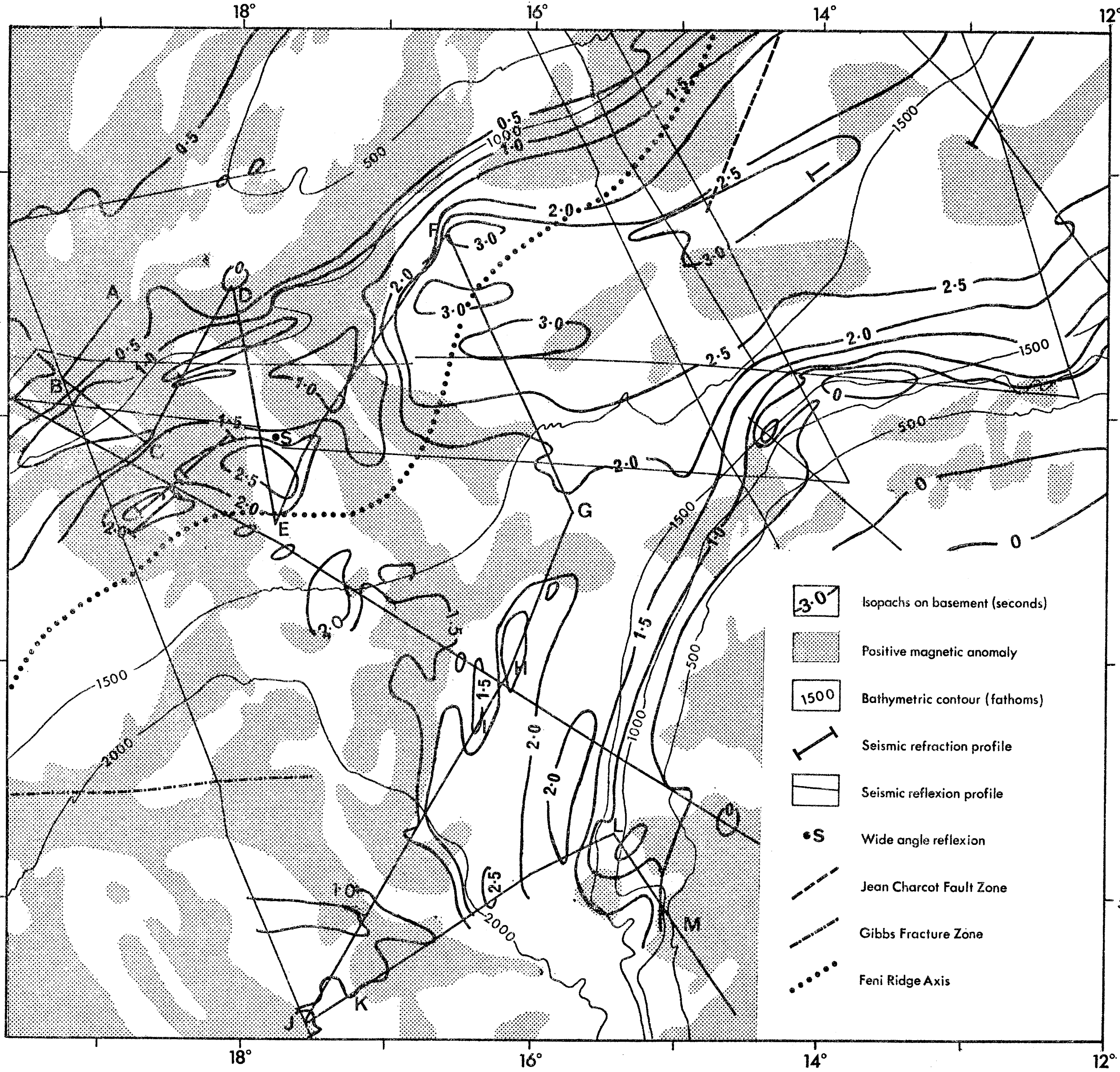

FIGURE 31. Isopachs on basement and magnetic anomalies across the Porcupine Fracture Zone (after Vogt & Avery 1974). Lettered tracks locate seismic profiles of figures 32, 33 and 34.

marks the boundary between the irregular 'continental' magnetic anomalies of Rockall Plateau and the broader longer wavelength anomalies of Rockall Trough (Vogt & Avery 1974). It may therefore mark the continent–ocean boundary. An eastward bulge in the 1.5 s isopach and associated east–west grabens may indicate another fracture zone to the south of the Porcupine. The trend of the 1.5 s isopach and new magnetic data (Vogt & Avery 1974) suggest that continental basement, contiguous with that west of Lorien Bank, may continue southward towards the Gibbs Fracture Zone.

The structural and age relation of the Gibbs Fracture Zone to the Rockall Trough has been

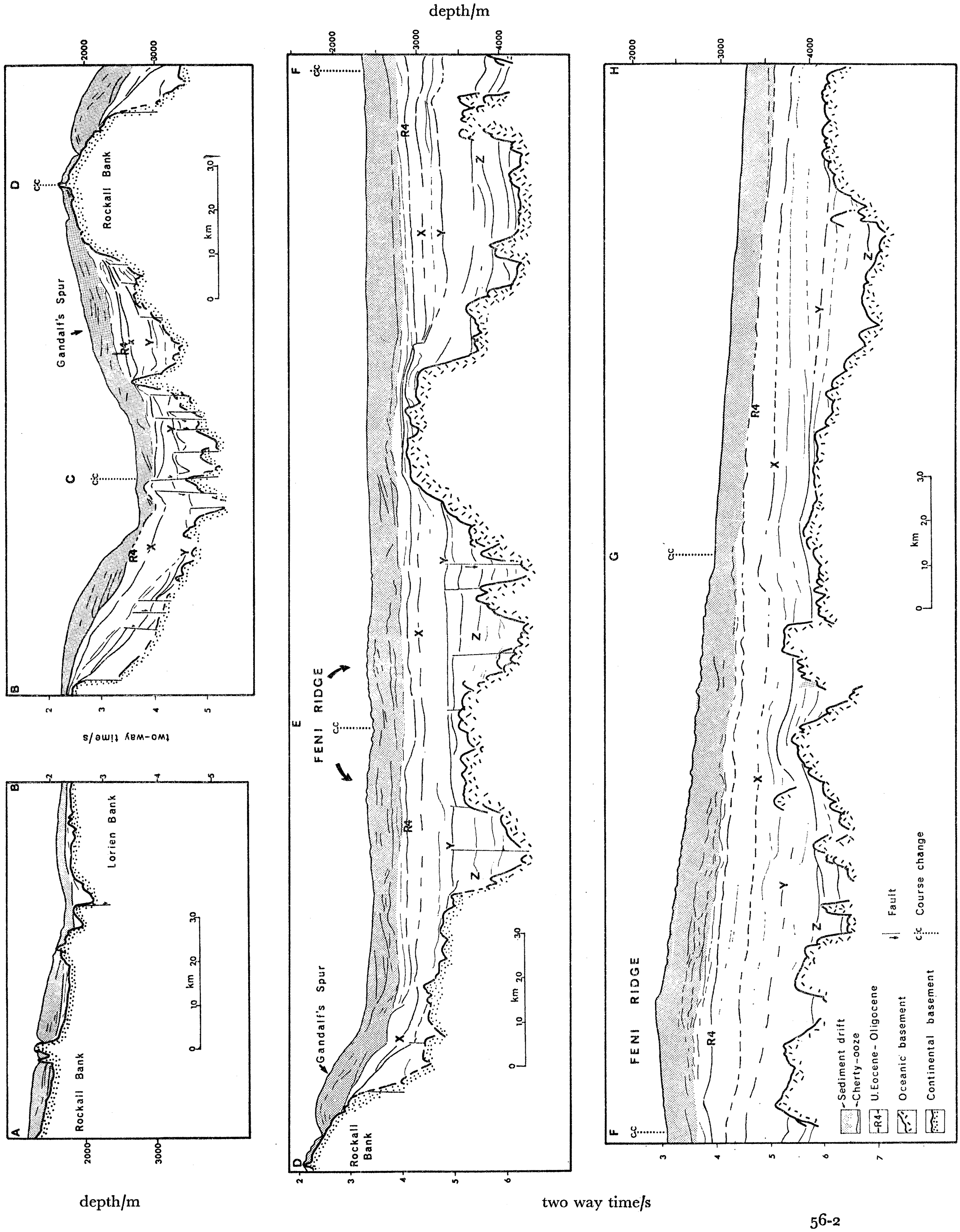

FIGURE 32. Seismic reflexion profiles across the Feni Ridge and Porcupine Fracture Zone. Profiles located in figure 31.

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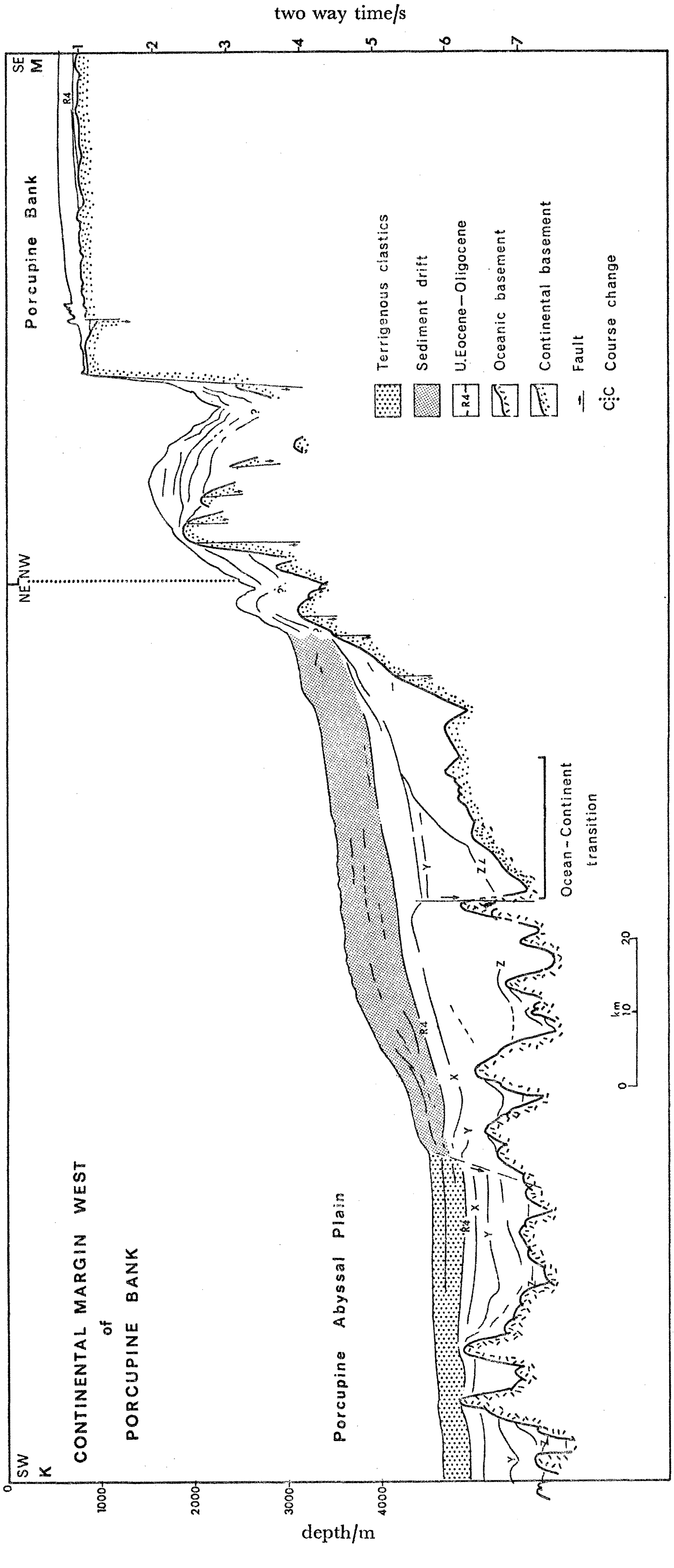


FIGURE 33. Seismic reflexion profile across the west margin of Porcupine Bank. Profile located in figure 31.

hitherto poorly understood. West of 21° W, it is a clearly defined WNW–ESE trending ridge and trough structure (Fleming *et al.* 1970). East of 20° W, it is largely buried (Cherkis, Fleming & Massingell 1973*b*) but can be followed ENE as far as 16° W (Roberts *et al.* 1973*a*). However, to the east of 16° W, the basement trends N–S and no structure comparable in dimension or character is present (figure 34). Magnetic anomaly maps and identifications (Scrutton *et al.* 1971; Vogt & Avery 1974) show the truncation of the fracture zone occurs at anomaly 32 (76 Ma). Between Porcupine Bank and anomaly 32 (figure 33) there is approximately 100 km of basement that is oceanic in aspect and must therefore be pre-76 Ma in age. An important reflector called Y pinches out against anomaly 32 crust but can be followed northward into Rockall Trough and eastward towards Porcupine Bank (figures 33 and 34). Although estimates of basement age derived from sedimentation rates are extremely speculative, a V_p of 3.2 km s^{-1} and an assumed sedimentation rate of *ca.* 10 cm per 1000 years suggests the pre-anomaly 32 crust is most probably about Lower Cretaceous–Upper Jurassic in age. A Triassic age is precluded because of the required low sedimentation rate. In support, Jones & Roberts have observed that the pre-32 anomaly sequence is correlative with that of the Bay of Biscay considered by Dewey, Pitman, Ryan & Bonnin (1973) to be 110–145 Ma. These results suggest formation of oceanic basement in Rockall Trough during Upper Jurassic–Early Cretaceous (Cimmerian?) time.

In the northern Rockall Trough, continuity of the Jean Charcot Fault Zone is interrupted by the Anton–Dohrn seamount though it may be represented by the scarp east and northeast of George Bligh Bank (figure 35). The northeastward trend of the 2.0 s isopachs suggests continuation of the oceanic Rockall Trough toward the Wyville–Thomson Ridge and Faeroe–Shetland Channel where basement depths and trends are comparable (Talwani & Eldholm, 1972). These data and the continental crust beneath the Faeroes (Bott *et al.* 1974) suggests the Faeroe–Shetland Channel may be the northward prolongation of Rockall Trough. The Wyville–Thomson Ridge is apparently composed of 'basement' divided into two ridges (Ewing & Ewing 1958; Ellett & Roberts 1973; Himsworth 1973) and speculatively may be a rejuvenated fracture zone. Note that the Wyville–Thomson Ridge and Porcupine Fracture Zone are not parallel to the Gibbs Fracture Zone (figure 21).

Between Rockall Bank and George Bligh Bank, there is no platform corresponding to that further south (figure 24) and the continuity of the slope is interrupted by the E–W graben mentioned previously. East of Rockall Bank, a series of NE–SW trending faults progressively downthrow the basement to the antithetically faulted platform (figures 29 and 30). Possible offsets in the 2.0 s isopach may be related to the Grenville Front or to the pre-lava fault at 57° 20′ N (figures 18 and 21).

The basement structure of the shelf comprising the east margin is summarized in figure 21. The NE–SW trending Caledonian mobile belt is the dominant feature and is separated by the Moine thrust plane from a northwest foreland composed of Scourian and Laxfordian granulite belts overlain by Late Precambrian and Cambro-Ordovician sediments. Both the foreland and Caledonian belt are cut by NE–SW trending Late Caledonian shears of which the best known are the Great Glen and Minch Faults. Syn-depositional Mesozoic dip-slip faulting along these shears has lead to the development of thick Permo-triassic and Jurassic sedimentary basins (Steele 1971; McQuillin & Binns 1973). Other important faults include the Highland Boundary and Southern Uplands Faults. In southwest Ireland, the WSW–ENE Hercynian orogenic belt is the dominant structure.

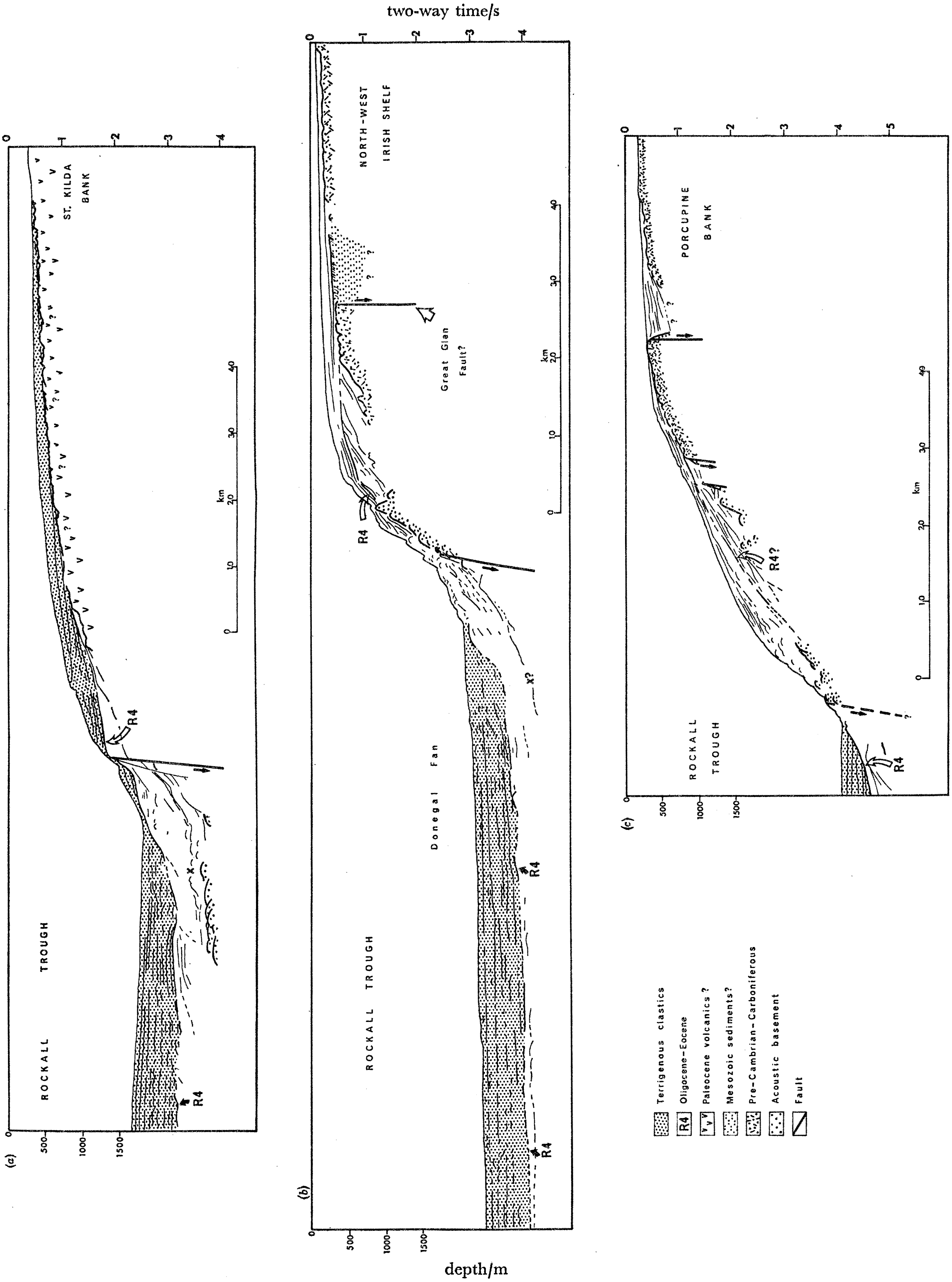


FIGURE 36. Seismic reflexion profiles (60 kJ arcer) across the continental margin west of the U.K.

In general, the basement structure of the slope and outer shelf consists of a series of down-faults occurring over the 50 km distance between the shelf edge and continent ocean boundary. Such faults are apparently responsible for the steep linear slopes west of St Kilda and Porcupine Bank (figures 21 and 33). Porcupine Seabight is controlled by N–S parallel faults (figure 21) that suggest an origin by orthogonal extension rather than rotation (cf. Bailey *et al.* 1971).

No obvious relation exists between the form of the margin and the trend of the Laxfordian, Scourian, and Caledonian belts. A relation between the Hercynian front and the Gibbs Fracture Zone postulated by Cherkis *et al.* (1973 *b*) is not supported by the truncation of the Gibbs at anomaly 32 and the absence of any associated offset in the margin. A palimpsest relation may exist however between the rift fabric and the Caledonian shears that are sub-parallel to various segments of the margin. The later Mesozoic sedimentary basin development may be related to the early rifting phase of the Rockall Trough. However, the margin clearly cuts across these basins in several cases. The most convincing relation seems to exist at 55° 30′ N where the Great Glen Fault may intersect the slope (Riddihough 1968; Roberts 1970; Bailey *et al.* 1974). The basement isopachs (figure 18) show a 25 km offset that does not continue deeper than the 2.0 s isopach or continent/ocean boundary. In view of the inferred Upper Jurassic–Late Cretaceous age of Rockall Trough, previously postulated Mesozoic and Tertiary dextral, strike-slip along the Great Glen Fault is untenable (Holgate 1969; Garson & Plant 1972). Evidence of an east–west fault downthrowing to the north (figure 36) at the offset (Bailey *et al.* 1974) is also inconsistent with southerly downthrow (McQuillin & Binns 1973) on the Great Glen Fault. These inconsistencies can be rationalized if the offset is associated with a fracture zone initially localized by the Great Glen Fault and thus post-dates the Permo-Jurassic sedimentary basin development of the Hebridean area (Steele 1971; Hudson 1964). On Porcupine Bank, evidence of a palimpsest control may be shown by a prominent ENE–WSW trending magnetic lineament (the Highland Boundary Fault?) that crosses the shelf and Bank (Vogt & Avery 1974). The linearity of this feature (figure 21) also precludes rotation of Porcupine Bank.

In view of the non-congruency of the opposite margins of the Rockall Trough, shown by the 500 fathom contours, it is worth comparing the congruency of the continent/ocean boundary defined by the Jean Charcot Fault Zone and 2.0 s isopach. North of the Porcupine Fracture Zone, a parallel sided sinuous congruent depression is so defined and a similar congruency may exist to the south. The continent/ocean boundaries thus have a substantially different plan to that implied by the bathymetry. A satisfactory closure of the Rockall Trough can be made at these continent/ocean boundaries and precludes rotation of Porcupine Bank (Roberts *et al.* 1973 *a*). This closure also implies on geometric grounds a northward continuation of the Rockall Trough via the Faeroe–Shetland Channel into the Norwegian Sea.

(ii) *The pre-R4 series*

The pre-R4 series in the Rockall Trough includes reflectors absent on and to the west and southwest of Rockall Plateau. It is thickest between the Jean Charcot Fault Zone and the east margin of Rockall Trough (figures 29 and 30). The three important reflectors in the series, here called X, Y and Z are most clearly observed in the southern Rockall Trough though their identification in the northern Rockall Trough is tentative because of the poorer quality seismic data and lack of tie-lines.

Reflector Z occurs only in the deepest parts of Rockall Trough. Within the axial part of the

Trough, it is not deformed and appears to be draped over the oceanic basement. Towards the margins however it is flat-lying suggesting deposition by gravity fill (figures 32, 33 and 34). Sediments between Z and basement exhibit two types of stratification. Towards the continent ocean boundary, the lowest pelagic reflectors are draped over the oceanic basement and this buried relief has been infilled by flat-lying sediments. At the foot of the marginal plexus however, the attitude of the weak reflectors and the radical increase in thickness suggest fan deposition (figure 34). A fan may also be developed at the foot of the Jean Charcot Fault Zone (figure 29). The pre-Z sequence also thins toward the Trough axis. These relations suggest

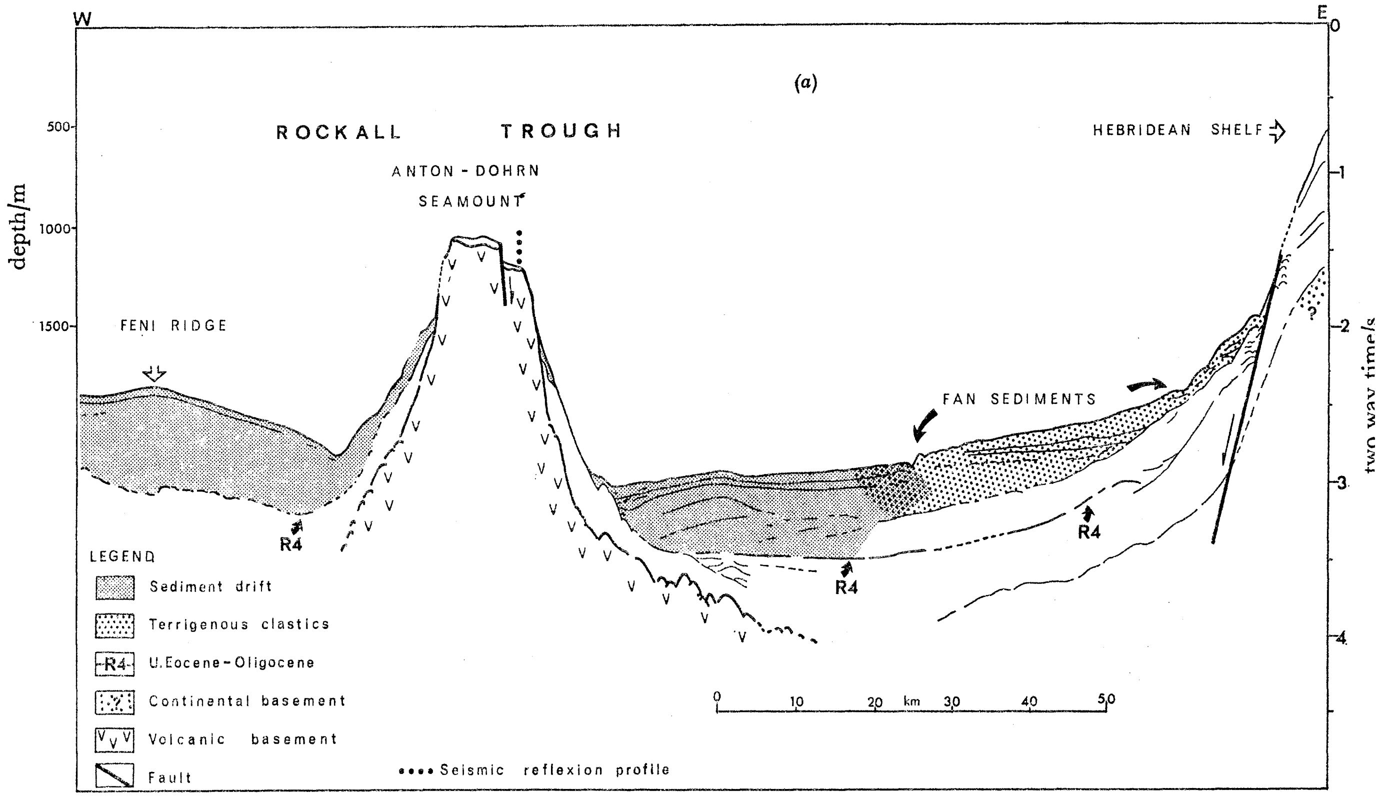


FIGURE 37 (*a*). For description see opposite.

that during the spreading of the Rockall Trough, the earliest pre-Z pelagic sediments were draped over the young oceanic basement while fans developed at the base of the marginal plexus to subsequently prograde westward, infilling the proximal oldest oceanic basement. Although precise stratigraphic data on the age of 'Z' do not exist, reflector 'Z' underlies reflector 'Y', previously argued to be 76 Ma. A mean intra-Z–Y thickness of *ca.* 0.5 s, sedimentation rate of 5 cm per 1000 years and a V_p of 3.0 km s^{-1} suggests a 100 Ma age for reflector 'Z'. Although such estimates are notoriously unreliable, it plausibly coincides with the 'Cenomanian' transgression and suggests spreading had ceased by 100 Ma.

Reflector Y pinches out on basement dated at anomaly 32 (76 Ma) and may therefore be about this age. The intra-Z–Y sediments are acoustically transparent and little can be said concerning their depositional history. Within the Trough, reflector Z and Y form a conformable sequence exemplified by draping and thinning of the intra-Z–Y sequence over the buried basement. Reflector Y pinches out against the basement comprising Anton–Dohrn and

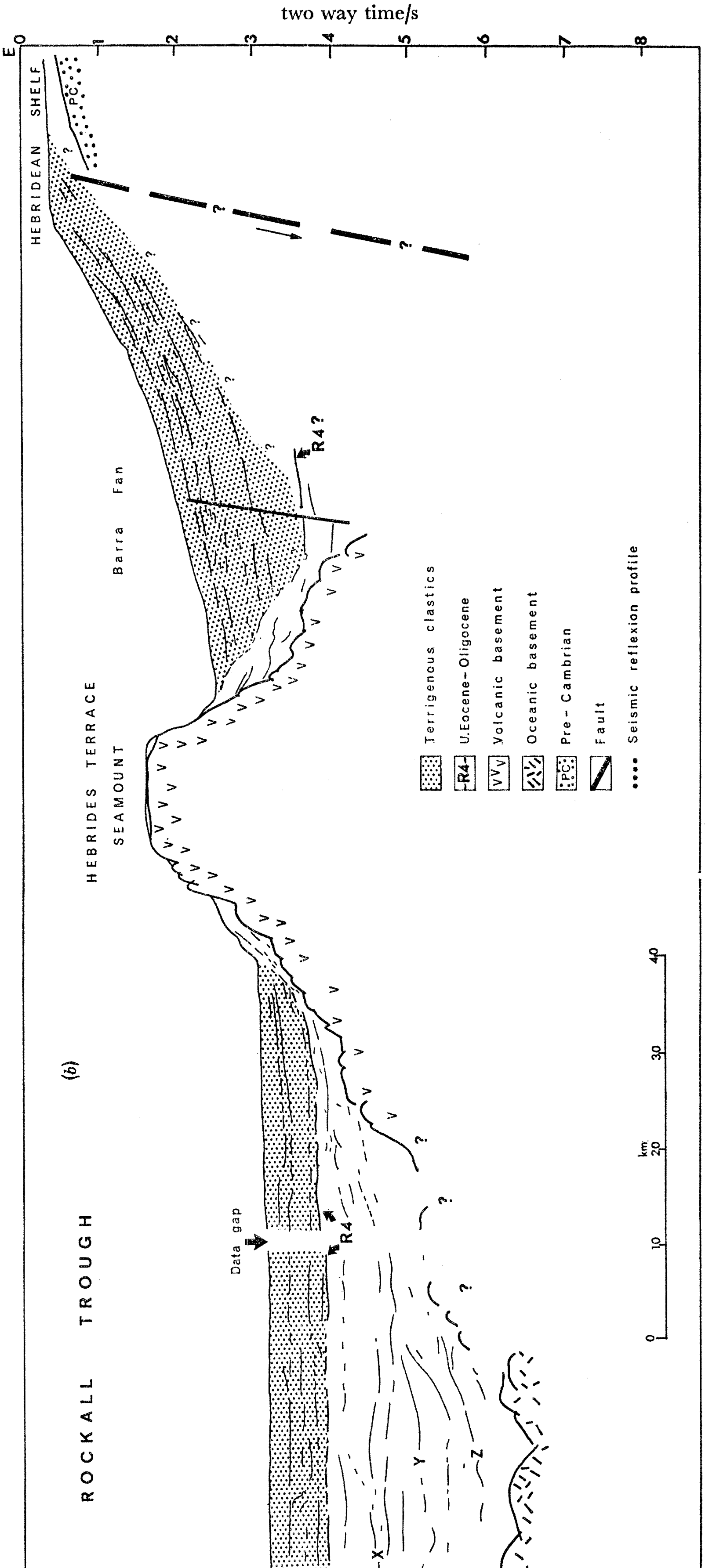


FIGURE 37(*a*). Sparker reflexion profile across Anton–Dohrn Terrace Seamount. (*b*), Air-gun reflexion profile across Hebrides Terrace Seamount.

Hebrides Terrace seamounts (figure 37*b*) thus indicating a pre-76 Ma age independently supported by a palaeomagnetic study of Rosemary Bank (Scrutton 1971). Beneath the east margins of the Trough, Y locally pinches out against Z but also overlaps it to subcrop against basement (figures 30 and 33). Adjacent to the Jean Charcot Fault Zone, little deformation of Y is evident and suggests the absence of subsequent deformation across the continent/ocean boundary. The possible presence of reflector Y in the grabens of the adjacent platform at the same stratigraphic level may indicate deposition did not begin there until between 76 and 100 Ma (figure 30). South of the Porcupine Fracture Zone on the margin of the Lorien Bank, the pattern of deposition and deformation is somewhat different. The intra-Z–Y sequence is thicker and the geometry of the reflexions suggests fan development (figure 32). Within this area, faulting of both Z and Y is present and is most intense adjacent to Lorien Bank and the south Rockall Bank. The NE–SW trending faults downthrow to the north by *ca.* 0.5 s and are thus antithetic to the marginal plexus (figures 21 and 32). Although faulting on the adjacent ocean crust is less intensive, regional warping of reflector Y is evidenced by the marked unconformity with reflector X (e.g. figure 32). Faulting is also present adjacent to the Gibbs Fracture Zone and anomaly 32 (figure 34) but diminishes northward and eastward. The pre-X faulting is therefore confined to the vicinity of Lorien Bank and the Gibbs Fracture Zone. This distribution may reflect the major reorganization in North Atlantic plate motions at 76 Ma that resulted in rifting and spreading between the southwest margin of Rockall Plateau and North America, and in the development of the Gibbs Fracture Zone. Contemporaneous uplift and erosion of Lorien Bank may be indicated by the syntectonically faulted and thickened sediments.

Reflector 'X' situated between reflector Y (76 Ma) and R4 (37 Ma), may be 60 Ma in age and correlative with the reflector 'X' observed southwest of Rockall Plateau. In the northern axial Rockall Trough, X, Y and R4 form a conformable sequence (figure 30). East of Lorien Bank, X is unconformable upon Y but conformable with R4 (figure 32). Towards the margins of the Trough, X pinches out against Y and is itself cut out by R4 (figures 29 and 30). Beneath the upper slope and outer shelf, a prominent unconformity separates an older prograding sequence resting on basement or the Mesozoic (Permo-Triassic to Jurassic) sediments from a younger prograding sequence (Stride *et al.* 1969; Bailey *et al.* 1974). This unconformity corresponds to R4 in one case (figure 37*a*) but in others may correspond to X thus indicating a prolonged hiatus from 60 Ma to post-R4 time. It therefore seems probable that at least part of the older prograding series may be pre-60 Ma. In support, Bailey *et al.* (1974) have reported derived Senonian forms close to an outcrop of the series. A common feature of profiles is the marked thickening of the intra-R4-x sequence at the base of the slope that suggests increased sediment deposition between 37 and 60 Ma.

(iii) *R4 and the post-R4 series*

Reflector R4 is a flat-lying reflector at depths of *ca.* 0.7 s previously called R by Jones *et al.* (1970) and identified as Tertiary–Cretaceous by Stride *et al.* (1969). Correlation of R4 with the Hatton–Rockall Basin and Reykjanes Ridge has been discussed earlier. As in these areas, R4 is associated with numerous overlapping hyperbolae that originate in the overlying 0.2–0.3 s of section and impart a characteristic crinkled appearance to the overlying beds. On the west side of Rockall Trough, impersistent cuspate reflectors characterise much of the post-R4 series (e.g. figure 29). The acoustic sequence is closely comparable to that in the Hatton–

Rockall Basin. R4 is thus Upper Eocene–Late Oligocene in age and the post-R4 series consists of Oligocene cherts overlain by Miocene to Recent sediments.

Within the central parts of the Rockall Trough, reflector R4 and the Oligocene cherts are conformable with reflector X. Beneath the east slope, R4 subcrops against X and is overlapped by the post-R4 series. The Oligocene cherts also pinch out against R4 and are themselves overlapped by younger sediments. Relations between basin and outer shelf–upper slope stratigraphy are obscured by slumping, side-echoes and thin sediments. The prominent shelf unconformity mentioned earlier is almost certainly correlative with R4 to the north of St Kilda and subcrops against the Palaeocene basement proximal to St Kilda. A prominent reflector that may be R4 underlies the Barra Fan and may also be present beneath the shelf (see, for example, Bailey *et al.* 1974, Fig. 2, profile 87-86). Bailey *et al.* (1974) show unconformities within their older series that may be correlative with reflector X. The existence of these unconformities and the correlation observed north of St Kilda suggests downwarping may have occurred in two phases at 60 and 37 Ma.

The Oligocene cherts are *ca.* 0.2–0.3 s thick and largely confined to the basinal parts of the Trough. Beneath the Feni Ridge, the post-Oligocene sediments are unconformable upon the cherts though the sequence is conformable to the east. Variations in the thickness of the cherts reflect contrasting depositional processes on the east and west sides of the Trough. Beneath the marginal channels associated with the seamounts and Feni Ridge sediment drift, the cherts are thinned and locally absent due to erosion (figure 32). Beneath the east side of the Trough, the crinkled aspect is subdued suggesting dilution by increased sediment supply (figure 29). Preservation adjacent to Rockall Plateau suggests lack of sediment supply and accords with the geological evidence of submergence by this time (Laughton, Berggren *et al.* 1972). There is also some indication that the cherts are absent beneath the Porcupine Abyssal Plain although this may reflect dilution by other sediments (figure 34).

The post-Oligocene stratigraphy is markedly different beneath the east and west sides of the Trough and emphasizes and reflects the contrasting physiography reviewed earlier. The post-chert sequence to the west was sedimented by differential deposition in contrast to the fan deposition that predominates to the east.

Between the Wyville–Thomson Ridge and 53° N, the isopachs and seismic profiles show the Feni Ridge is underlain by a triangular or lensoid body of sediments extending to the axis of maximum depth in Rockall Trough. These sediments are unconformable with the cherts and seabed. The unconformity with the cherts indicates differential deposition began in post-Oligocene–Early Miocene time.

The Feni Ridge sediment drift begins at the gap between the Wyville–Thomson and Ymir Ridges (Ellett & Roberts 1973). Along the sediment drift, there are systematic variations in thickness and physiography that reflect temporal and spatial variations in differential deposition. In general there seems to be a close relation between drift thickness, sea-bed physiography marginal channel dimensions and the trend and gradient of the adjacent slope. A rather unique example is shown by the systematic clockwise decrease in sediment thickness beneath channel and ridge around Anton–Dohrn and Rosemary Bank seamounts (Roberts *et al.* 1974). At 59° N, a weakly developed ridge is underlain by 0.75 s of sediment and the seabed is characterized by the small scale roughness discussed earlier (figure 35). Further south, as the east slope of George Bligh Bank steepens, enhanced differential deposition and perhaps erosion is indicated by the truncation of R4 and virtual absence of sediments beneath the marginal channel; erosion over a depth range of 2000 m may also be shown by reflectors truncated against the stepped slope of George Bligh Bank (figure 35). South of Anton–Dohrn seamount, the drift thickens and the impersistent cuspate reflectors accurately mirror undulations in the seabed (figure 29) though these features do not persist to 55° 30′ N (cf. figures 29 and 30). Between 55° 30′ N and 53° N, there is a remarkably close relation between drift thickness, physiography and slope trend. As the Feni Ridge diverges from the margin, there is a progressive southward increase in thickness to 1.25 s (figure 27) followed by a subsequent decrease to the south (although the drift is still thicker). On the east side of the drift, undulations concomitantly developed persist down section (figure 32) but do not continue laterally beyond an abrupt physiographic boundary with a smooth relief typical of uniform pelagic sedimentation. In contrast, no undulations are developed west of the ridge crest and the marginal channels are underlain by only 0.5 s of sediment, about half the thickness beneath the ridge axis. The stepped inner walls of the channels north and south of Gandalf's Spur (figure 32) indicate erosion although the northern channel terminates in the thicker sediments of the semicircular depression. The sinuosity of the ridge, the appearance and disappearance of marginal channels, the semicircular depression all seem closely related to Gandalf's Spur. The seismic profiles show Gandalf's Spur is a sediment drift, built on east–west trending basement, that has grown eastward and outward across the regional trend of the slope and the adjacent marginal channel. Erosion in 900 fathoms depth compared to 1300 fathoms depth in the marginal channel is indicated by the stepped slope north of the spur. Development of Gandalf's Spur seems closely related to the adjacent non-depositional channel that originates between Rockall Bank and Lorien Bank (figure 32) and continues downslope to the head of the channel at the foot of Gandalf's Spur.

To the south of 53° N, the Feni Ridge turns northward around Lorien Bank decreasing in relief to merge with the drifted sediments of the southwest margin. A long spur extending south of 52° N may also be an extension of the Feni Ridge. A sediment drift may be present west and north of Porcupine Bank.

Evidence of a change in depositional regime is given by a change in the attitude of reflectors and migration of the ridge crest. In the northern Rockall Trough, sub-surface undulations not present in the seabed appear at about 0.5 s depth (figure 35). South of Anton–Dohrn seamount (figure 29), the sub-surface undulations first appear at this depth but continue up section. Beneath the marginal channel, unconformities suggest migration of the Feni Ridge axis accompanied by erosion. This event may correspond to the Late Miocene change in bottom current regime observed on the Rockall Plateau and Reykjanes Ridge. Finally, Pleistocene sedimentation rates on the Feni Ridge axis are *ca.* 15 cm per 1000 years (McIntyre & Ruddiman 1972) compared to the mean Post-Oligocene rate of *ca.* 3 cm per 1000 years.

On the east margin of the Trough, a series of coalesced fans best developed in the uppermost 0.5 s of section are present between 58° and 54° N. Within the fans, numerous cut-outs and channels are present. The optimum development of the Barra and Donegal Fans around the Hebrides Terrace Seamount reflects the greater sediment supply available along canyons and the shelf transport paths that are there oriented perpendicular to the shelf edge. The sub-surface relation of the fans to the Feni Ridge drift is not clear. Topographic evidence (figure 9) suggests axial transport has inhibited eastward extension of the drift. Most probably, fan and drift sediments interfinger at a scale too fine to be resolved by the seismic reflexion profiles. The fans do not extend southwest of Porcupine Bank (figure 9).

The preservation of the two contrasting sedimentation regimes is primarily a function of sediment supply. To preserve cherts and the Feni Ridge and the drift west of Porcupine Bank, little sediment can have been available from Rockall Bank and Porcupine Bank in post-Upper Eocene time. This conclusion implies that both areas were probably below sea level at that time. Palaeoenvironmental JOIDES data on the deeper western part of Rockall Bank support this conclusion (Laughton, Berggren *et al.* 1972). In contrast, post-Upper Eocene downwarping of the shelf accompanied by relative uplift and erosion in Scotland has been responsible for fan growth and shelf progradation to the east.

(*e*) *Porcupine Bank and Seabight*

Although only limited deep reflexion data are available for Porcupine Seabight, it is useful to compare the stratigraphy with that of Rockall Trough.

Within the Seabight, there are three prominent reflectors here called X, Y and R4 that may be correlative with those in the Rockall Trough. All three are not faulted and bear an onlapping relation to each other that indicates progressive subsidence of the Seabight and its margins (figure 13).

One feature of particular interest is the striking change in depositional regime above R4 that may be related to that observed in the Rockall Trough. The more numerous reflectors indicate an influx of terrigenous material that can only have been derived from uplift of the Irish area. Beneath the west margin of the Seabight, the lower part of the post-R4 sequence is prograded but the upper part is in contrast acoustically transparent suggesting decreasing sediment supply due to the submergence of Porcupine Bank. Bailey *et al.* (1974) have reported folded sediments within the NE–SW trending Slyne Trough (figure 21) beneath the Northern Seabight. Folding is not apparent on the deep seismic profiles. Possibly, the Slyne Trough may be an early Mesozoic basin analogous to the Minch.

4. Patterns of Post-Palaeogene sedimentation

The common feature of Post-Oligocene stratigraphy is the widespread predominance of differential deposition in the Rockall Trough, on Rockall Plateau and on the east flank of the Reykjanes Ridge (figure 9). The pervasion and conservation of differential deposition is undoubtedly due to the absence of significant sediment supply from the submerged Rockall Plateau.

A number of problems remain however. These include the origins of the Post-Oligocene differential deposition and the Late Miocene change in depositional regime. Furthermore, their occurrence coupled with the continuum of differential deposition over depth ranges of 1000–4000 m must be examined in the perspective of present and past ocean circulation. The origins of the variations in physiography, erosion and deposition rate across the drifts should also be examined in this perspective.

The geologically important role of deep ocean circulation in depositing and distributing sediments has been known for sometime (see, for example, Heezen & Hollister 1964; Hollister & Heezen 1972; Jones *et al.* 1970; Le Pichon *et al.* 1971*a*, *b*). In the present North Atlantic Ocean, saline surface water flows northward into the Norwegian Sea where it cools and becomes denser (figure 38). This water sinks to ultimately overflow back into the North Atlantic Basins via the Denmark Strait, Iceland–Faeroes Ridge, Faeroe Bank Channel and Wyville–Thomson

Ridge (Worthington 1970). After debouching from the Faeroe Bank Channel and overflowing the Iceland–Faeroes Ridge, the Norwegian Sea overflow water with sediment in suspension initially flows WSW and then southwestward along the Reykjanes Ridge where it mantles the Gardar Ridge (Crease 1965; Meincke 1972; Lee & Ellett 1965, 1967; Steele *et al.* 1962; Johnson & Schneider 1969; Davies & Laughton 1972; Ruddiman 1972). In the case of the Rockall Trough, intermittent overflow across the Wyville–Thomson Ridge is associated with the Feni Ridge (Ellett & Roberts 1973). The clear first order relation between the latter drifts

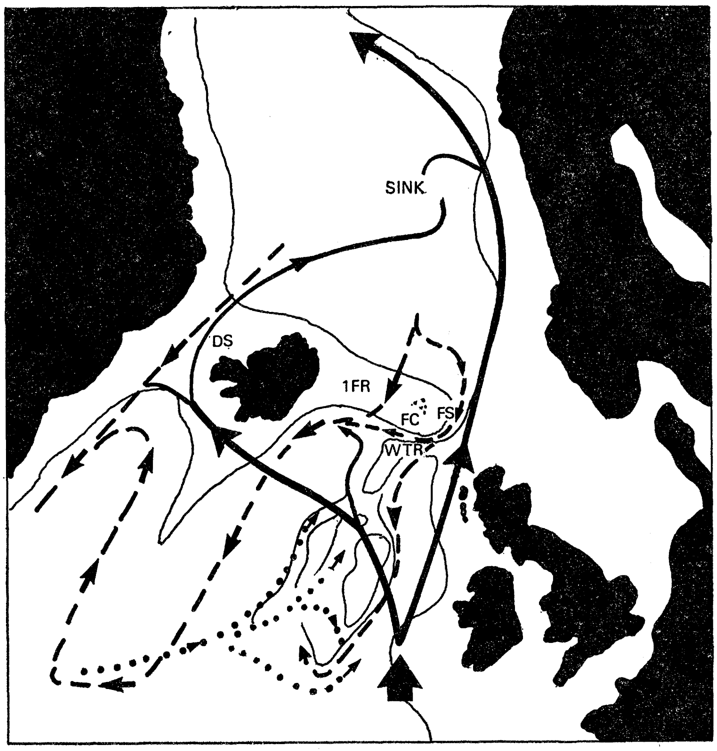


FIGURE 38. Northeast Atlantic circulation. Heavy solid lines, inflow; dashed line, Norwegian Sea overflow; dotted line, Labrador Sea Water (after Worthington 1970; Ellett & Martin 1973).

and the regional bottom flow (cf. figures 38 and 9) demonstrates that the present day circulation is essentially Post-Oligocene in age although the thickening of the pre-R4 sequence on the Reykjanes Ridge may indicate an earlier phase (Ruddiman 1972). The onset of differential deposition was, however, preceded by deposition of biogenous silica now represented as cherts (site 116, Laughton, Berggren *et al.* 1972). Biogenous silica deposition contrasts with the general predominance of carbonate deposition in Upper Eocene and Post-Oligocene time and is also at variance with the present low surface silica values of the North Atlantic. Berger (1970) has compared silica values in the Pacific and Atlantic Oceans and has suggested high silica production is associated with an 'estuarine', i.e. closed basin circulation. The data may thus suggest a change in Atlantic circulation from open to closed in Eocene–Oligocene time possibly due to partial closure of the Panama isthmus and the Tethys (Malfait & Dinkelmann 1972; Dewey *et al.* 1973). The subsequent return of open circulation conditions may be related to subsidence of the Iceland–Faeroes Ridge below sea level estimated at either 30 or 10 Ma, which would permit exchange of water with the Norwegian Sea and the establishment of the present day circulation (Vogt 1972). Vogt's 30 Ma estimate is not supported by the stratigraphic evidence although the 10 Ma date is in broad agreement with the Late Miocene change in depositional regime. This model may be reconciled with the stratigraphic data if the earlier outflow came through the Faeroe Bank, Faeroe–Shetland Channel system as the area subsided. In support, seismic evidence suggests an open seaway (Stride *et al.* 1969) and the present sill depth of the Faeroe Bank channel is 350 m greater than the Iceland–Faeroes Ridge. Subsequent subsidence of the Iceland–Faeroes Ridge below sea level at 10 Ma would have allowed a greater exchange of water with the Norwegian Sea resulting in a more vigorous circulation perhaps evidenced by the lateral migration of sediment ridges. Finally, the greater thickness of the Gardar Ridge compared to the Feni Ridge suggests transport across the Wyville–Thomson Ridge has been volumetrically less significant.

Studies of the present day distribution of Norwegian Sea Water (Lee & Ellett 1965, 1967; Worthington & Wright 1970) show it is absent on the Rockall Plateau and is confined to the adjacent ocean basins. The source of the bottom flow responsible for continuing differential deposition on Rockall Plateau is thus of some interest. Hydrographic studies show the Norwegian Sea Water is overlain by Labrador Sea Water typified by distinctive salinities and temperatures (Lee & Ellett 1967). The water mass overlying the southwest Hatton–Rockall Basin and the Hatton Drift can thus be identified as Labrador Sea Water (Dietrich 1969; J. Crease, private communication). To reach these areas, the Labrador Sea Water must turn northward after flowing eastward over the Mid-Atlantic Ridge (figure 38). In the Hatton–Rockall Basin, the northward decrease in thickness of differentially deposited sediments supports a northward flow. However, a divergent southeastward flow may be also be implied by the non-depositional channel in the southeast Hatton–Rockall Basin and the presence of Labrador Sea Water in the Rockall Trough (Ellett & Martin 1973). Alternatively, differential deposition in the Hatton–Rockall Basin may reflect interaction of the surface Irminger current with the incoming Labrador Sea Water (Ruddiman 1972).

A thorough understanding of the sedimentary processes responsible for the growth of these sediment drifts is impeded by the lack of quantitative oceanographic data and the coarse resolution of seismic profiles. Typically, the resolution is of the order of 10^6 Ma in contrast to the transcience of ocean circulation events (*ca.* 1 a) and the effects of circulation changes likely to have occurred in glacial and interglacial periods (*ca.* 10^4–10^5 a). The problem is well illustrated on the Feni Ridge by the difference between Pleistocene (*ca.* 15 cm per 1000 years) and the mean post-Oligocene (*ca.* 3 cm per 1000 years) deposition rates. Using similar reasoning, Worthington & Wright (1971) have queried the use of sediment drifts as palaeocirculation indicators. However, the correlation between drift distribution and bottom circulation does demonstrate the continued existence of a bottom circulation similar in pattern though not necessarily in intensity to that observed today.

The more important problems arise from the apparent stability of drifts and channels for at least 10 Ma and their widespread growth on a flat substrate. The dimensions and complexity of the drifts and channels around Edoras Bank do not accord with the wide regional flow shown by the hydrographic data (cf. dimensions of 20 and 10^3 km). Qualitative observations regarding scales of differential deposition are important in this respect. In the Rockall Trough, differential deposition has occurred over distances between ridge and channel axis of 50–100 km. Yet the undulations east of the ridge axis demonstrate differential deposition operating parallel to the drift axis on a horizontal scale of 1 km. Comparative vertical scales are *ca.* 400 m and 50 m respectively. Further, the abrupt boundary between undulations and smooth relief suggests an invariant limit to the differential depositional regime. Closely comparable features though smaller in scale are present in the Hatton–Rockall Basin. Erosion extending over 2000 m east of George Bligh Bank does not accord with near bottom depth of Norwegian Sea Water (Ellett & Martin 1973; Ellett & Roberts 1973) or indeed the slow current velocities (*ca.* 5 cm s^{-1}) recorded in the Rockall Trough (R.R.S. *Shackleton*, unpublished data cf. 15–35 cm s^{-1} experimental critical erosion velocities of Southard, Young & Hollister (1971)). These observations suggest the whole water column and indeed the regional flow may be perturbed in a way that ensures independence from short term changes in circulation and therefore depositional stability. The seismic and topographic data have empirically suggested that the topography of the underlying basement or substrate is the common factor. For example, the Gardar Ridge drift seems to have initially grown against the 45 Ma basement scarp and the offsets in its axis accurately reflect underlying fracture zones offsets of the basement scarp (Ruddiman 1972). In the Rockall Trough and Hatton–Rockall Basin there are similar relations. A number of mechanisms have been offered to explain drift and channel development. Davies & Laughton (1972) suggested secondary currents generated by seamounts and ridges might result in ridges and channels along which high potential density water may now flow. Langseth & Boyer (1972) invoked outward directed flow in an Ekman boundary layer to explain the absence of sediment beneath the marginal channel; such a mechanism may be applicable to initial growth of the Gardar Ridge. Internal tides have been invoked to explain erosion on Horizon guyot (Lonsdale, Normark & Newman 1972). Roberts *et al.* (1974) have postulated that Taylor column development atop Anton–Dohrn seamount is responsible for the asymmetric sediment distribution.

Laboratory experiments performed by Hide & Ibbetson (1966) and Hide (1968) show unusual water velocity distributions that are compatible with many aspects of the ridges and channels and are independent of water mass distribution. In their experiments, a current impinging on a scarp was accelerated (non-deposition and/or erosion) and then deflected back into the main flow at the end of the scarp; the current passing over the scarp was decelerated (deposition). By analogy, intersection of the southwest flowing Norwegian Sea Water by the offset N–S basement scarp may have led to the sinuous growth of the Gardar Ridge. Similarly in the Rockall Trough, accelerated water adjacent to George Bligh Bank is evidenced by erosion, non-deposition and the small-scale bottom roughness typically associated with faster currents (Schneider *et al.* 1967). Further south the relation between Gandalf's Spur, the non-depositional channel between Lorien Bank and Rockall Bank, and the sinuosity of Feni Ridge suggests intersection of two differentially depositing agencies that may be analogous to the source–sink flows modelled by Hide (1968). In his model (see Hide 1968, Fig. 12*c*), irregular features developed upstream of the sink and extending around it may be comparable to the pattern of undulations east of the Feni Ridge. The semicircular depression may correspond to the sink. In the case of Edoras Bank, it is difficult to conceive a mechanism responsible for the complex drifts and channels. Enhanced deposition on top of Edoras Bank may indicate Taylor column development and the ridges and channels effects comparable to these discussed above. It is highly desirable to obtain quantitative current and sedimentologic data to rigorously analyse the origin of these features and especially since their linearity indicates considerable value as palaeoenvironmental indicators in the geological record.

5. Perspectives of continental margin development

In the introduction, inhomogeneities in the lithosphere were considered to be important controls on rift margin development. The change in plate tectonics regime through time (Sutton 1973) suggests such inhomogeneities are a consequence of both the age and the history of the lithosphere. The continental lithosphere of Rockall Plateau and the British Isles has been progressively modified and generated during the Keltidian, Scourian, Laxfordian, Grenvillian, Caledonian and Hercynian orogenies. It is clear that the margins of the Rockall Plateau and Trough cut across the trends of these orogenic belts (Roberts *et al.* 1973 *a*; McQuillin & Watson 1973) suggesting that rift margin development is independent of the age of the lithosphere. Further, the low-angle thrust fabric of orogenic belts may inhibit development of the vertical, normal fault fabric of rift margins. The structures that have affected the continental margin and controlled sedimentary basin development include the Caledonian shears which must affect the lithosphere. The significant factor affecting the form of the rift margin may therefore be the presence or absence of a fault fabric affecting the whole lithosphere.

The influence and interaction of the pre-existing fault fabric and fault fabric imposed during the various rifting phases is more difficult to assess. West of the British Isles, initial basin formation in Permo-Triassic times was primarily due to regional extension with palimpsest control by Caledonian trends (Bott & Watts 1970, 1971; Hall & Smythe 1973). Renewed subsidence and uplift in Middle to Upper Jurassic times (Bacon & Chesher 1974) may reflect rifting and the formation of oceanic lithosphere in the Rockall Trough. Note however that basin trends are truncated at the margins. Crustal thinning that may be cumulative is evident beneath the Hatton–Rockall Basin and Porcupine Seabight (Matthews & Smith 1971). On Rockall Plateau, evidence of a fault fabric imposed during spreading of Rockall Trough is not extant. In the Hatton–Rockall Basin however, the north–south faulting and subsidence in contrast to the NE–SW faulting may reflect the response of the palimpsest fabric initially developed during formation of the Labrador Sea to later rifting and subsidence.

The subsequent history of the major sedimentary basins has been one of progressive and/or intermittent subsidence. Remarkably, uplift and subsidence are contemporaneous over a very wide area far from the edge of the accreting plate and apparently correlate with spreading rate changes at the mid-ocean ridge (figure 39). For example, the 40 Ma spreading rate change on the Reykjanes Ridge is correlative with the start of the post-Lower Oligocene subsidence of the Hatton–Rockall Basin, downwarping of the shelf west of the British Isles, subsidence in the Porcupine Seabight and uplift in Scotland and Ireland (George 1966, 1967) and perhaps with similar events in the English Channel and South Irish Sea (Roberts 1974). At 60 Ma, the initiation of spreading between Greenland and Rockall is matched by post-Palaeocene subsidence in the Hatton–Rockall Basin, warping of reflector X in the Rockall Trough and Porcupine Seabight and also by the intra-Cretaceous Palaeogene deformation in Scotland and Ireland (George 1966, 1967). A second problematic feature is the gentle warping that characterizes the Rockall Trough in contrast to the more intense contemporaneous deformation observed on the continental parts of the plate, e.g. Hatton–Rockall Basin.

Two models have been developed to explain basin and continental margin subsidence. Sleep (1971) has shown the margin cools and subsides with a time constant of about 50 Ma that is comparable to the subsidence of the oceanic part of the plate. On the other hand, Walcott (1971, 1972) and Bott (1971, 1972) have suggested differential loading across the margin causes flexure and faulting that is likely to develop parallel to the margin in continental crust. Bott (1971) also suggests the stress differences between oceanic and continental crust produce hot creep in the lower and middle continental crust causing it to thin by flow towards the oceanic upper mantle. The two mechanisms are not mutually exclusive since differential subsidence will be concurrent with thermal subsidence of the margin.

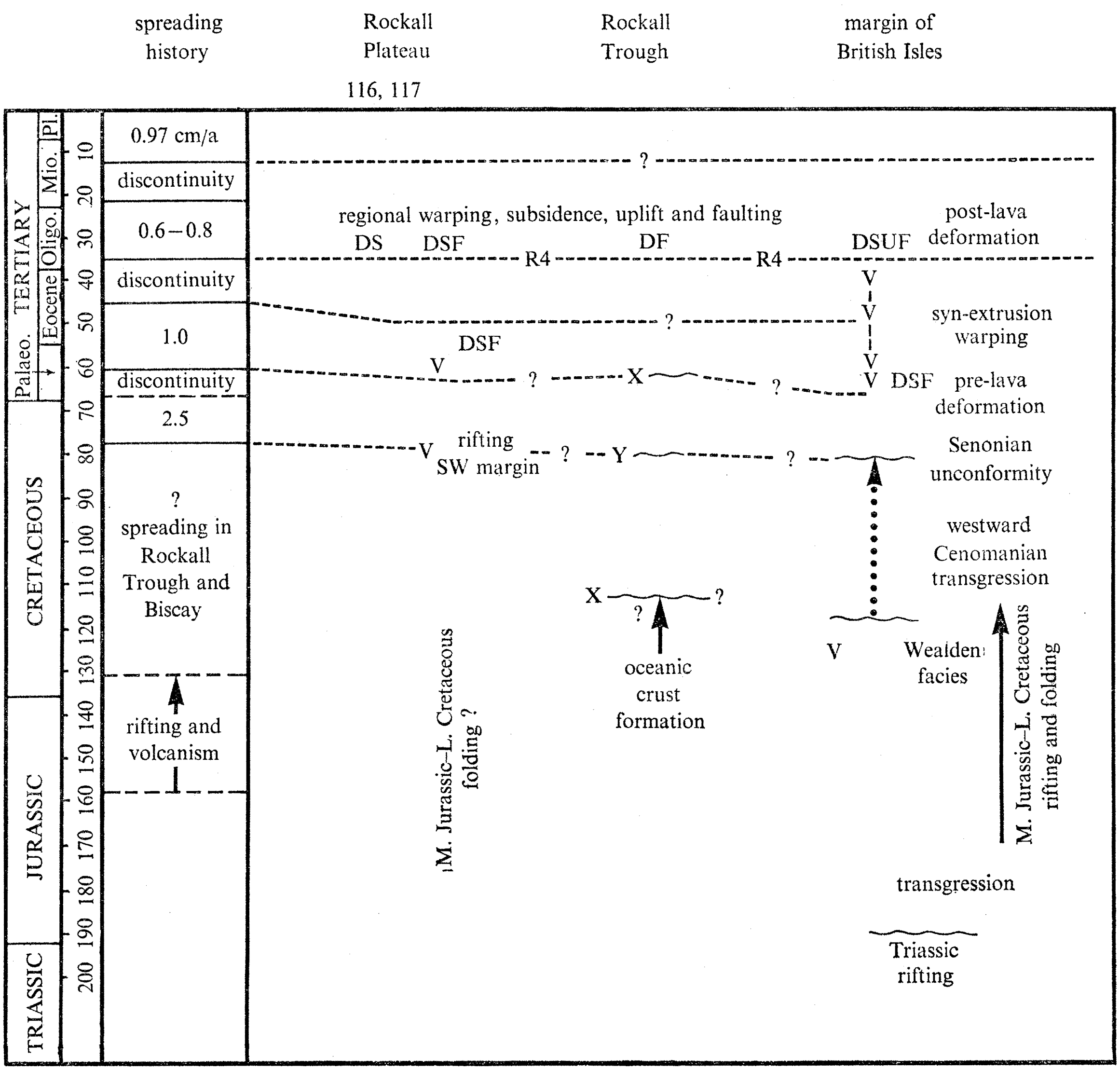
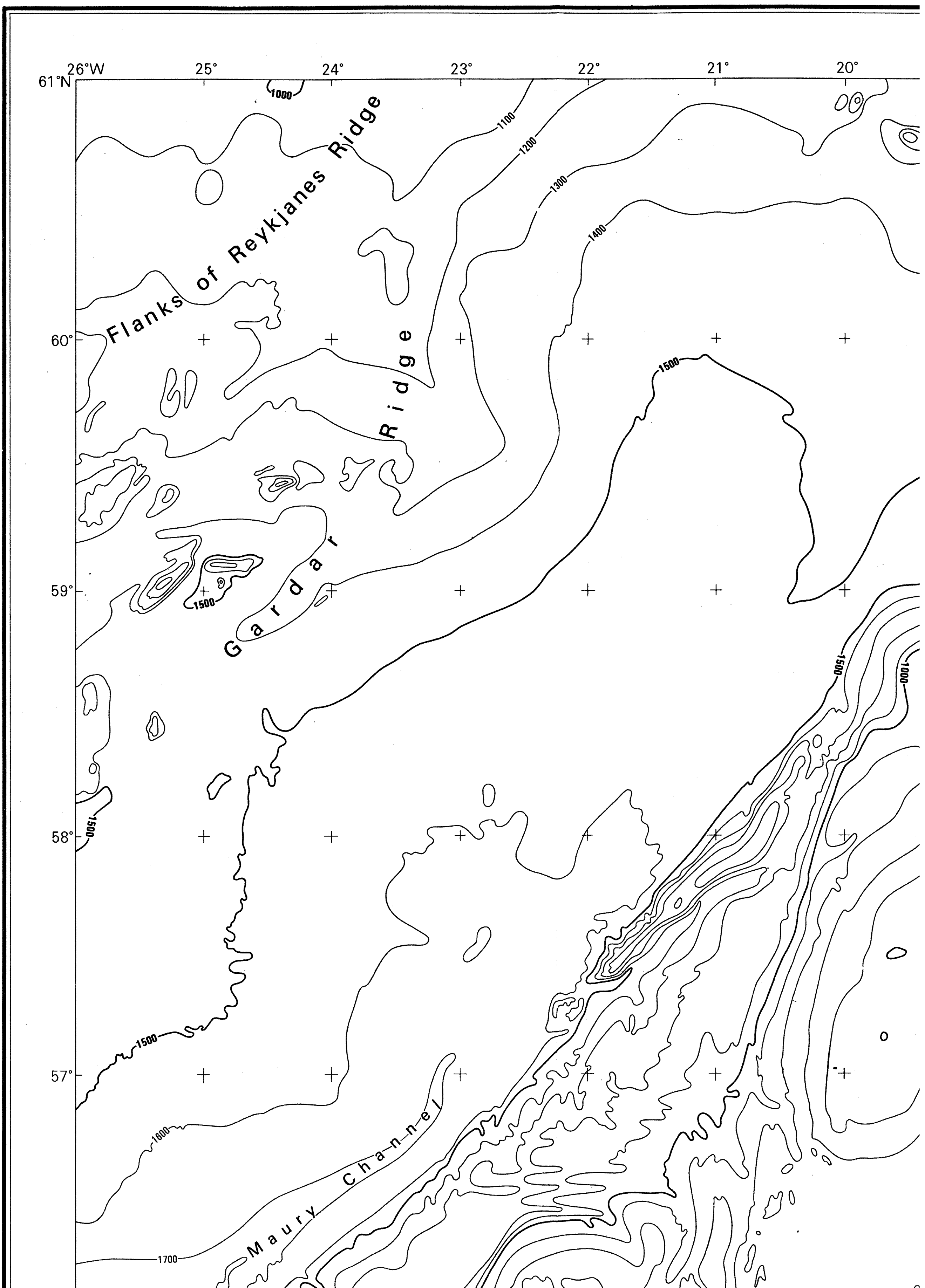


FIGURE 39. Regional geological history. V, Volcanics; DS, downwarping/subsidence; F, faulting; U, uplift.

If the estimated Jurassic–Early Cretaceous age of Rockall Trough is correct, the progressive westward transgression of the higher members of the Cretaceous (Walsh 1966; Hancock 1961) over the following 50 Ma suggests thermal subsidence of the margin consistent with Sleep's model (figure 39). However, in the case of the Hatton–Rockall Basin, subsidence occurred only 10 Ma and 23 Ma after the initiation of spreading and in the Rockall Trough, considerably after the margin had supposedly cooled. The hot creep model is particularly difficult to reconcile with the Rockall Plateau micro-continent because a sink is necessarily implied beneath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Flanks of Reykjanes Ridge
Gardar Ridge
Maury Channel


19°
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Hatton Bank
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George Bligh Bank
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Feni Ridge
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Basin
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700
Hatton Rockall
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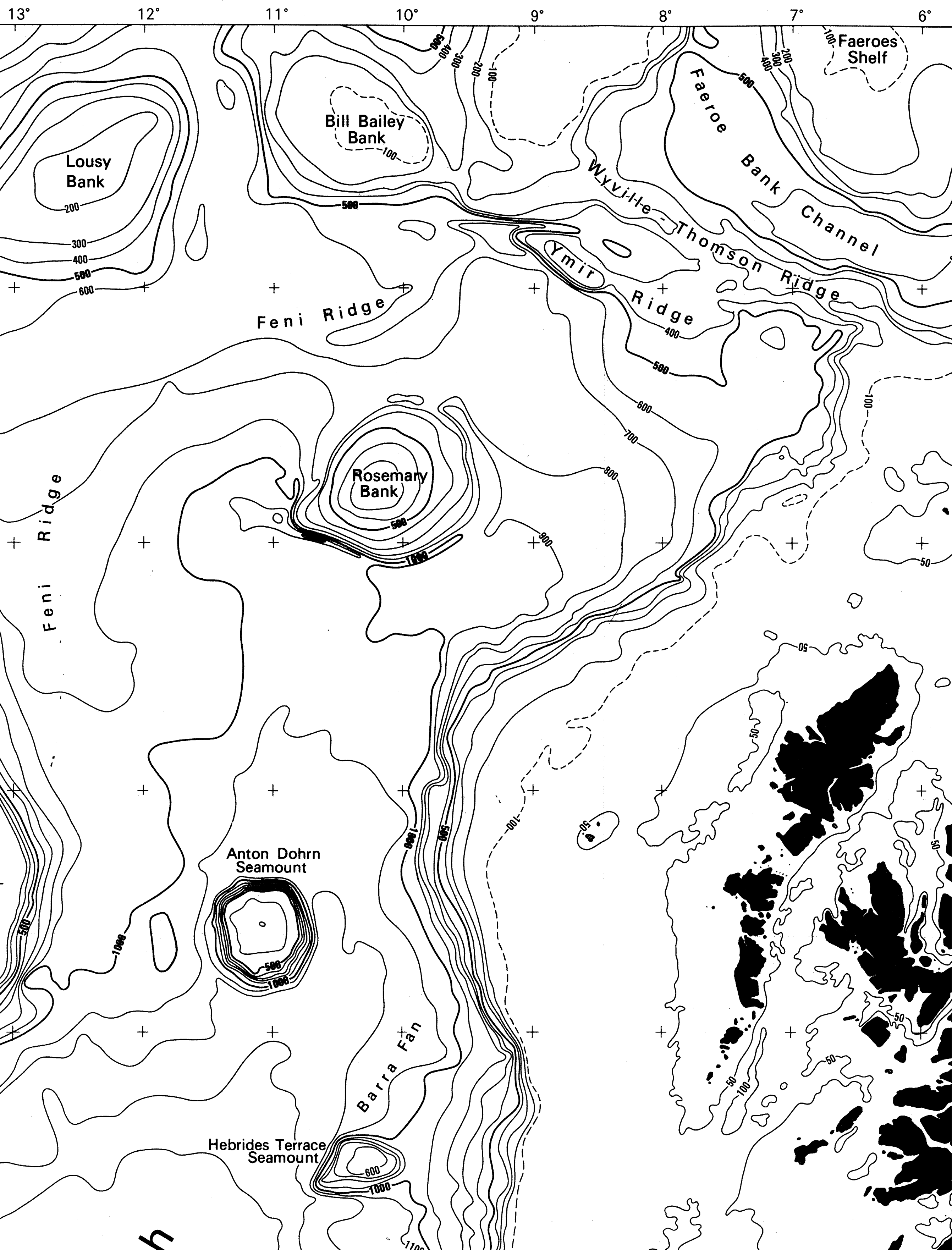
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Faeroes
Shelf
Lousy
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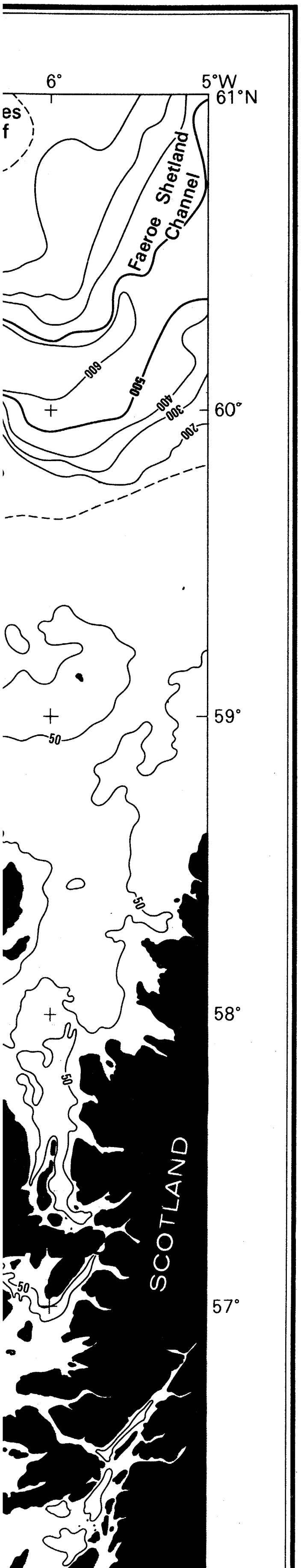
Bill Bailey
Bank
Faeroe Bank Channel
Wyville-Thomson Ridge
Ymir Ridge
Feni Ridge
Feni Ridge
Rosemary
Bank
Anton Dohrn
Seamount
Barra Fan
Hebrides Terrace
Seamount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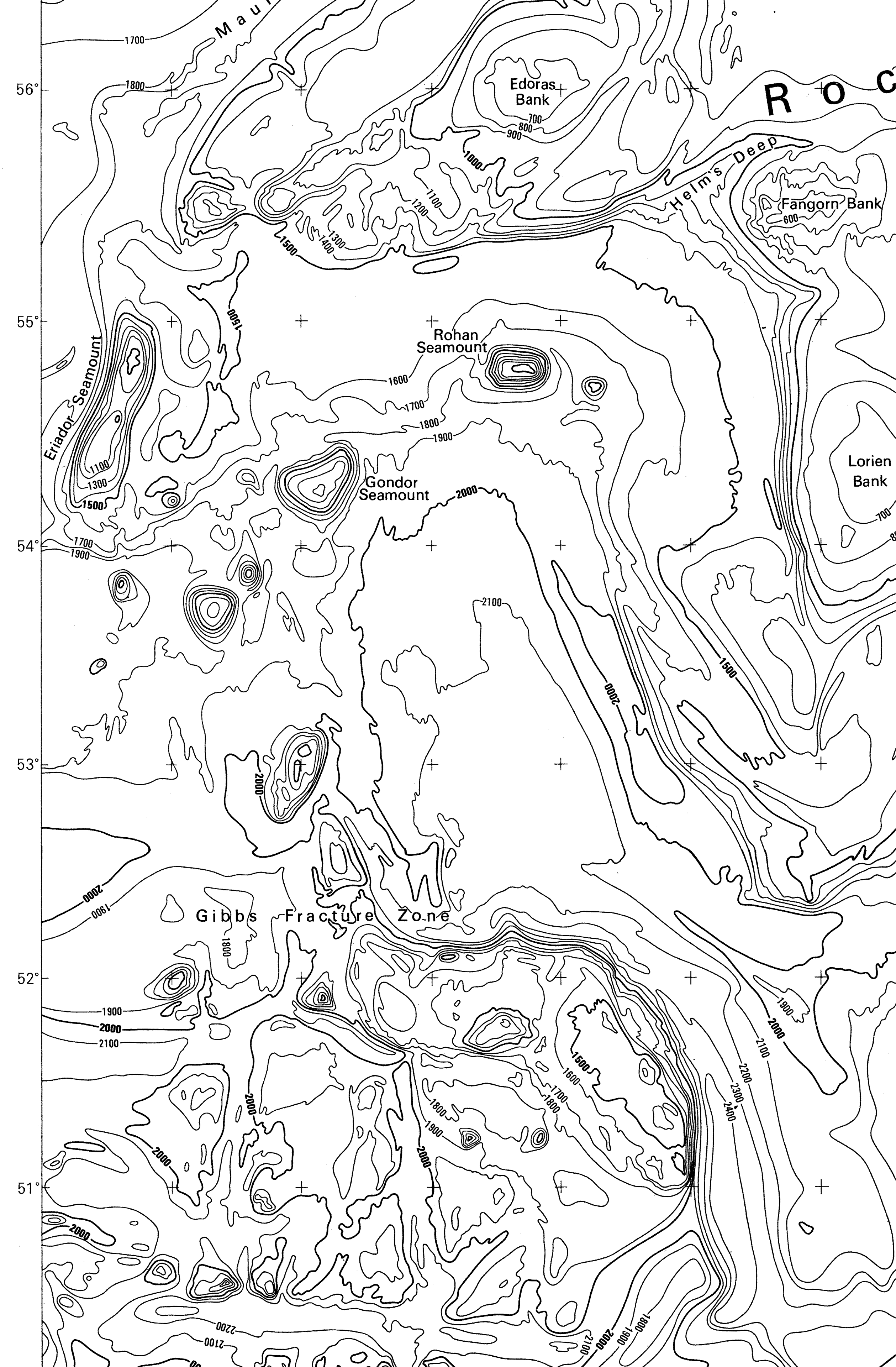
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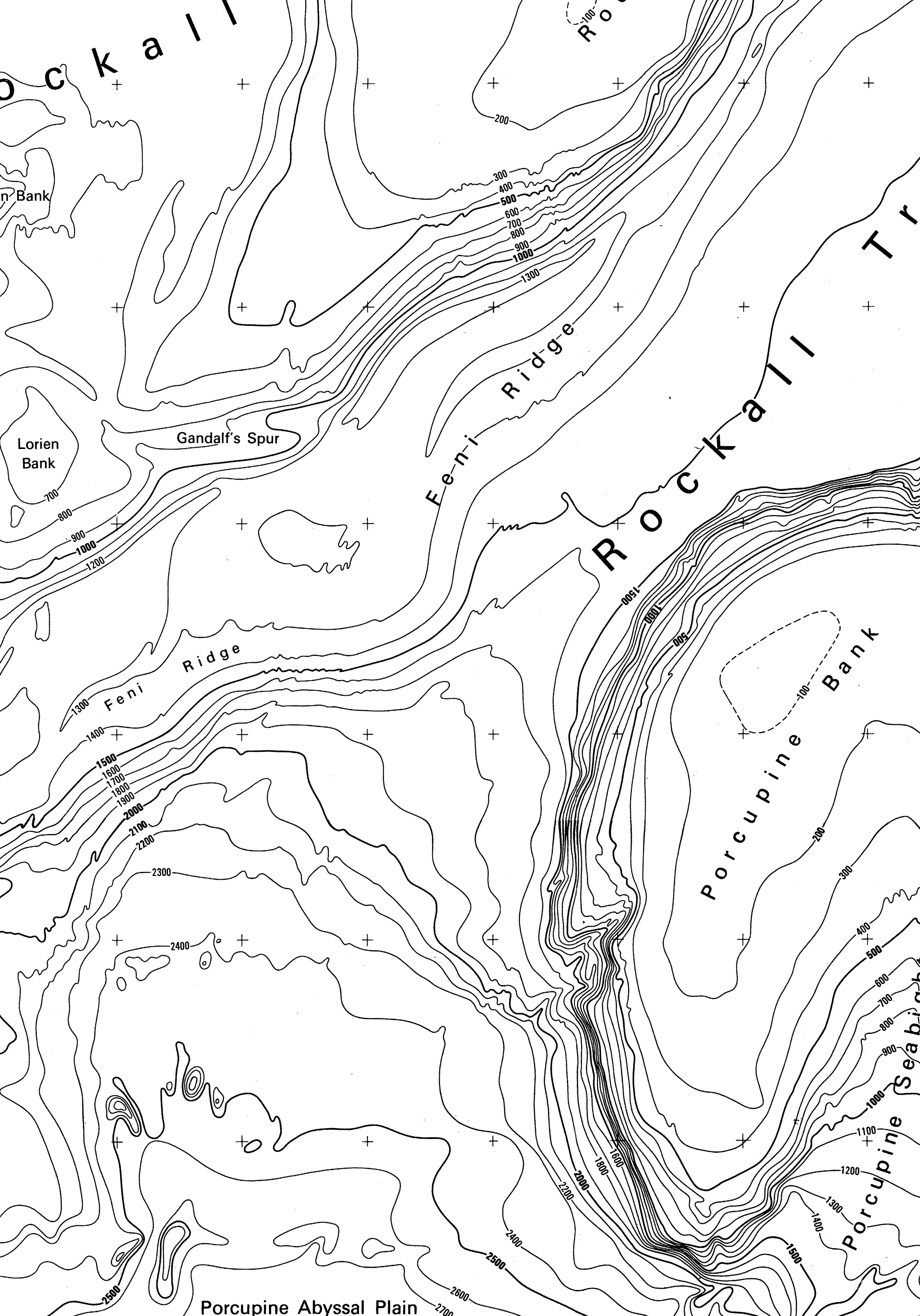
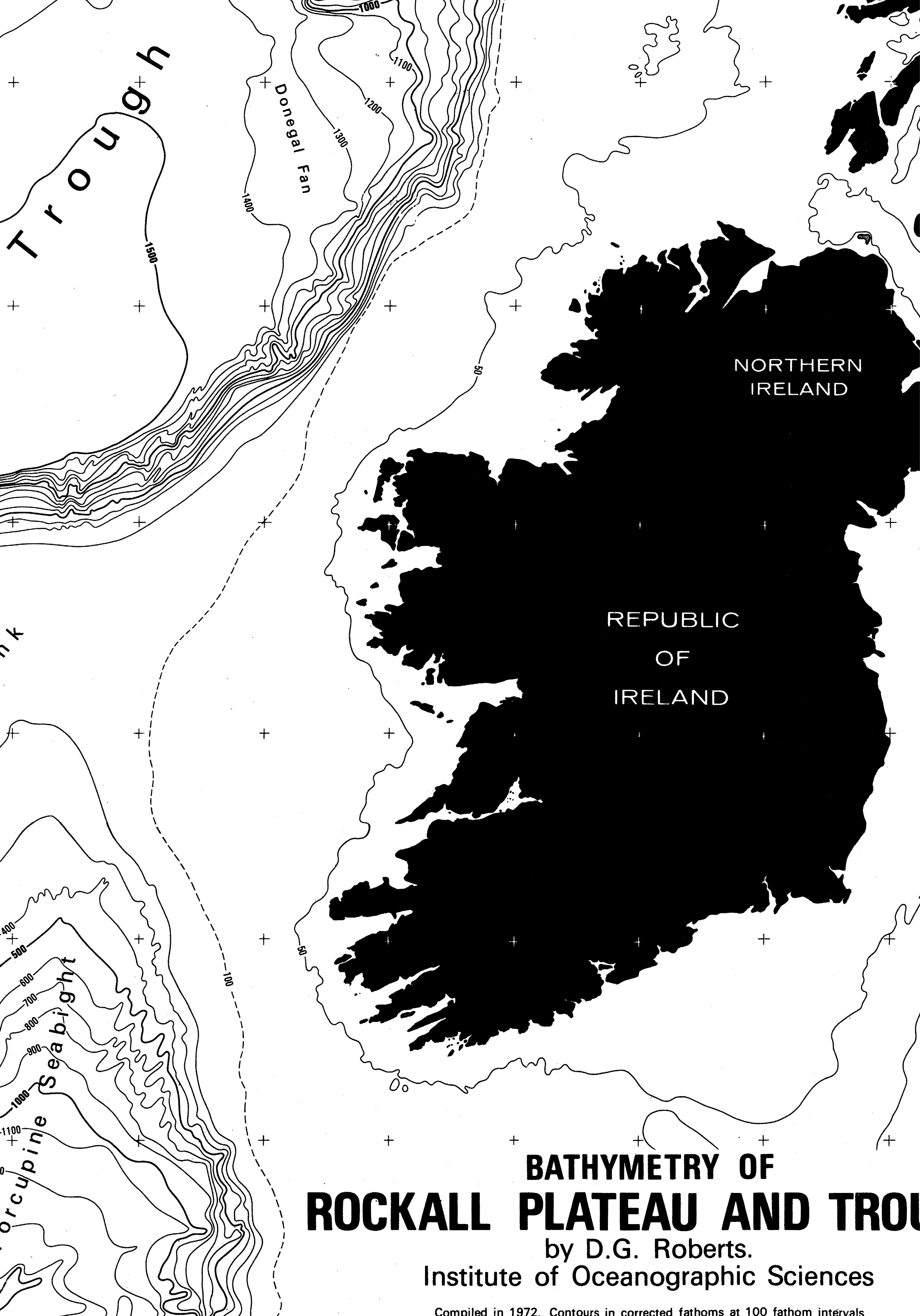
Edoras Bank
Helm's Deep
Fangorn Bank
Rohan Seamount
Eriador Seamount
Gondor Seamount
Lorien Bank
Gibbs Fracture Zone



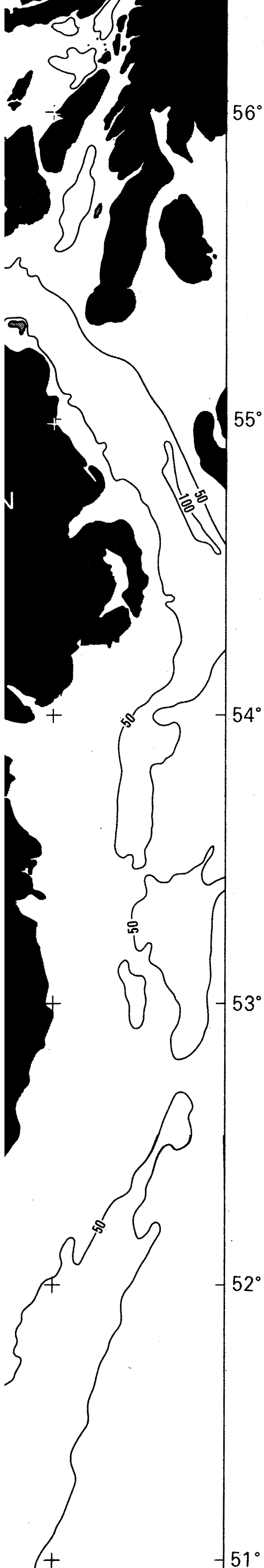
Rockall
Rockall Trough
Lorien Bank
Gandalf's Spur
Feni Ridge
Feni Ridge
Porcupine Bank
Porcupine Seabight
Porcupine Abyssal Plain



BATHYMETRY OF
ROCKALL PLATEAU AND TROU

by D.G. Roberts.
Institute of Oceanographic Sciences

Compiled in 1972. Contours in corrected fathoms at 100 fathom intervals



ROUGH

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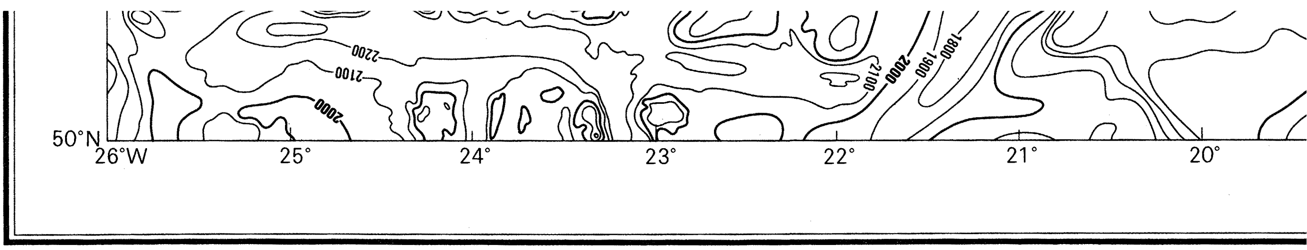
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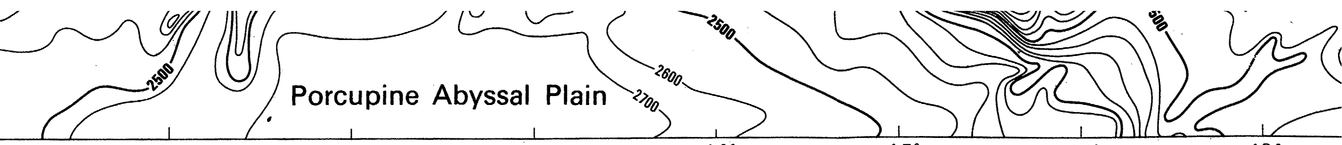

Porcupine Abyssal Plain
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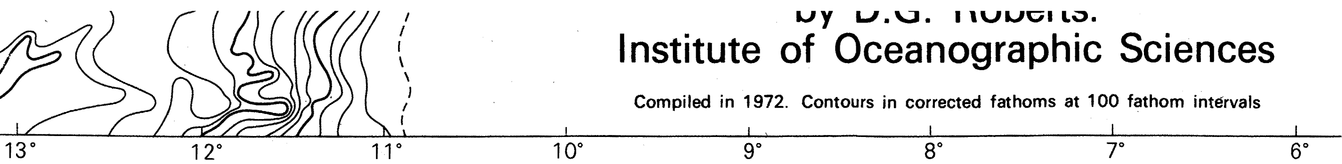
FIGURE 40.


Institute of Oceanographic Sciences
Compiled in 1972. Contours in corrected fathoms at 100 fathom intervals
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Rockall Trough to accomodate the outflow which must also have occurred in widely different directions.

Differential loading of the lithosphere has considerable appeal since it adroitly explains the contrasting deformation between the oceanic and continental parts of the plate. However, the lithosphere must be regionally stressed to produce contemporaneous uplift and subsidence. In the case of a cooled margin, the stress due to differential loading may be largely in equilibrium and further faulting or warping can only occur by superimposing another stress on the lithosphere or by increasing the differential loading, e.g. by increased sedimentation. The coincidence of the subsidence on the still cooling Rockall Plateau with spreading rate changes suggest the former may be the case. Localization of basin subsidence in areas of crustal thinning or on lithospheric fractures implies changes at the base of the lithosphere may be responsible although the precise mechanism is unknown.

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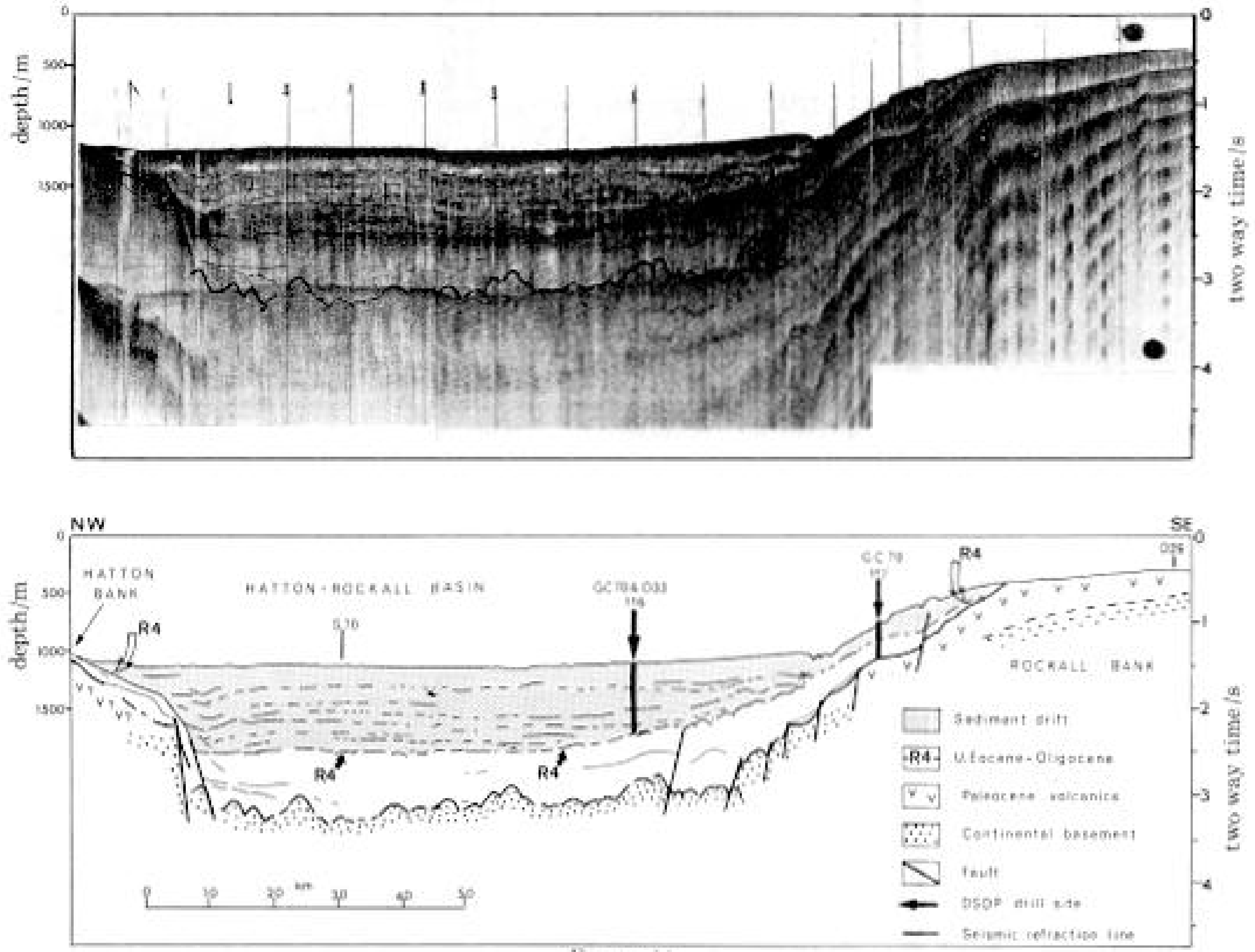


FIGURE 14. Seismic reflexion profile (14) across the Hatton-Rockall Basin through JOIDES sites 116 and 117. Profile is located in figure 12.

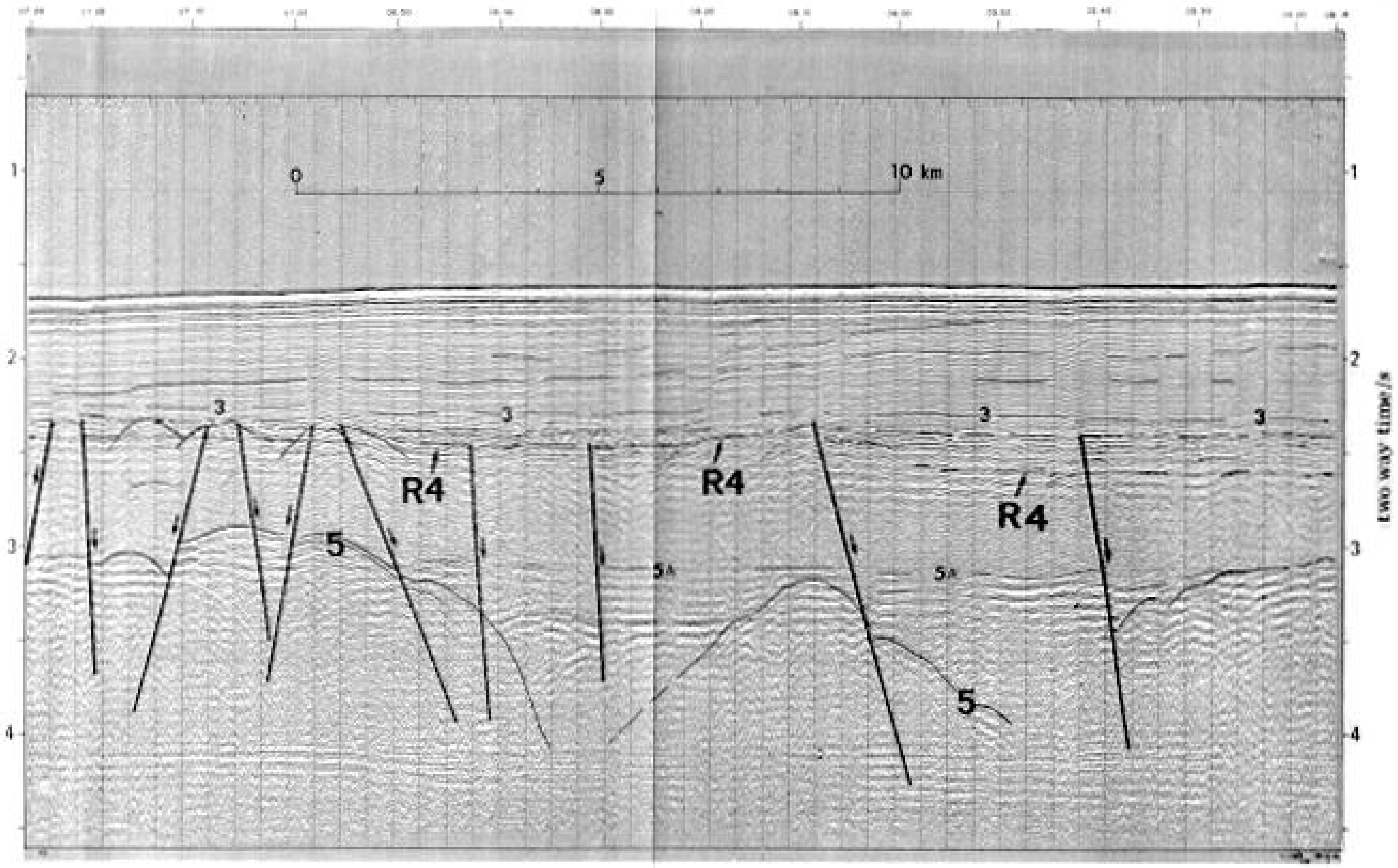


FIGURE 16. Deconvolved seismic reflexion profile in the Hatton-Rockall Basin. Section is located in figure 22.

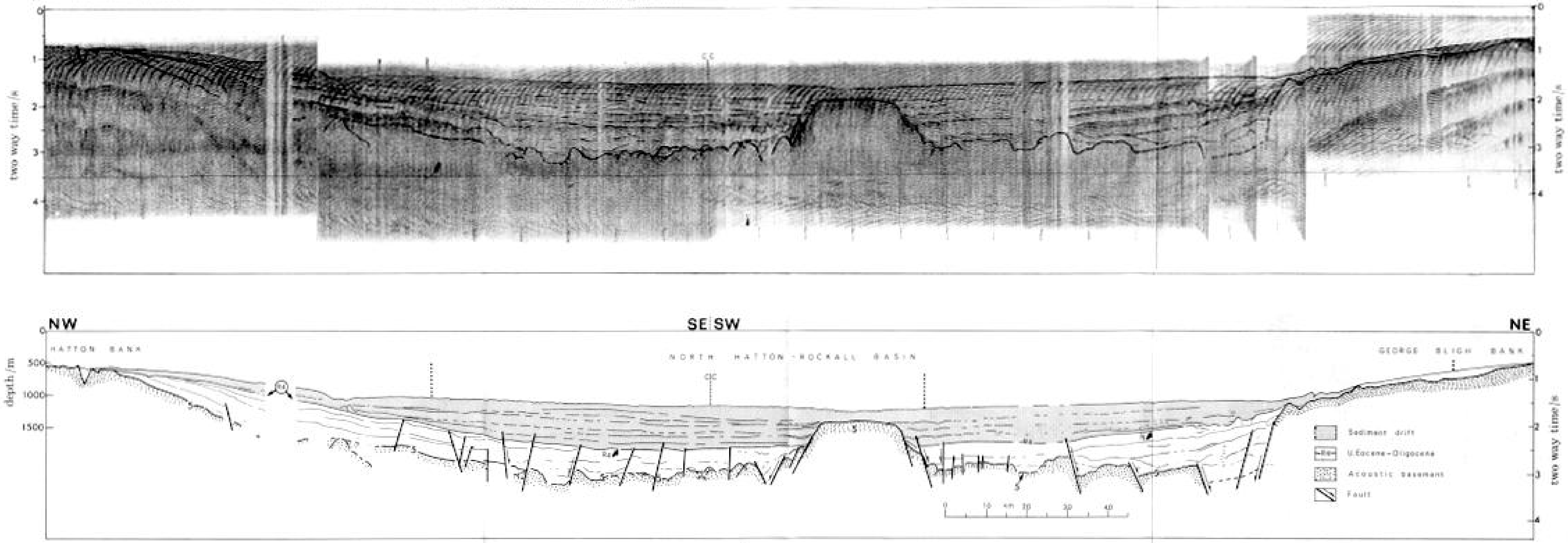


FIGURE 19. Seismic reflexion profile (19) across the northern Hatton-Rockall Basin. Profile is located in figure 12.

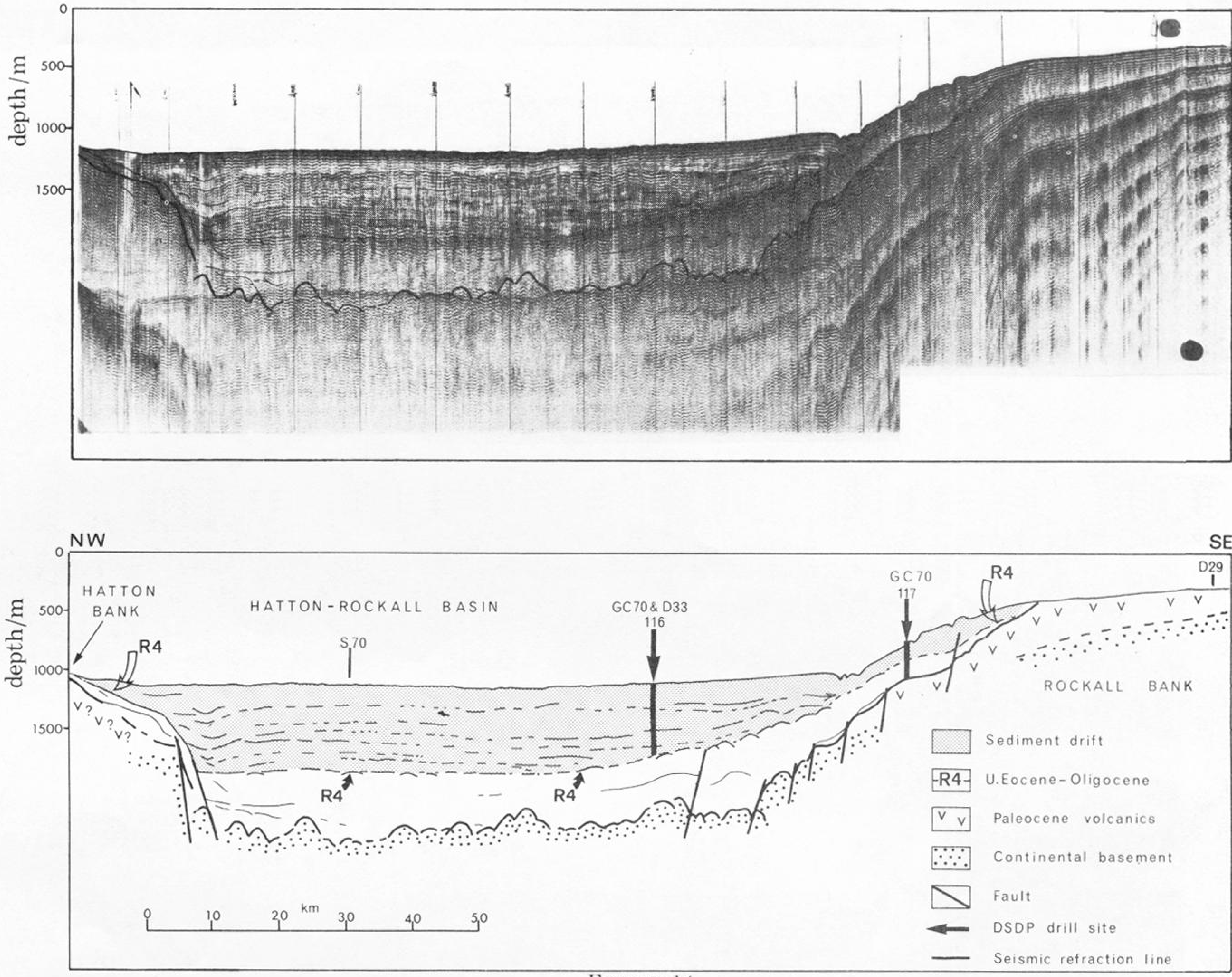
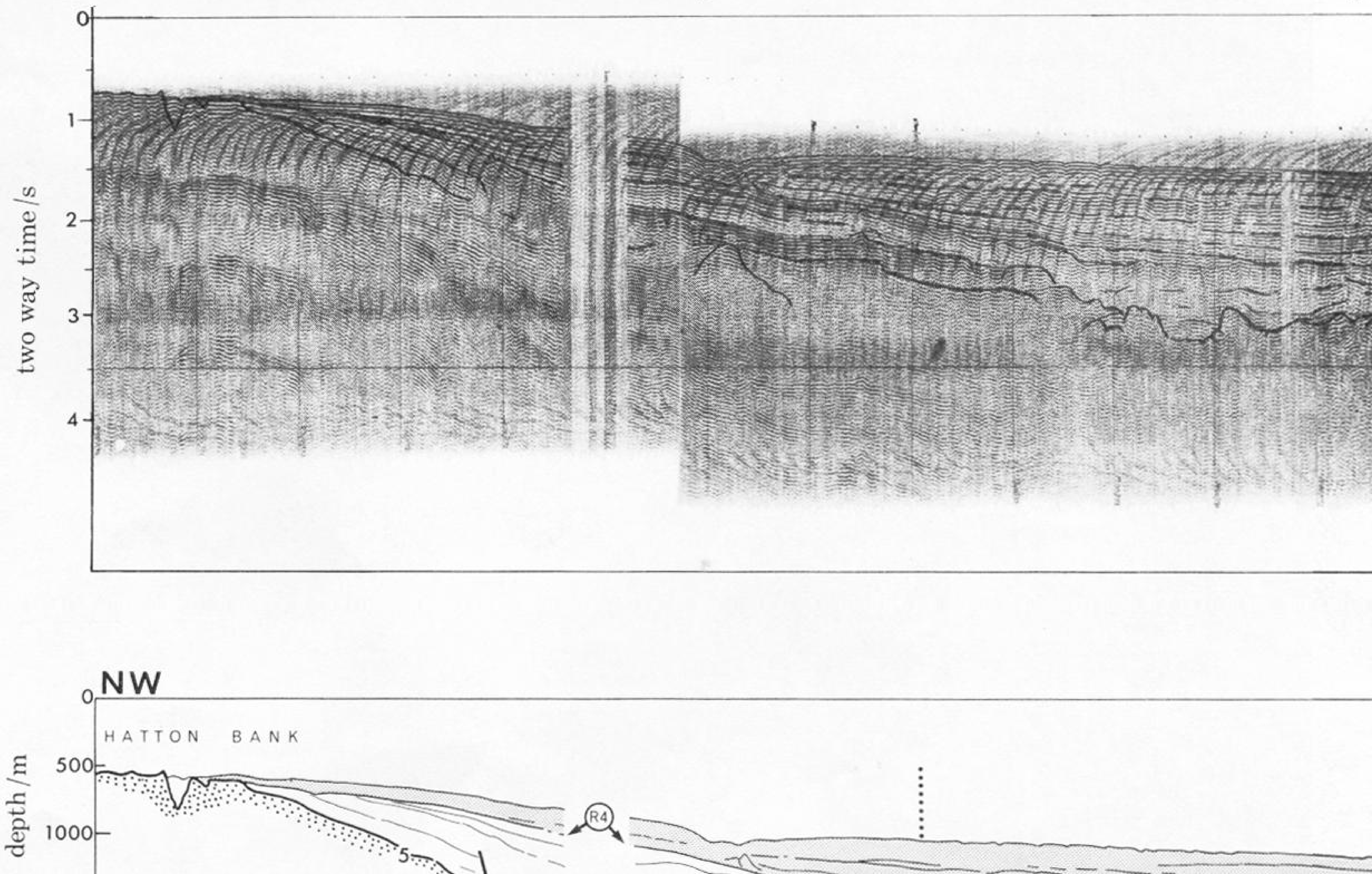
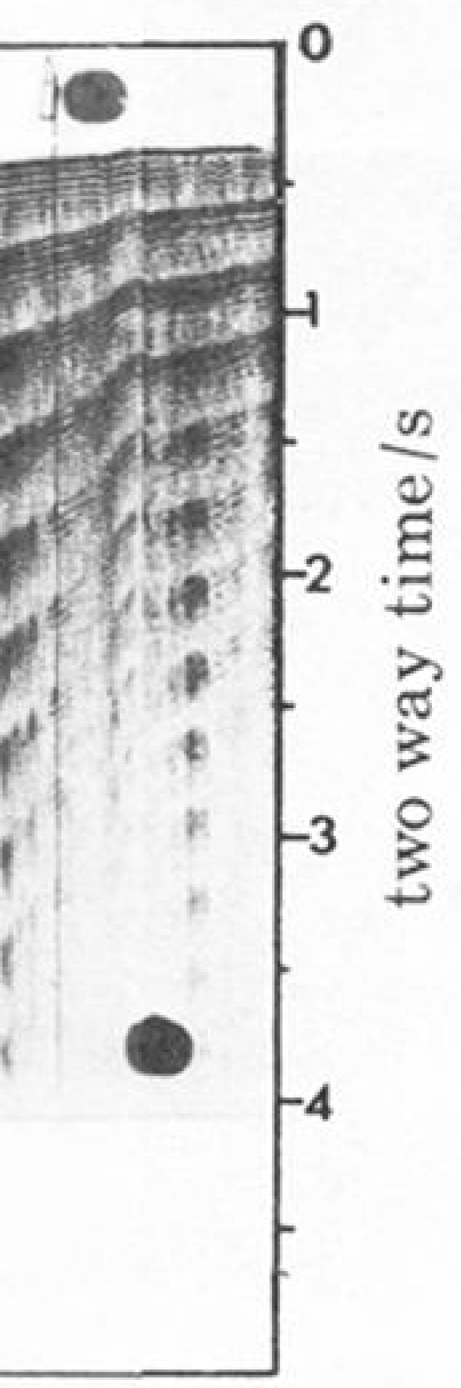
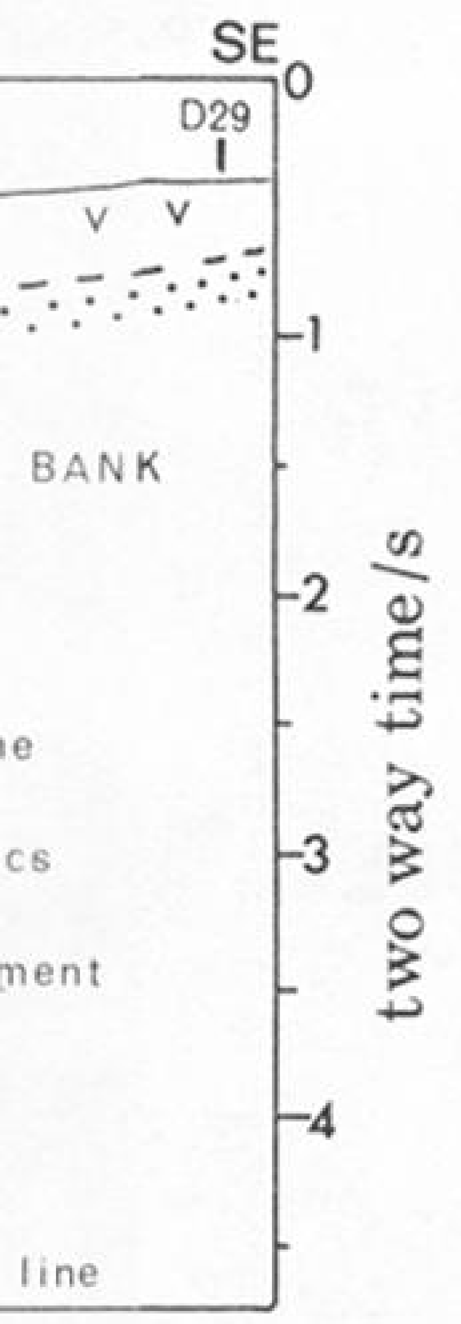


FIGURE 14.
Seismic reflexion profile (14) across the Hatton-Rockall Basin through JOIDES sites 116 and 117. Profile is located in figu







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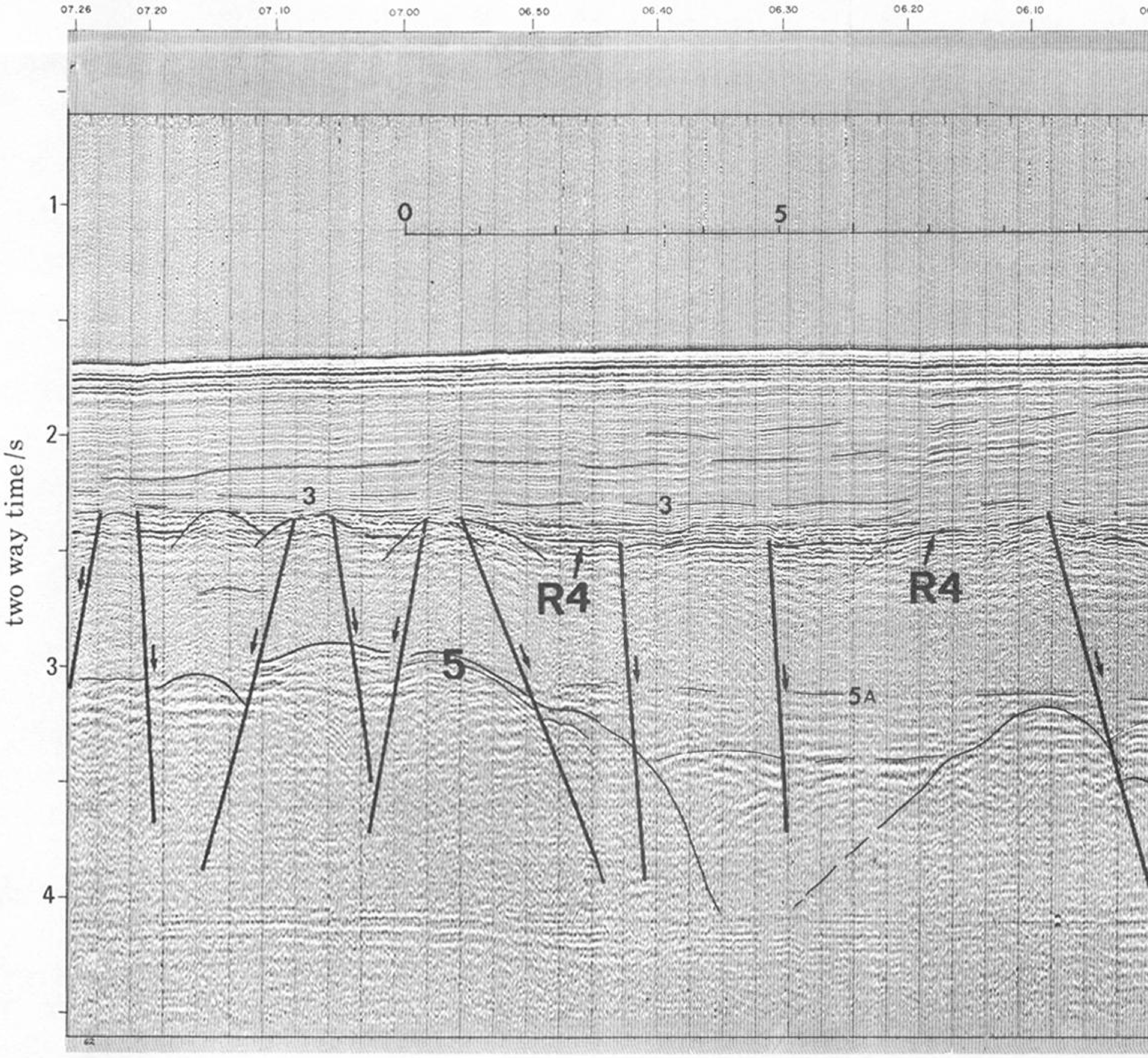

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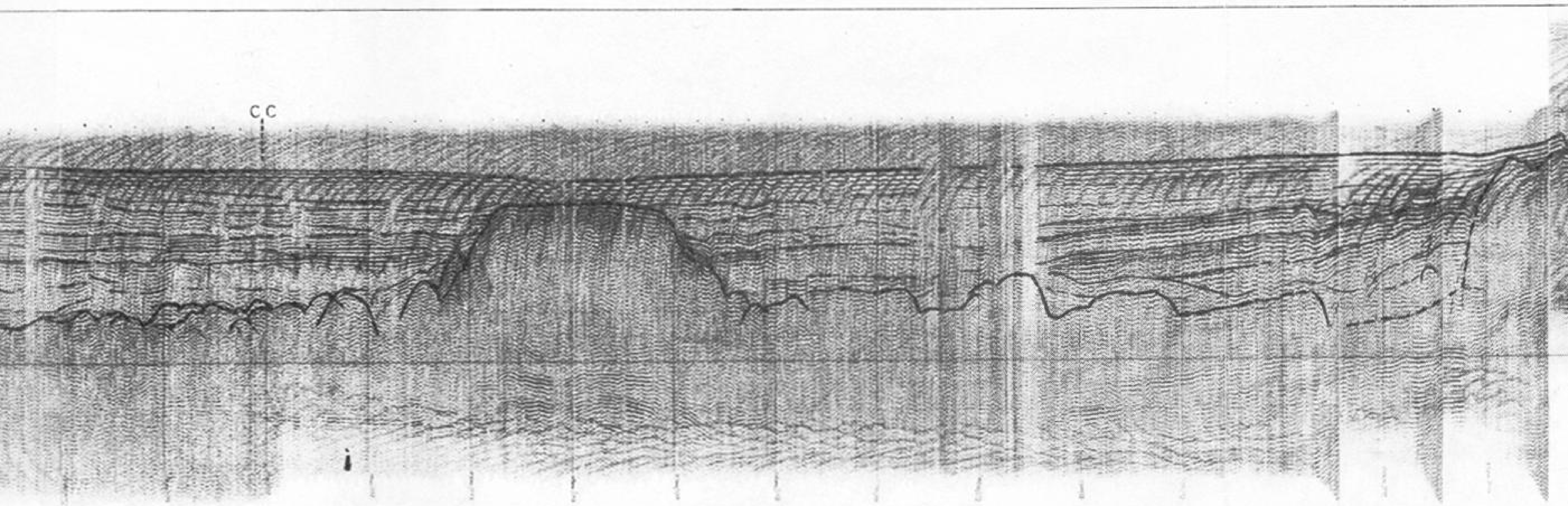
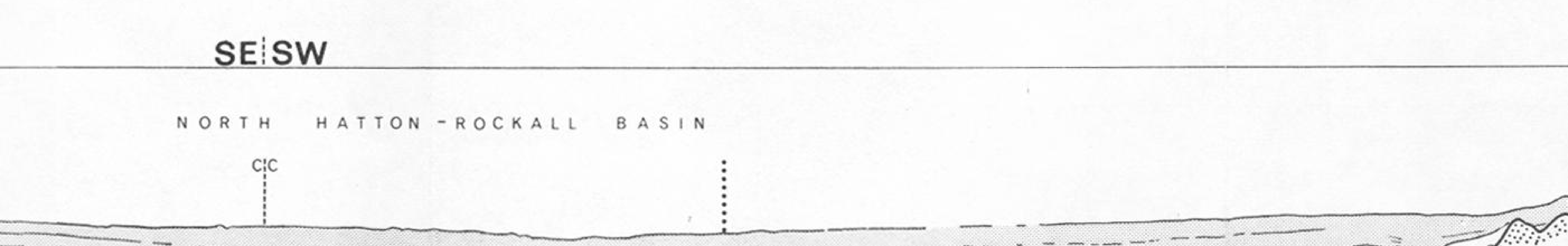
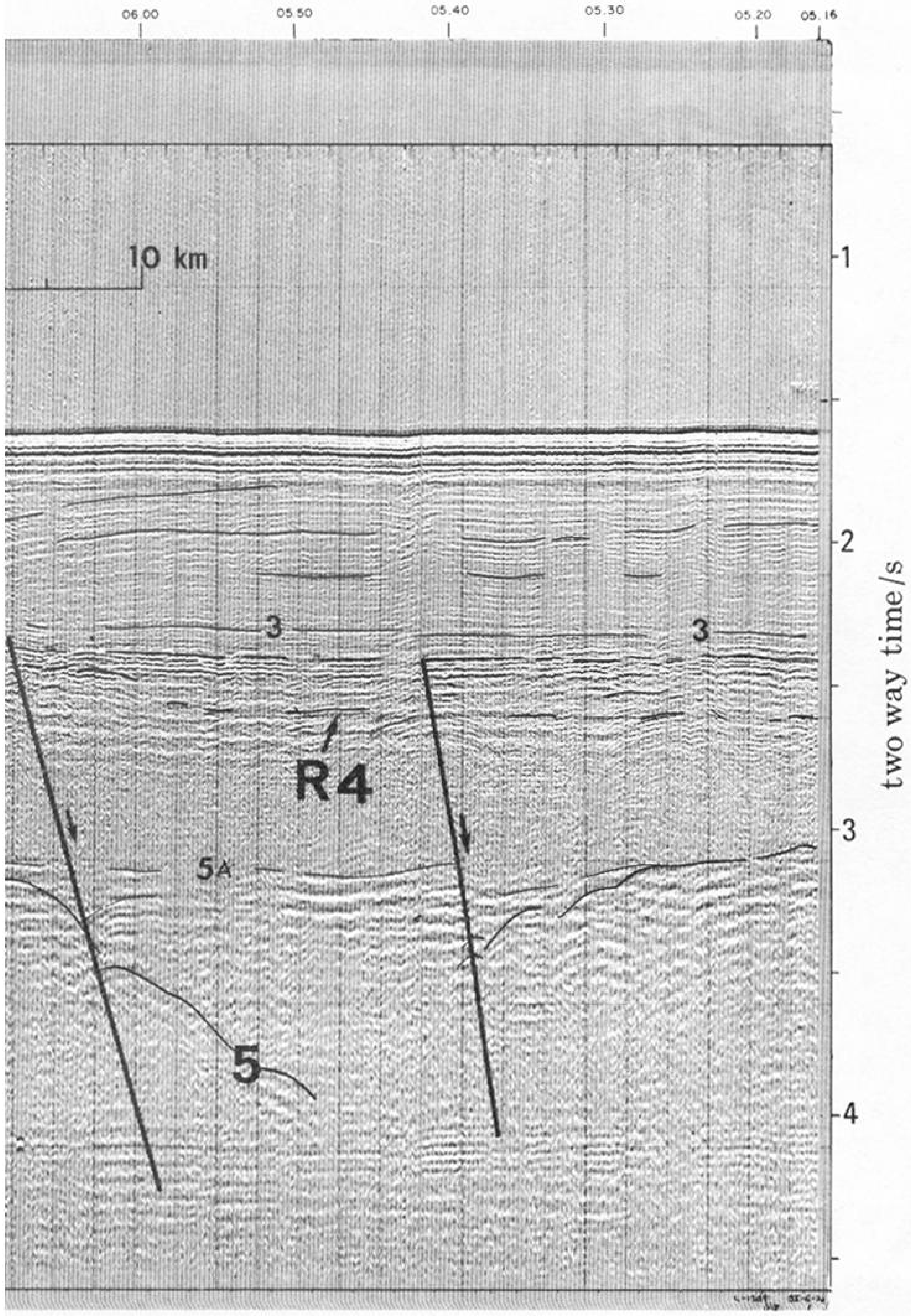
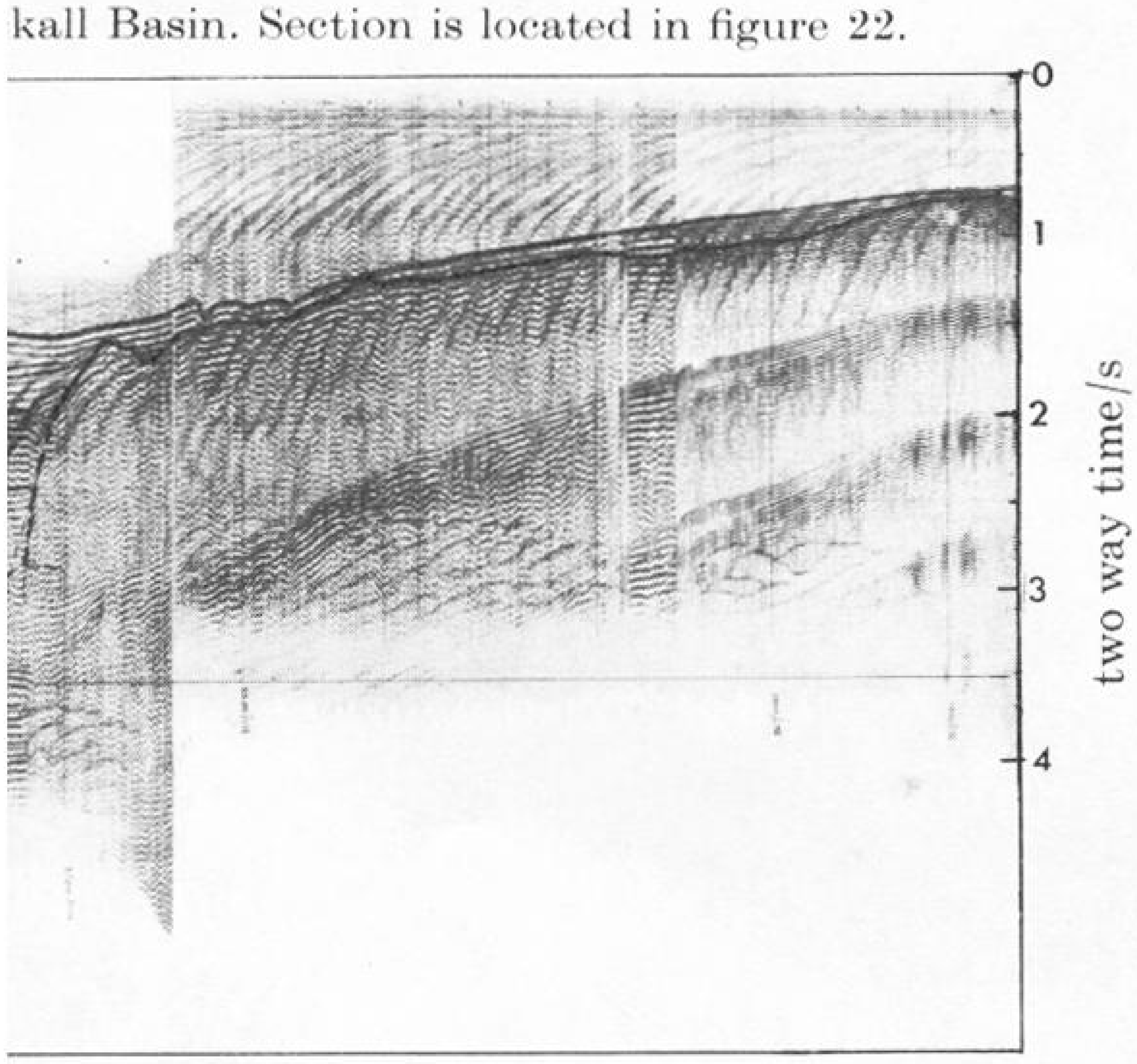
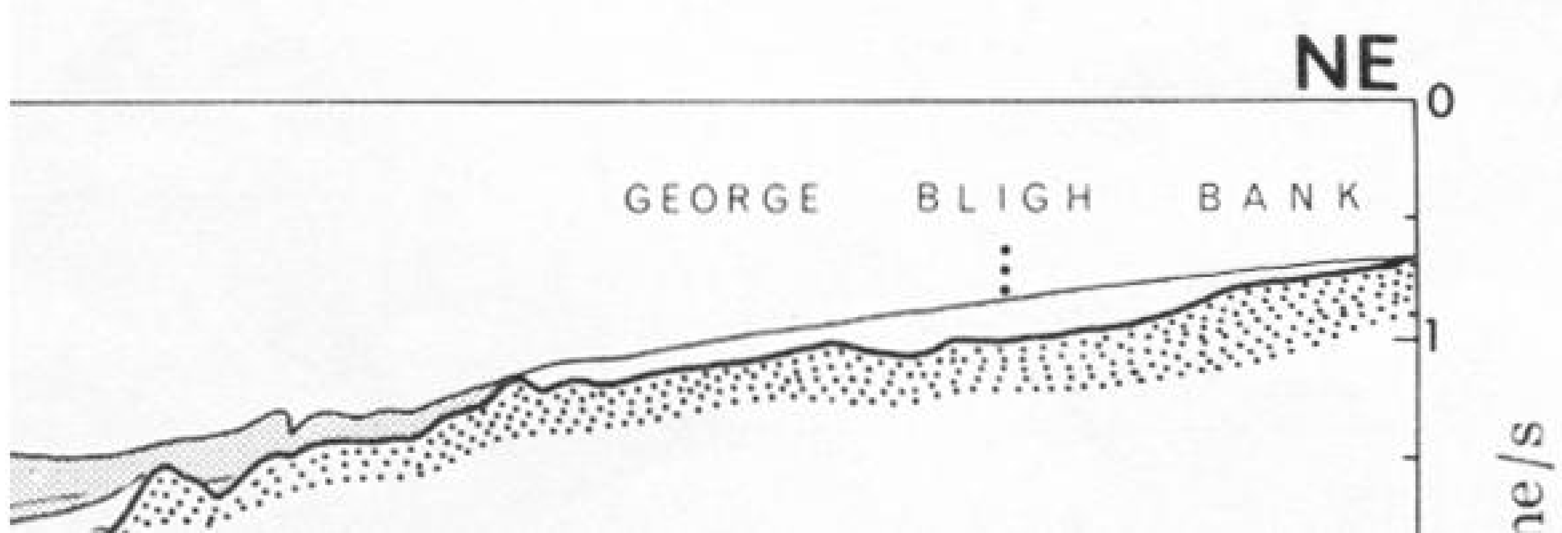






kall Basin. Section is located in figure 22.





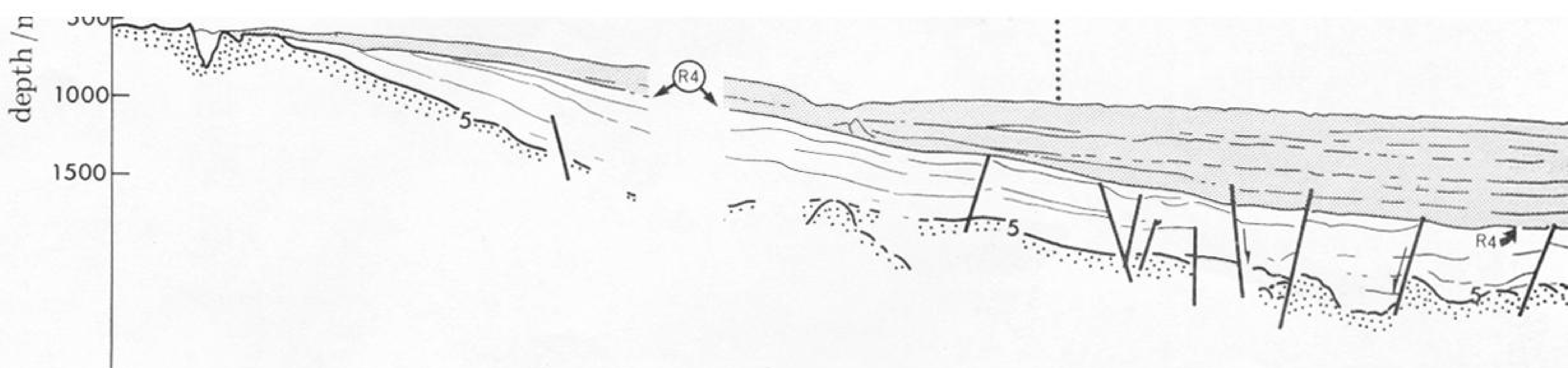
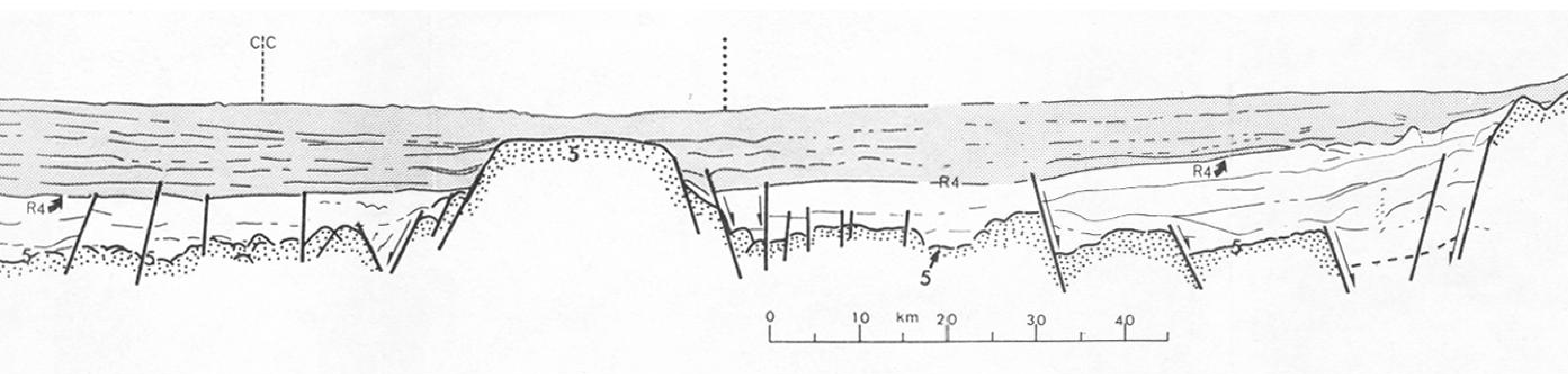
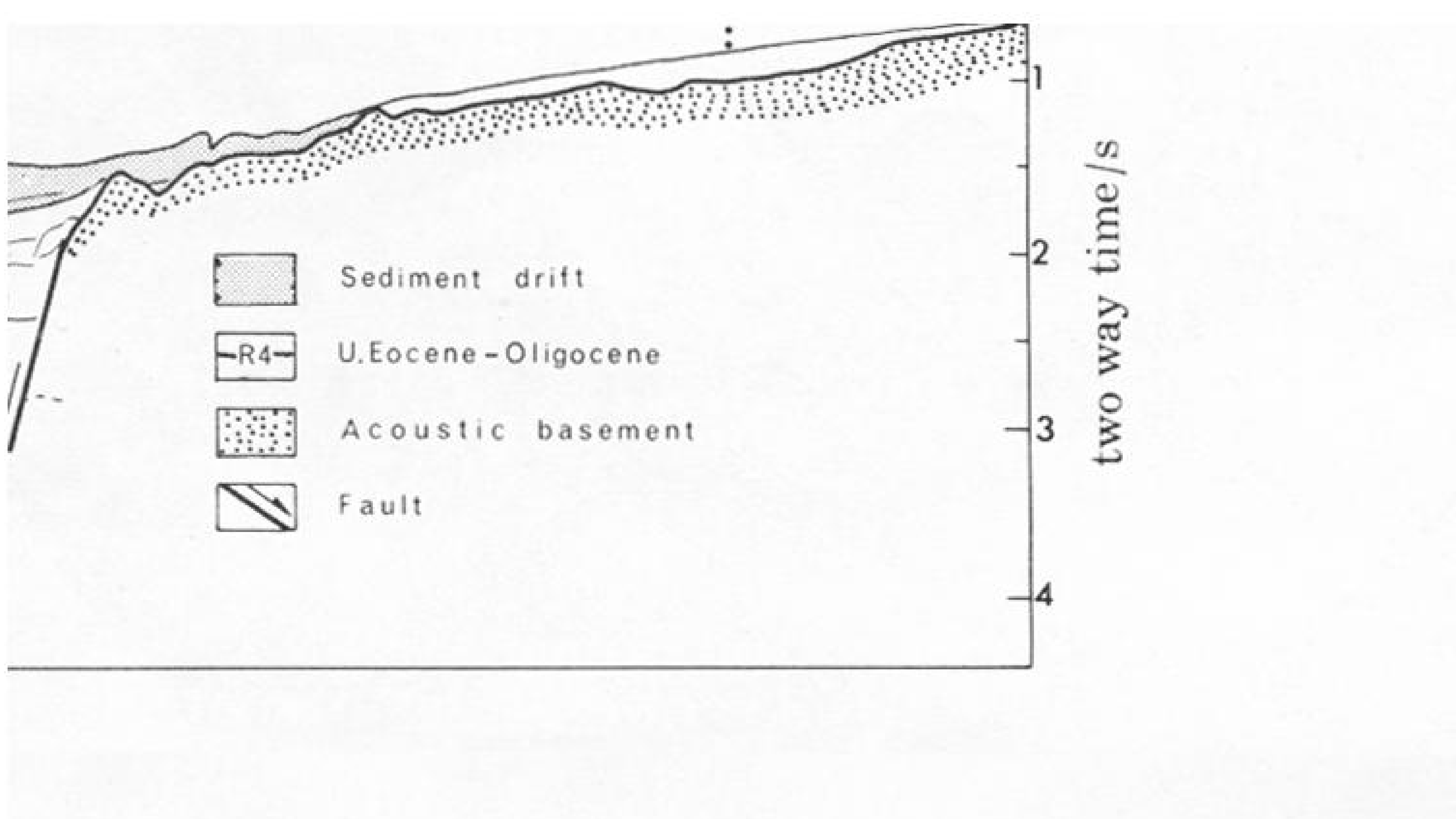


FIGURE 19. Seismic reflexion pr



lexion profile (19) across the northern Hatton-Rockall Basin. Profile is located in figure 12.



Sediment drift
R4
U. Eocene - Oligocene
Acoustic basement
Fault
1
2
3
4
two way time/s

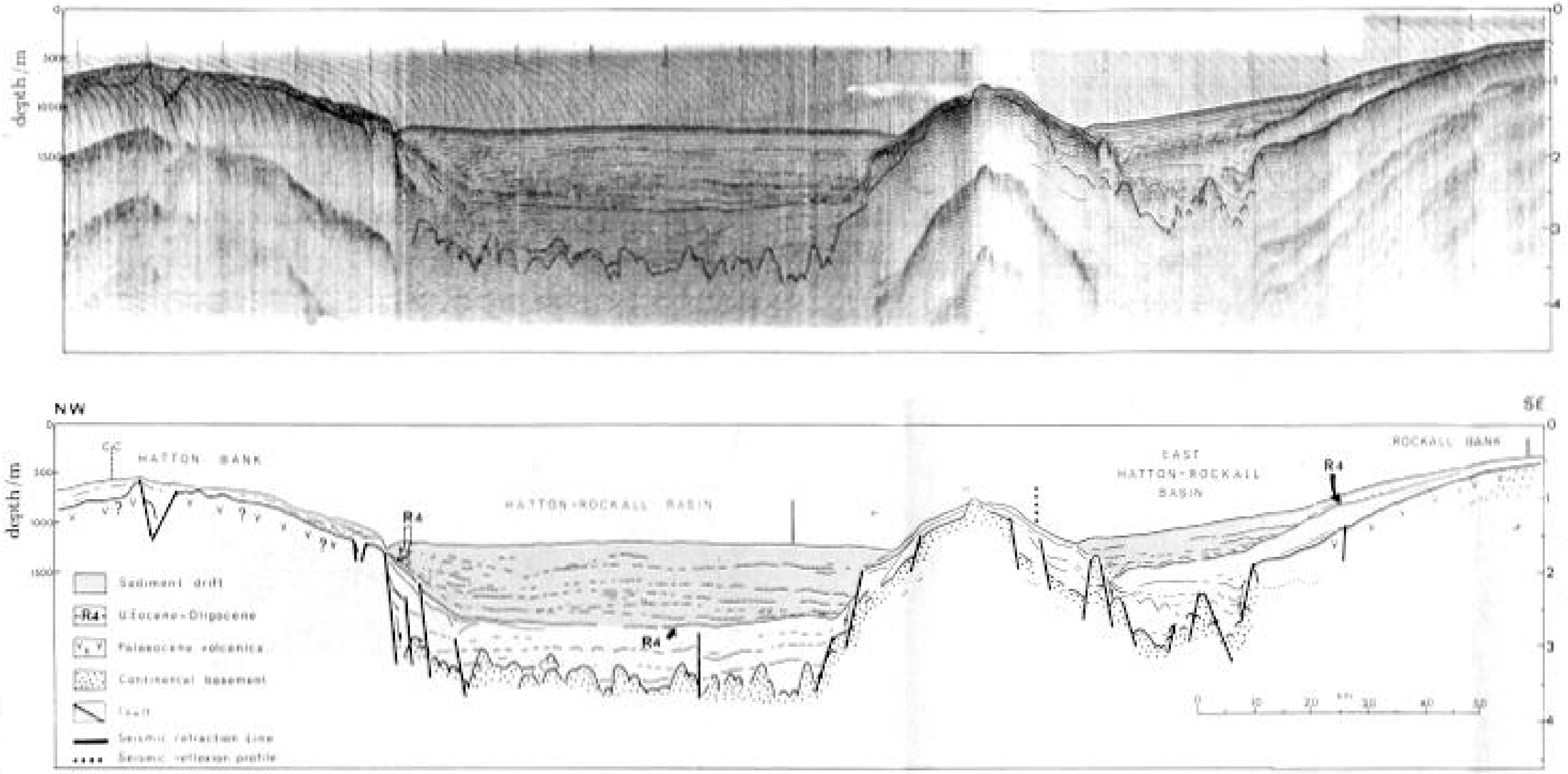


FIGURE 20. Seismic reflexion profile (20) across the central Hatton-Rockall Basin. Profile is located in figure 12.

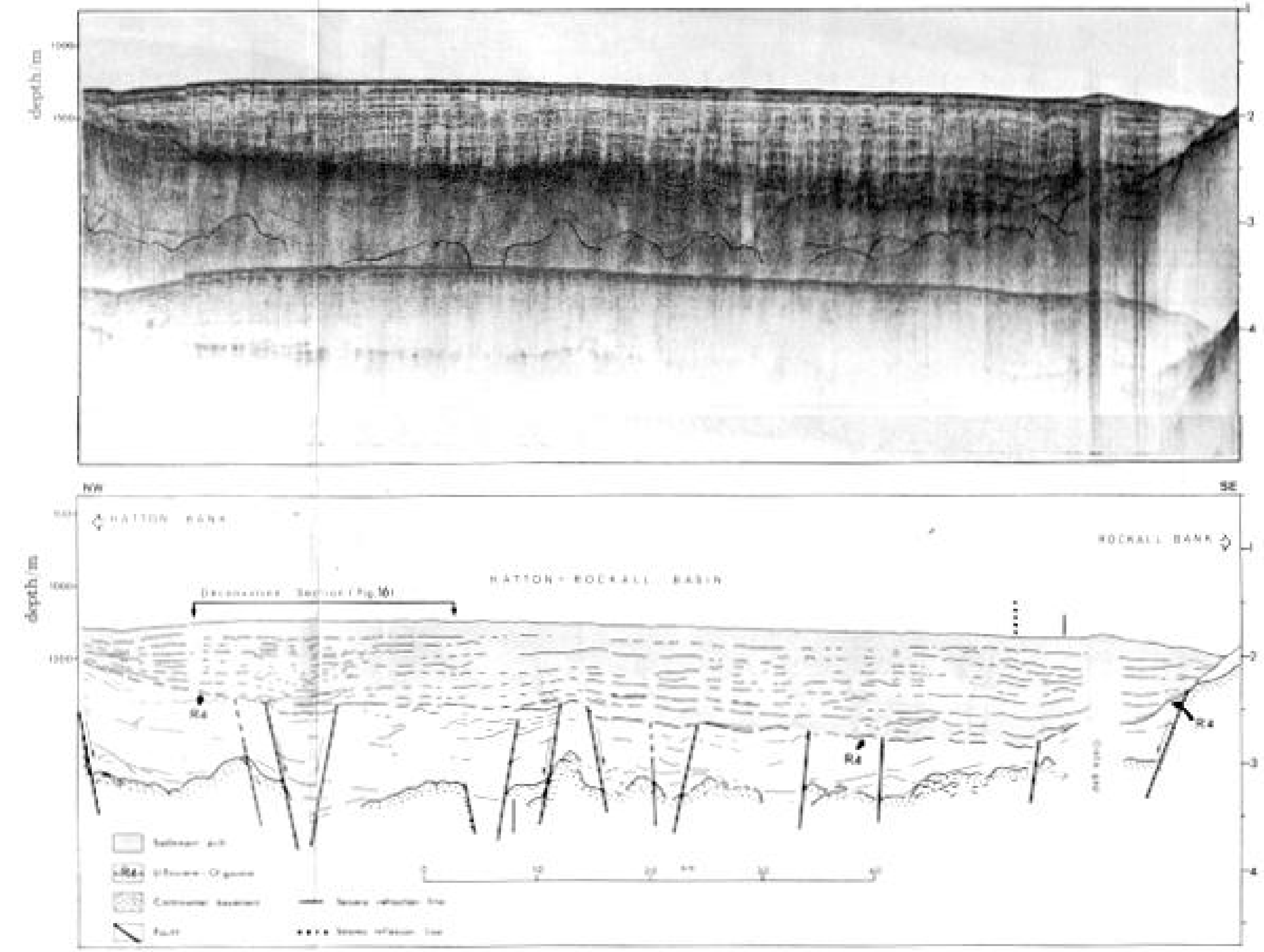


FIGURE 22. Seismic reflexion profile (22) across the central Hatton-Rockall Basin. Profile is located in figure 12.

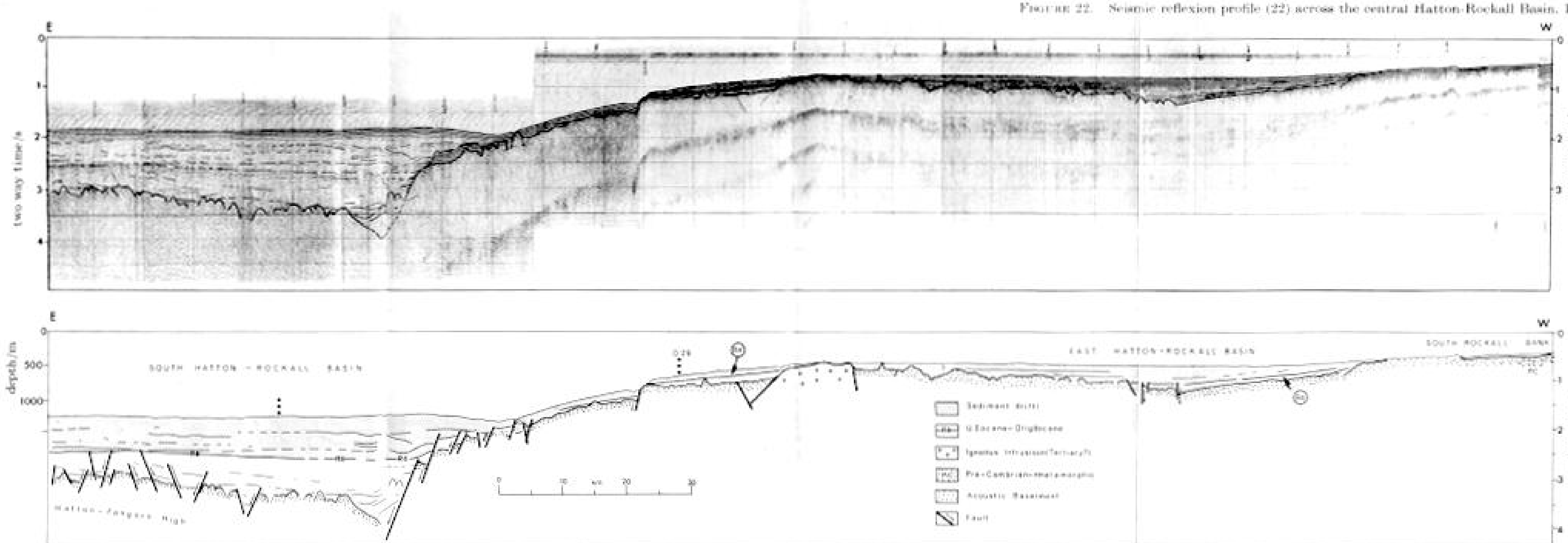


FIGURE 23. Seismic reflexion profile (23) between the south Hatton-Rockall Basin, Faeroes High and Rocka

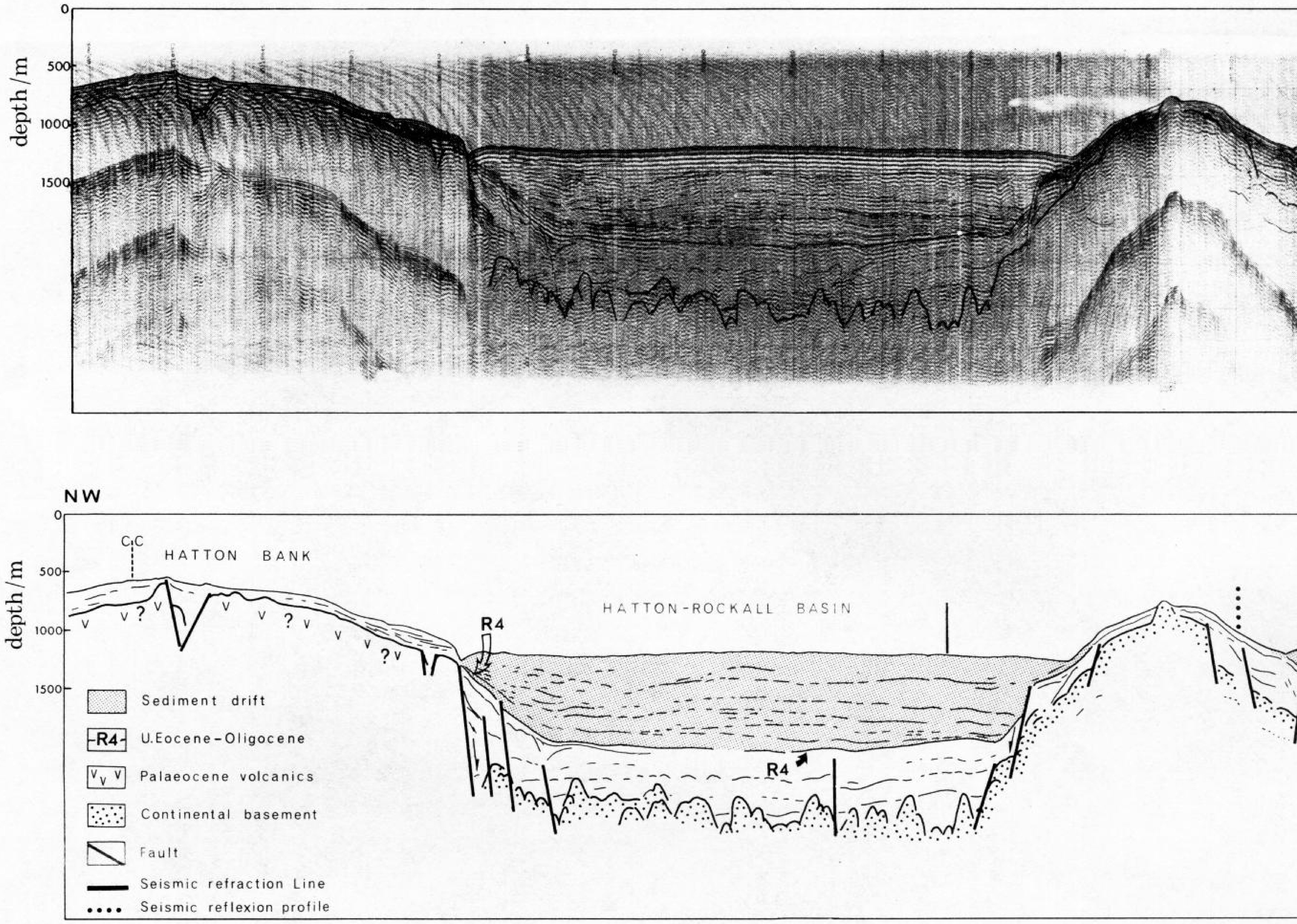
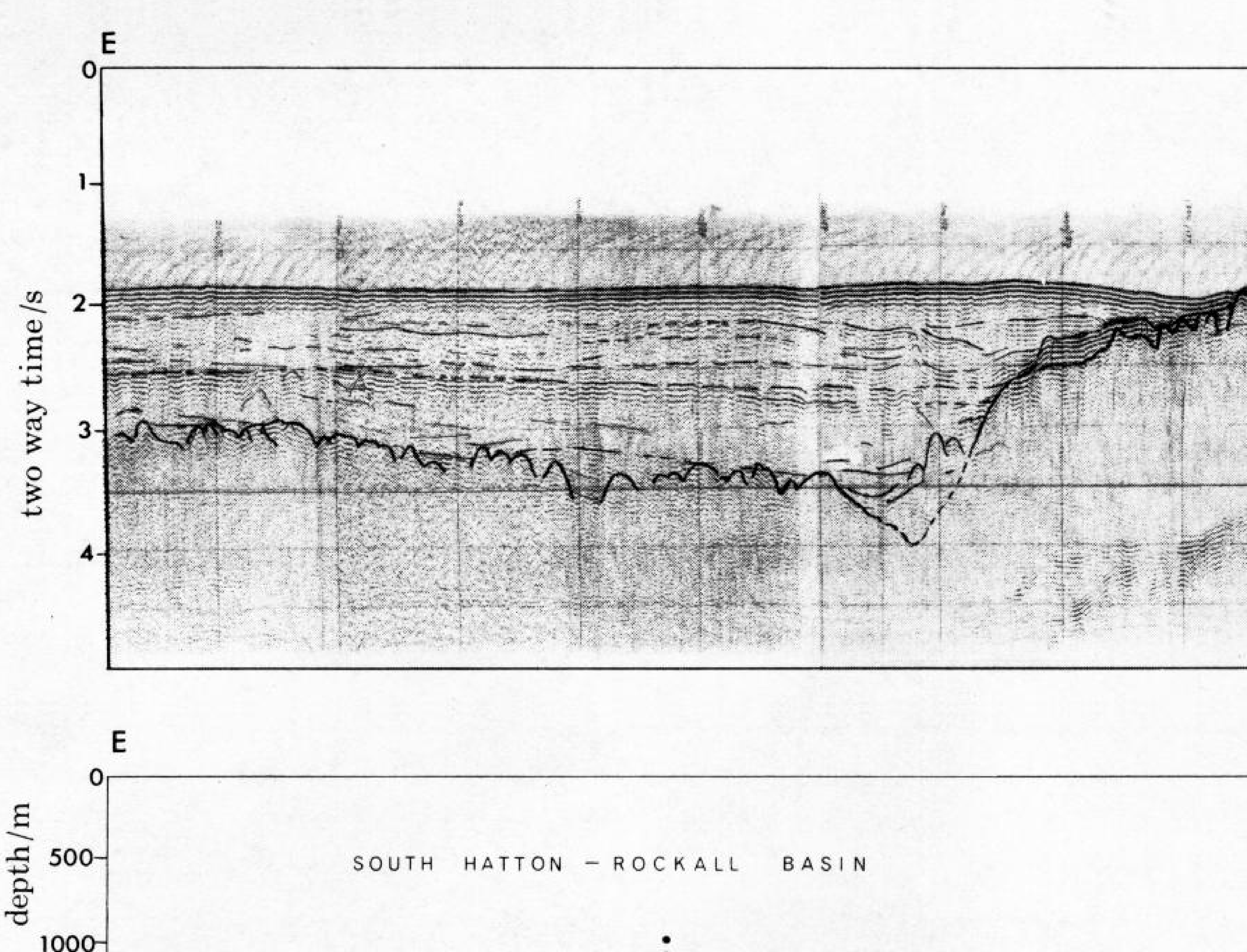
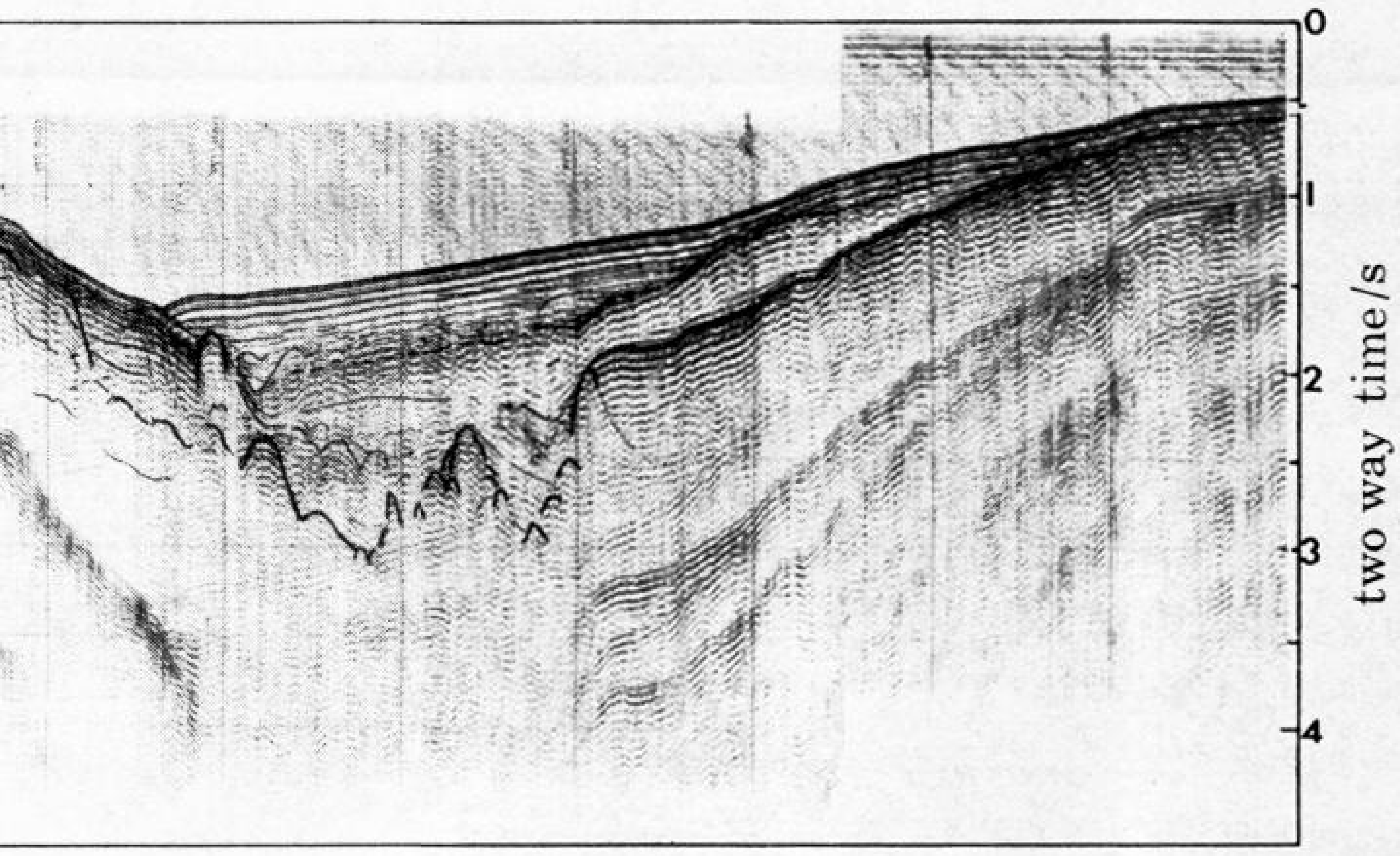
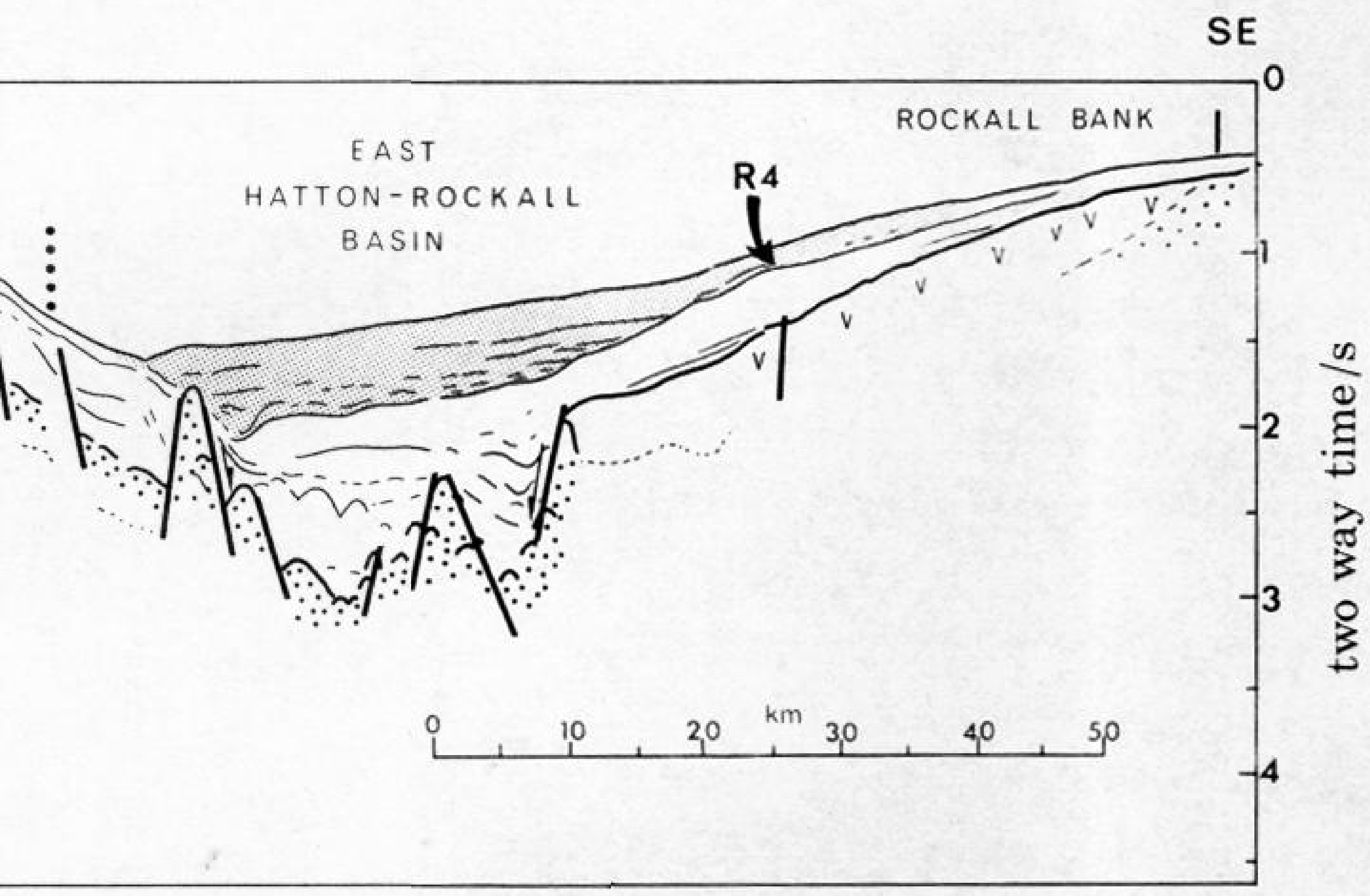


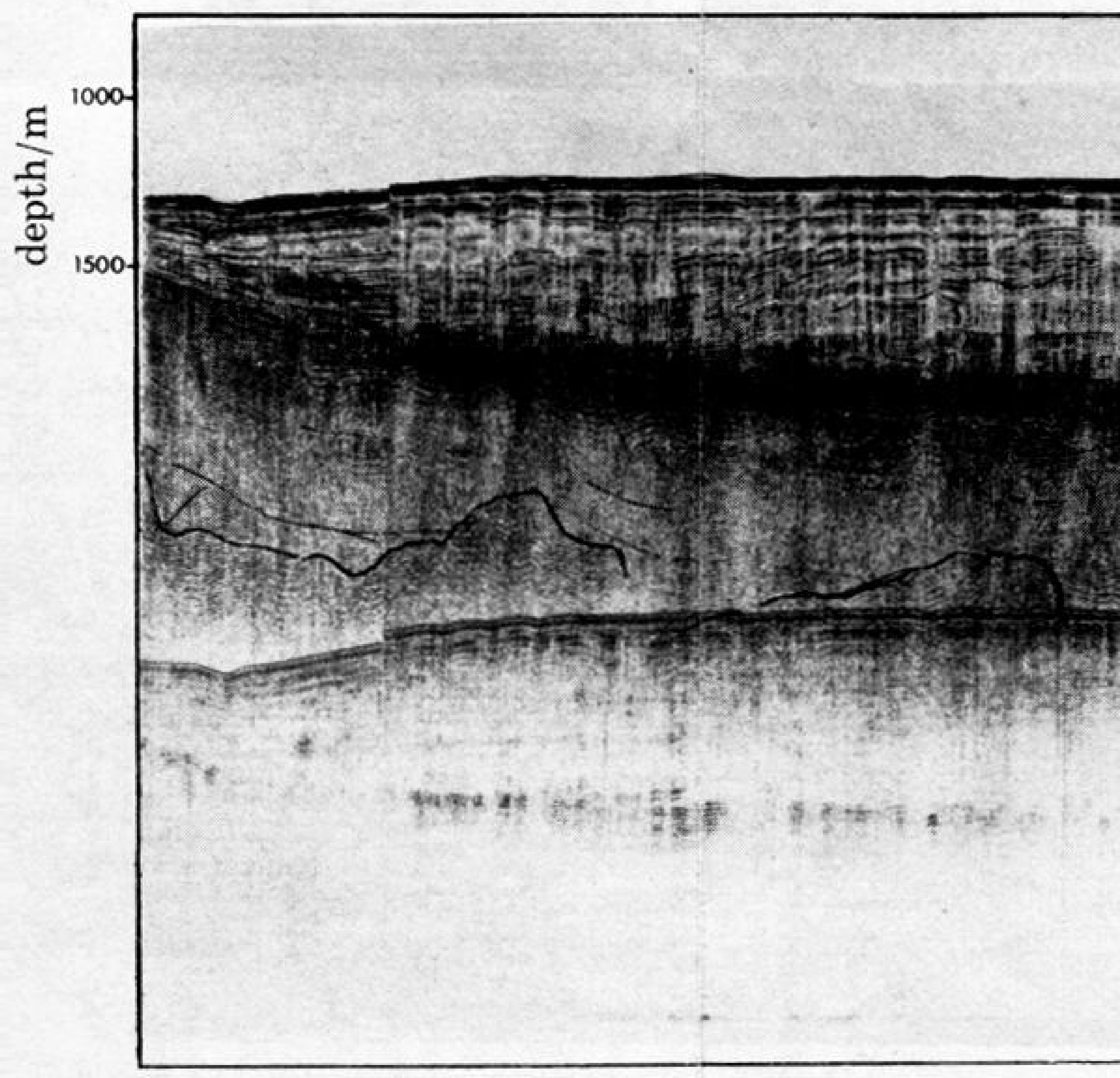
FIGURE 20. Seismic reflexion profile (20) across the central Hatton-Rockall Basin. Profi







in. Profile is located in figure 12.



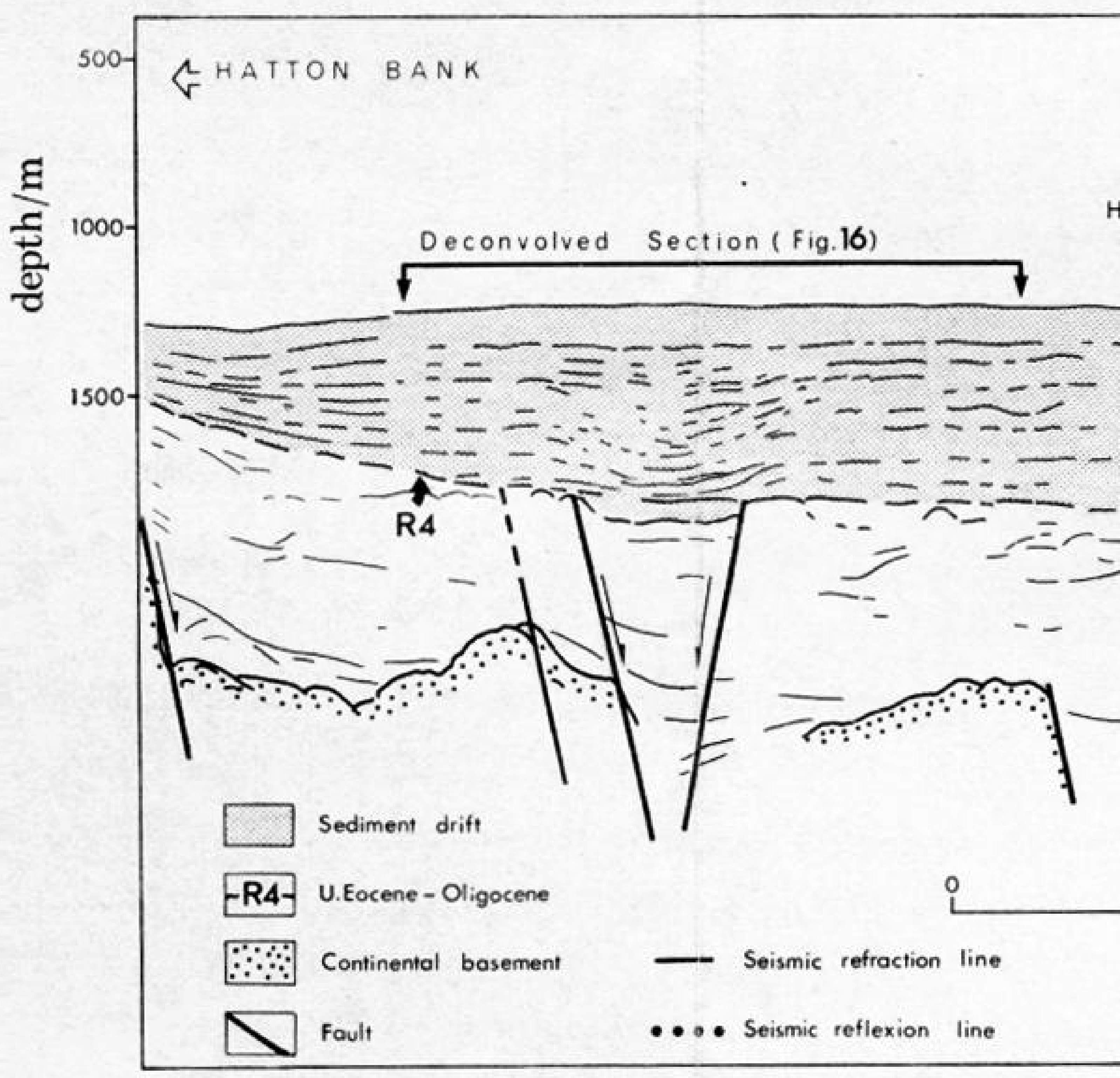
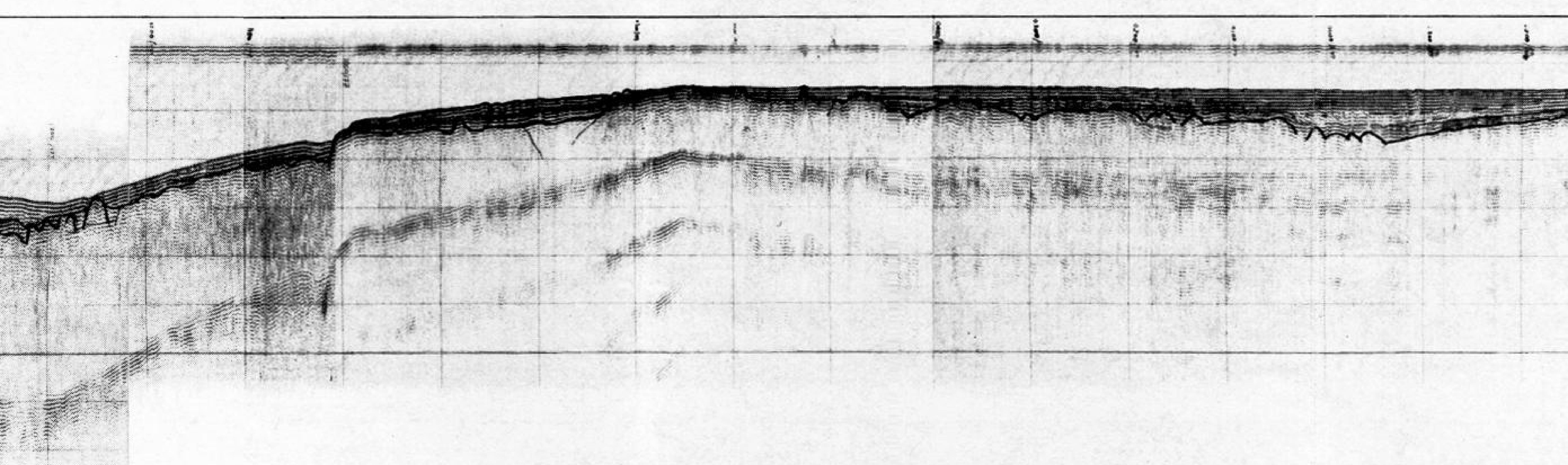
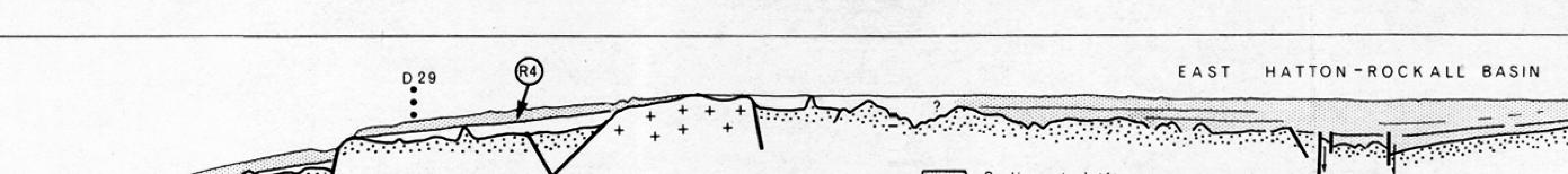
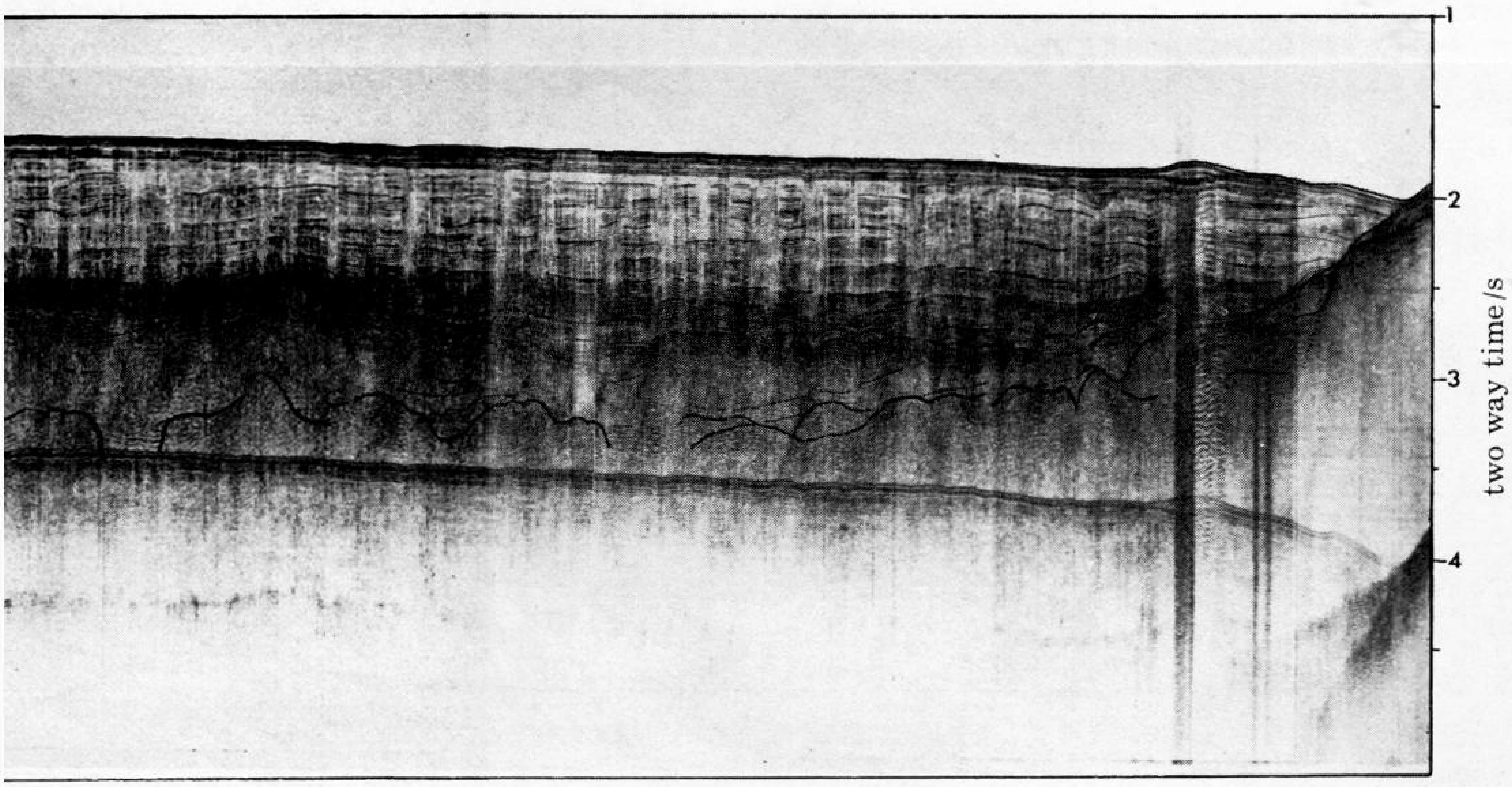
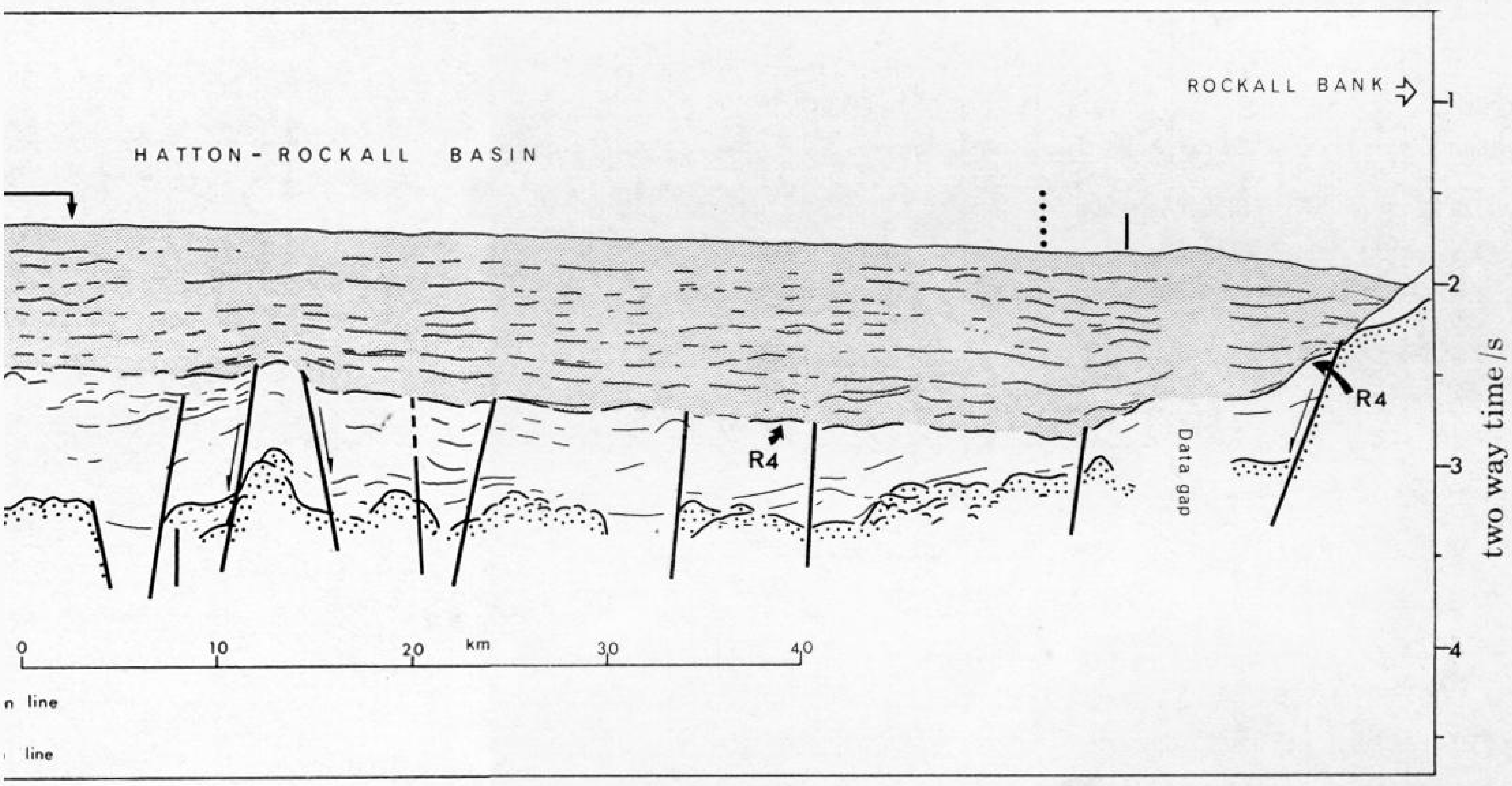


FIGURE 22. Seismic reflexion profile (

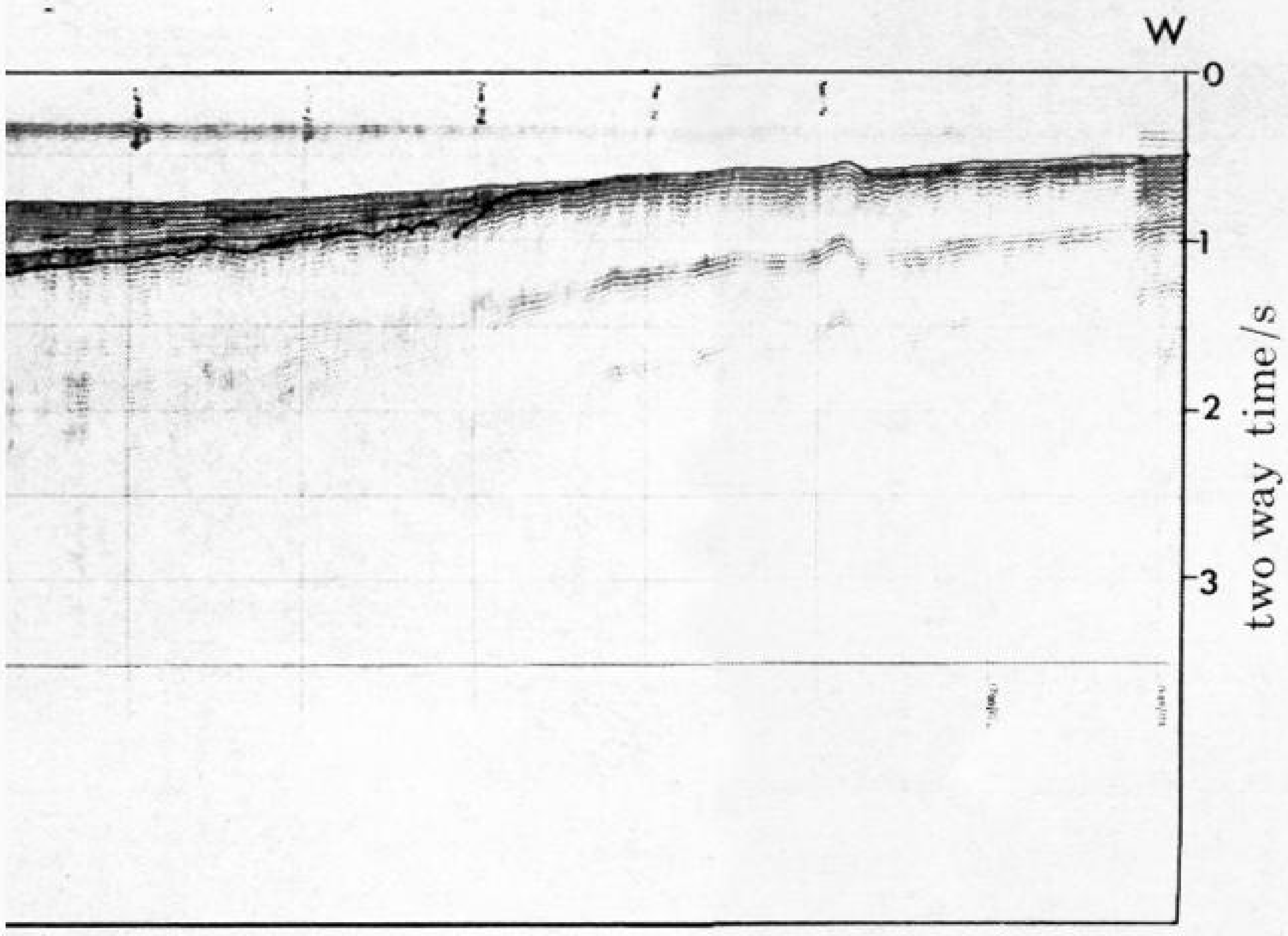


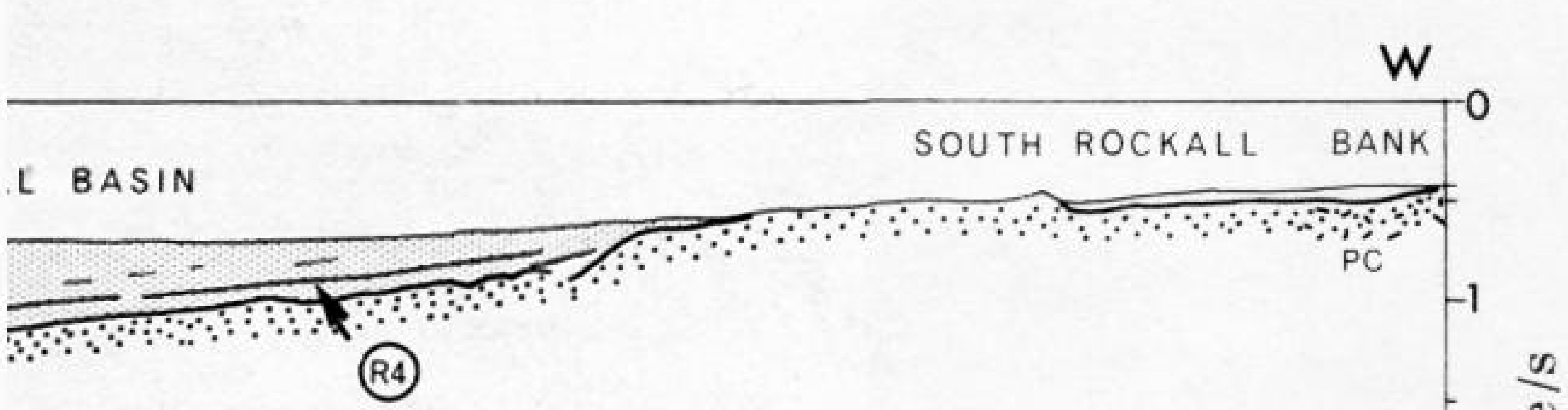






n profile (22) across the central Hatton-Rockall Basin. Profile is located in figure 12.





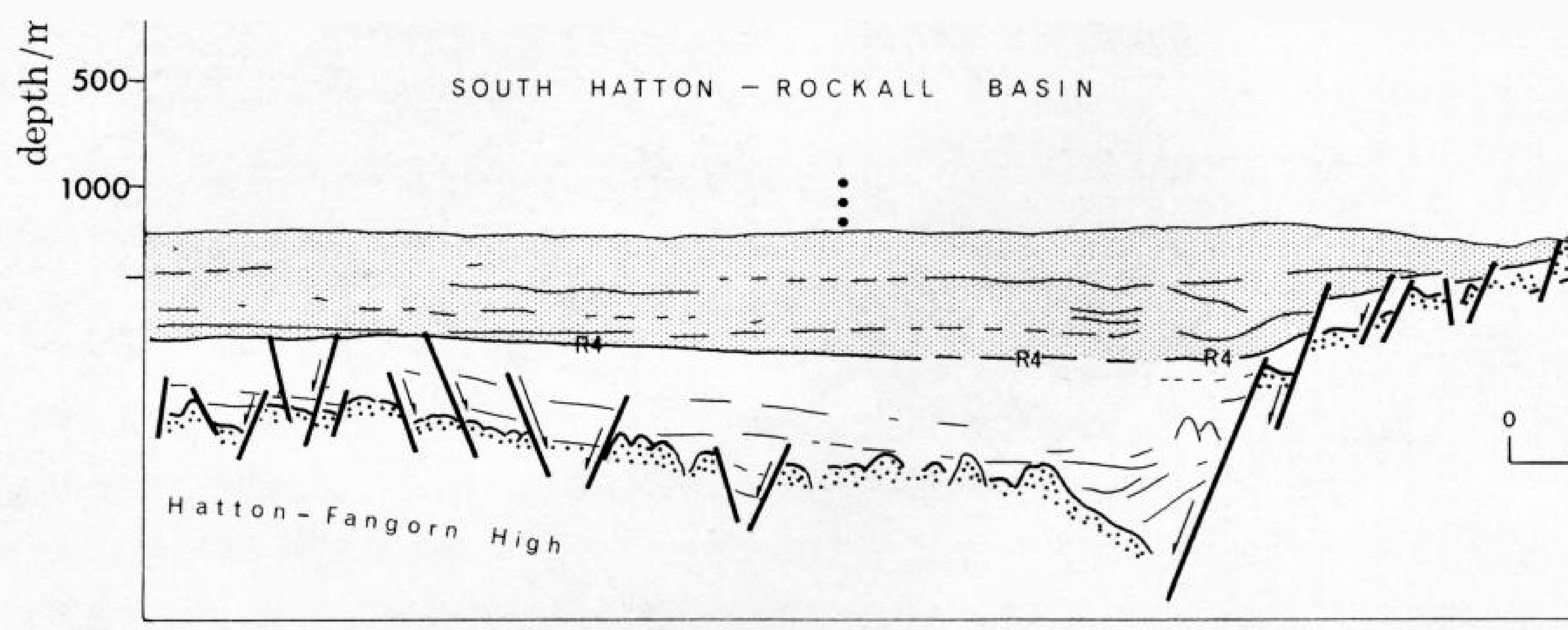
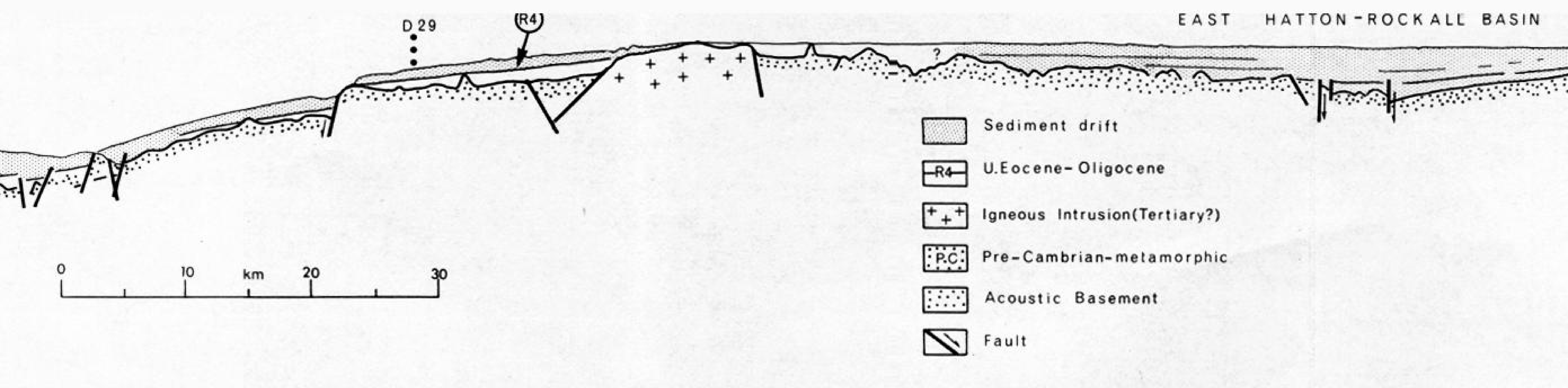
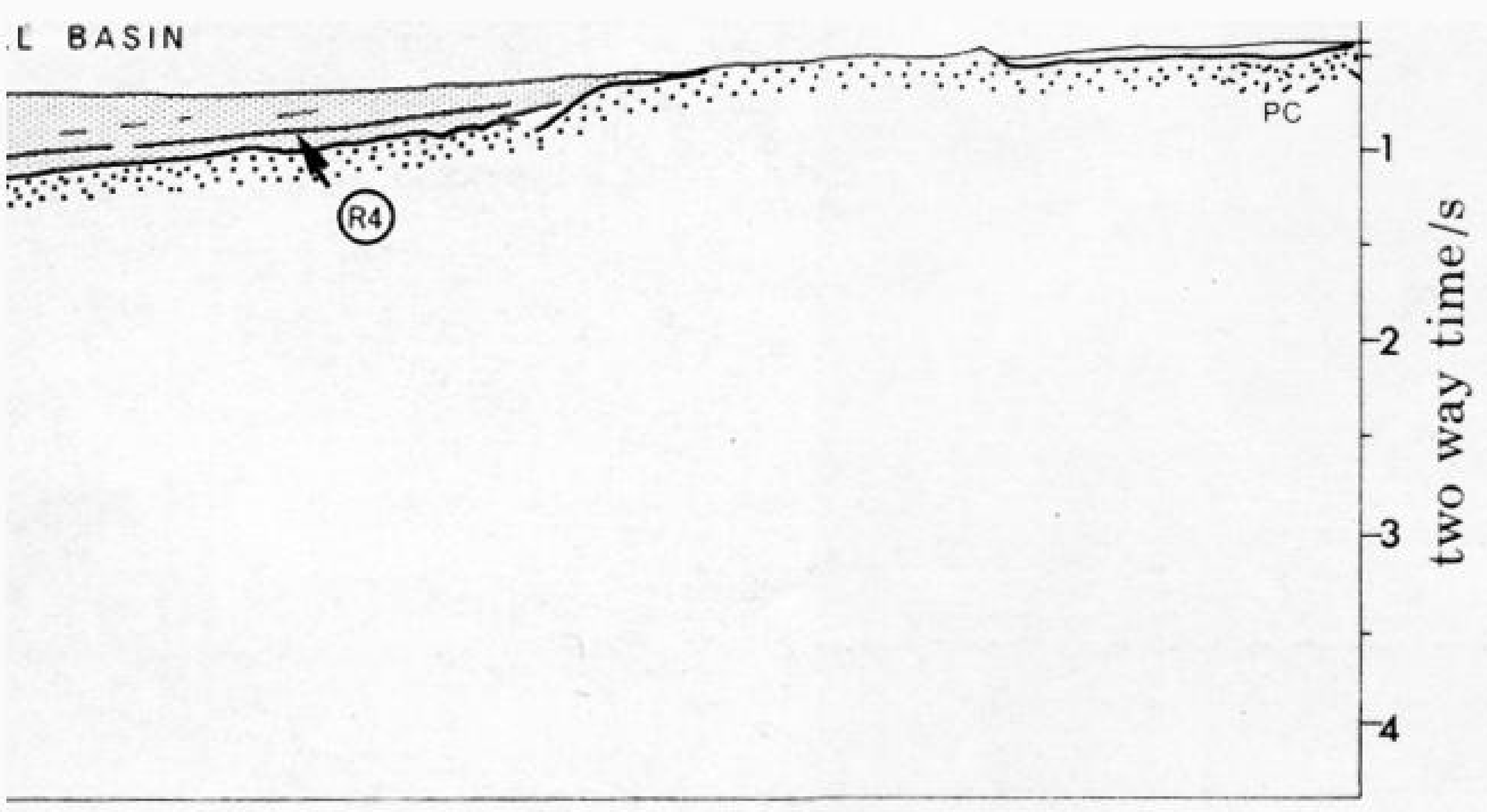


FIGURE 23. Seismic reflexion



eflexion profile (23) between the south Hatton-Rockall Basin, Fangorn High and Rockall Bank. Profile is located in figure



n figure 12.

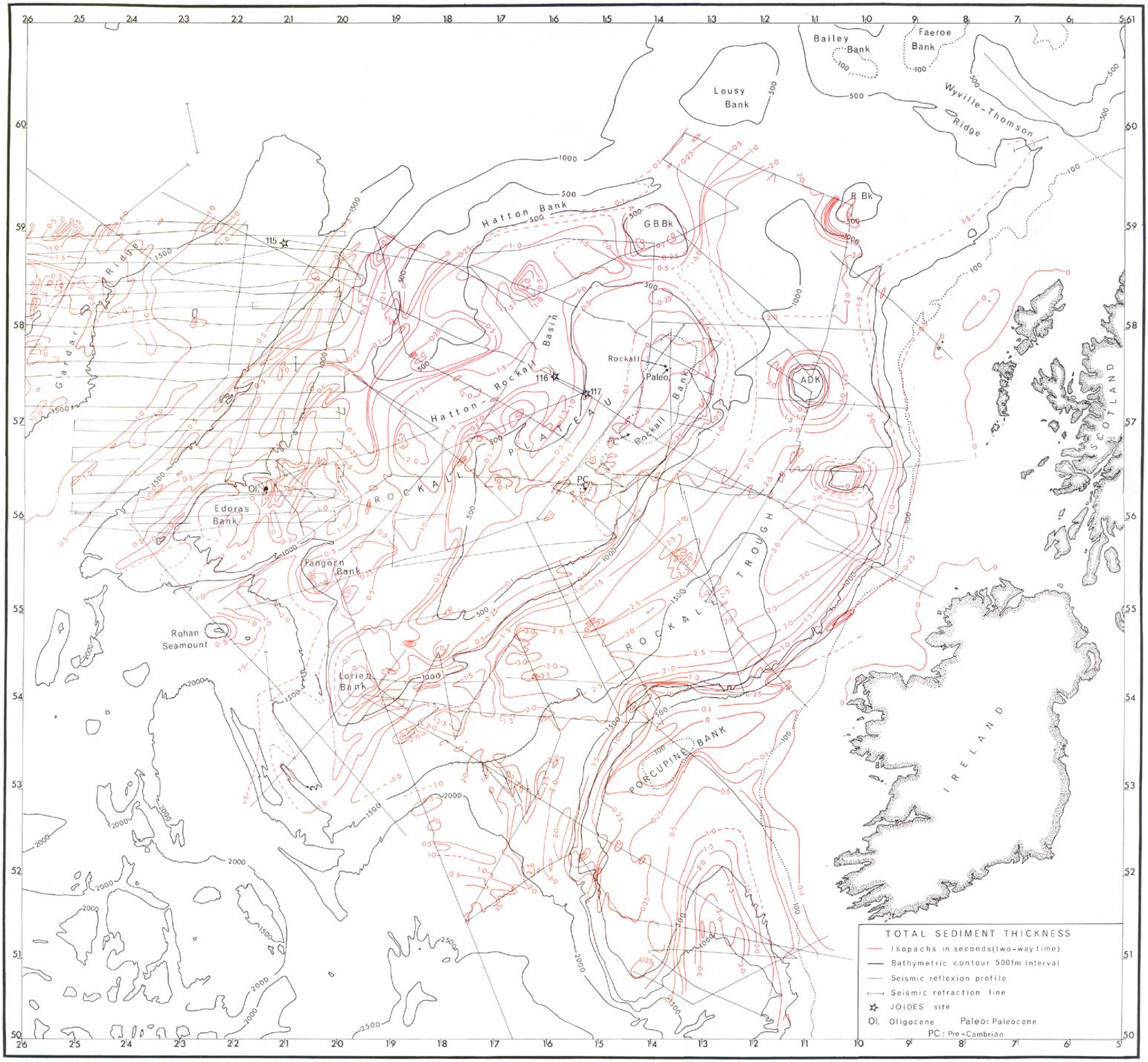
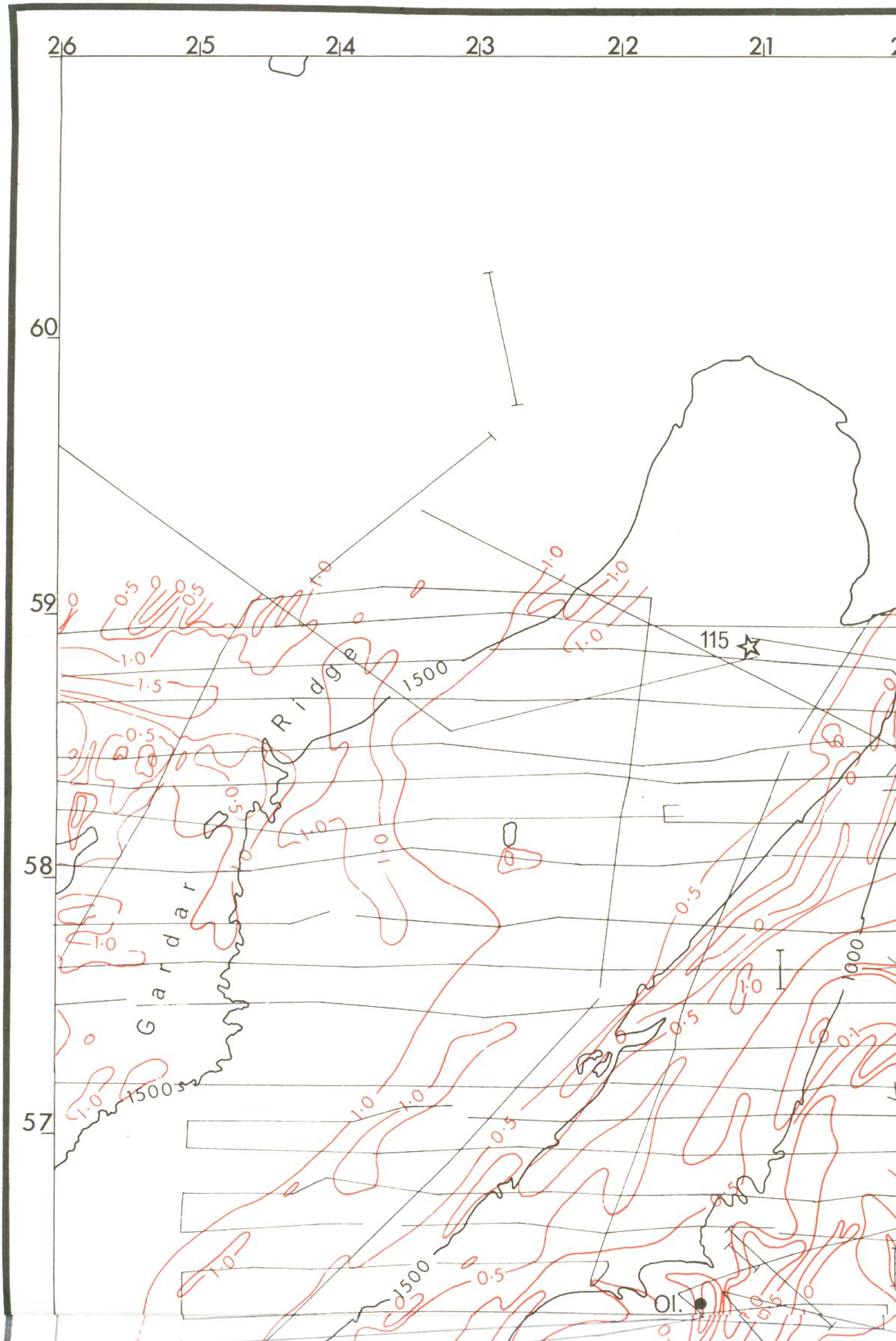
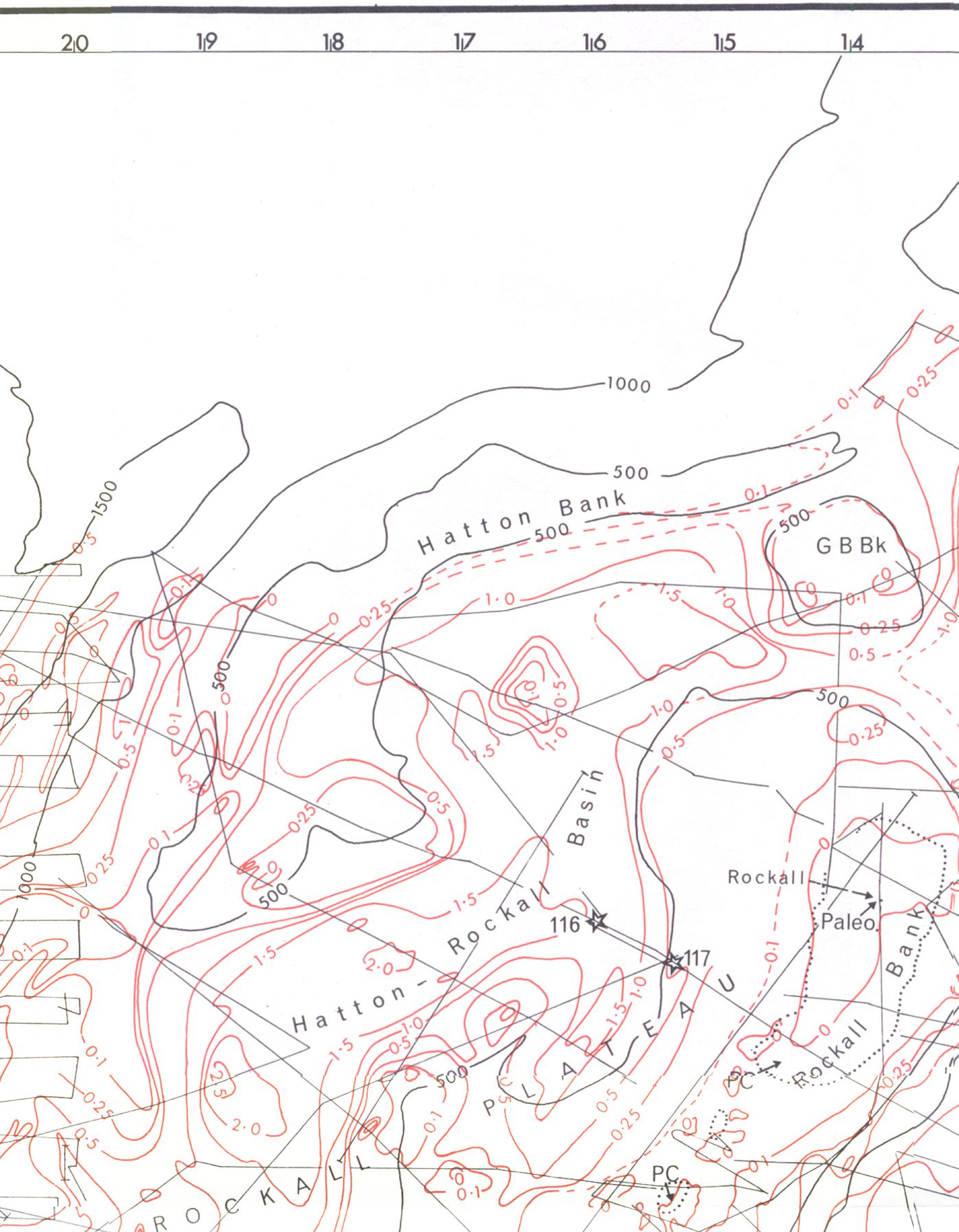
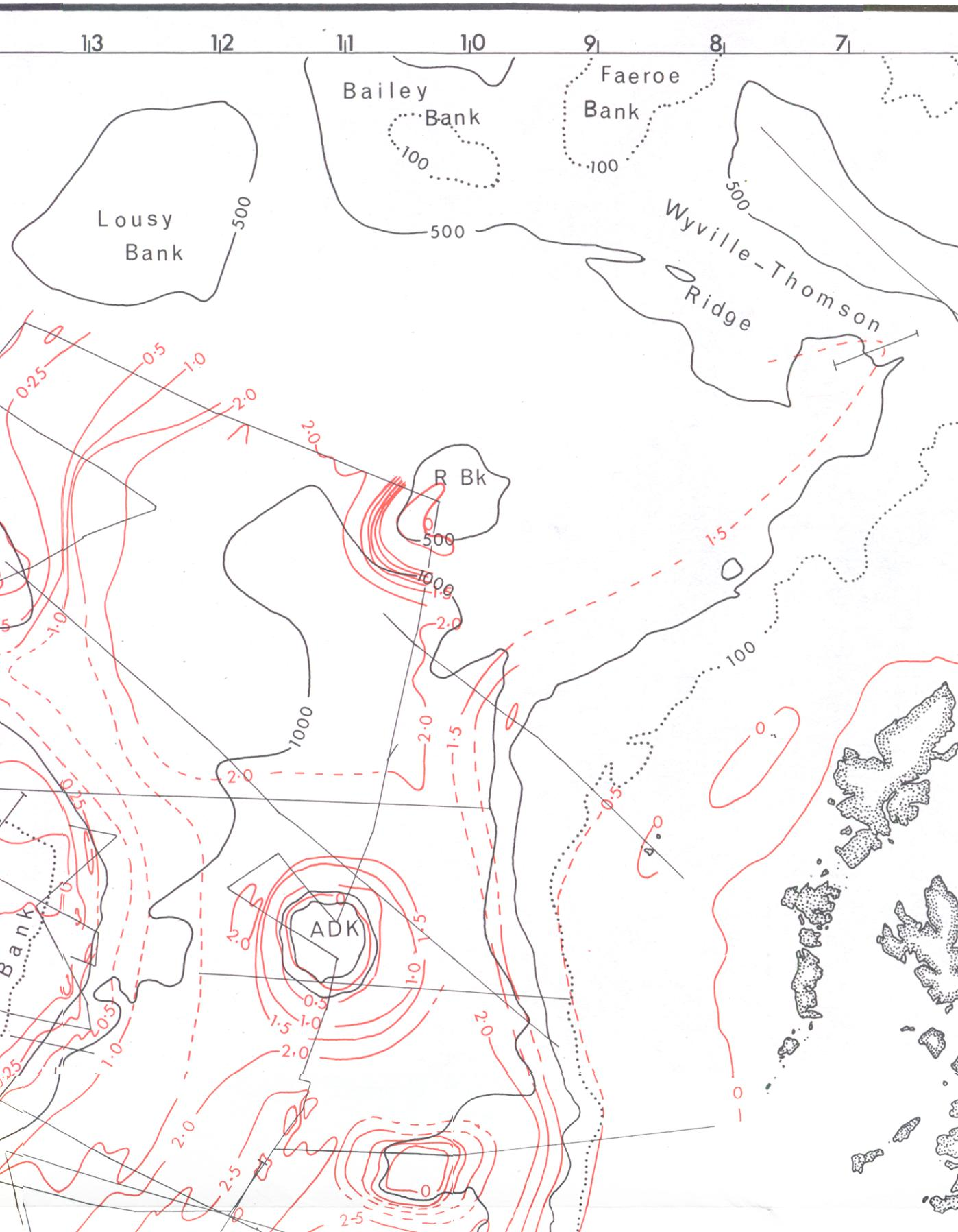


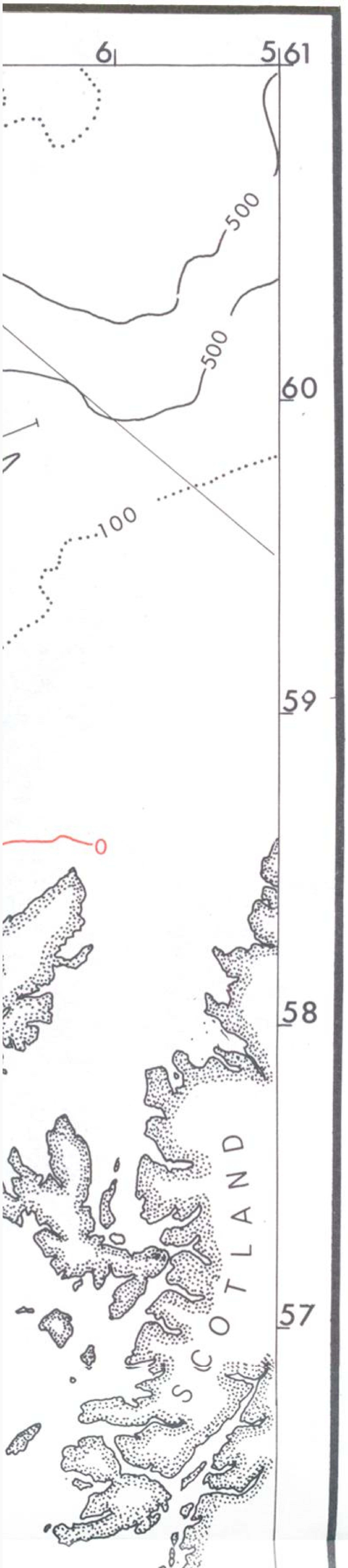
Figure 18. Isopachs on basement in two way time below seabed.


26
25
24
23
22
21
60
59
58
57
115
01.
Gardar Ridge
1500
1000
0·5
1·0
1·5
0



20
19
18
17
16
15
14
1000
1500
500
Hatton Bank
G B Bk
Hatton - Rockall Basin
116
117
Rockall
Paleo.
Rockall Bank
PC
ROCKALL PLATEAU

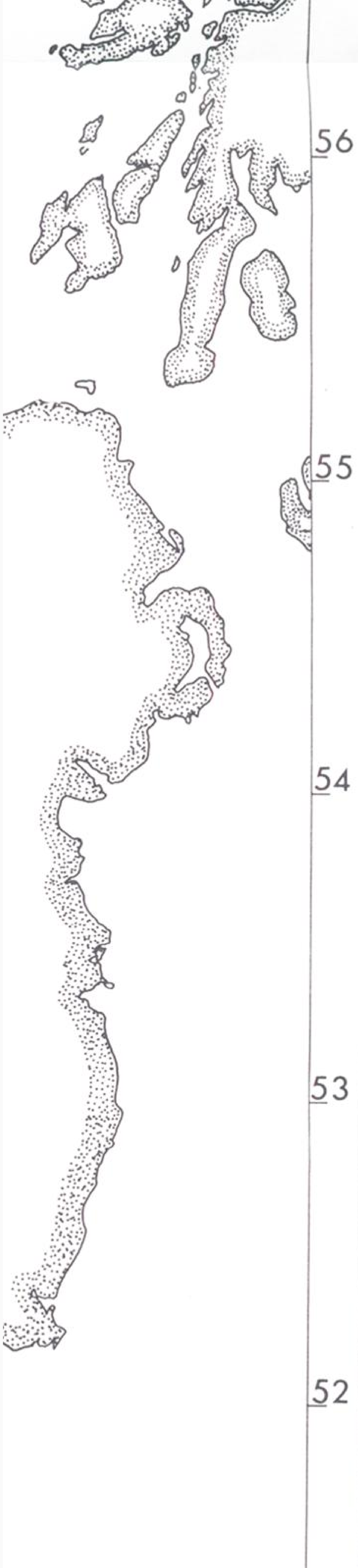

13
12
11
10
9
8
7
Bailey Bank
Faeroe Bank
Lousy Bank
Wyville-Thomson Ridge
R Bk
ADK
Bank
100
500
1000
0
0·25
0·5
1·0
1·5
2·0
2·5


6
5
61
500
500
60
100
59
0
58
SCOTLAND
57

56
55
54
53
52
51
Edoras
Bank
OI.
Fangor
Rohan
Seamount
1000
1500
2000
0·5
1·0
1·5
2·0
0


ROCKALL
PC
Rosemary?
angorn Bank
Lorien Bank
ROCKAL
PORCUPINE B
100
500
1000
1500
2000
2500

TROUGH
BANK
IRELAND
TOTAL SEDIMENT THICKNES
Isopachs in seconds(two-way time)
Bathymetric contour 500fm interval
Seismic reflexion profile
Seismic refraction line
JOIDES site
Ol. Oligocene
Paleo: Paleocene



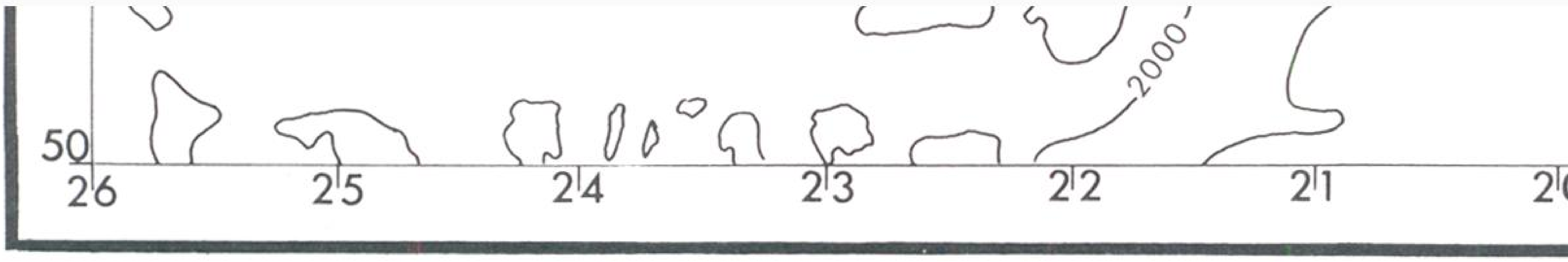
KNESS

time)

erval

51

50


2000
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26
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24
23
22
21

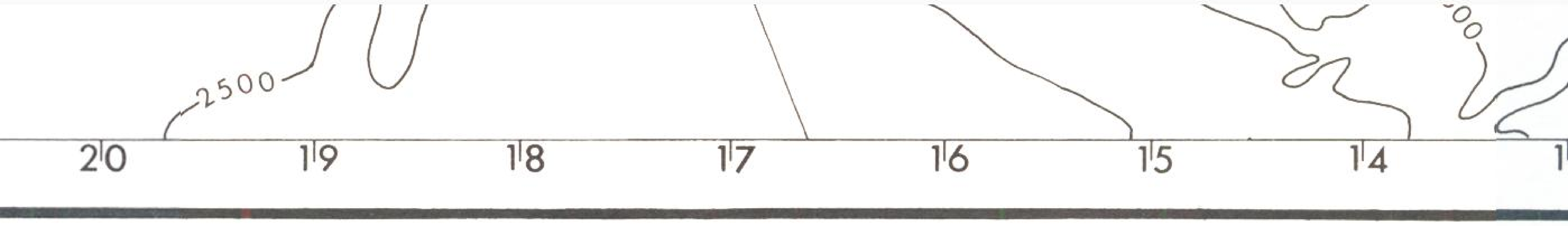
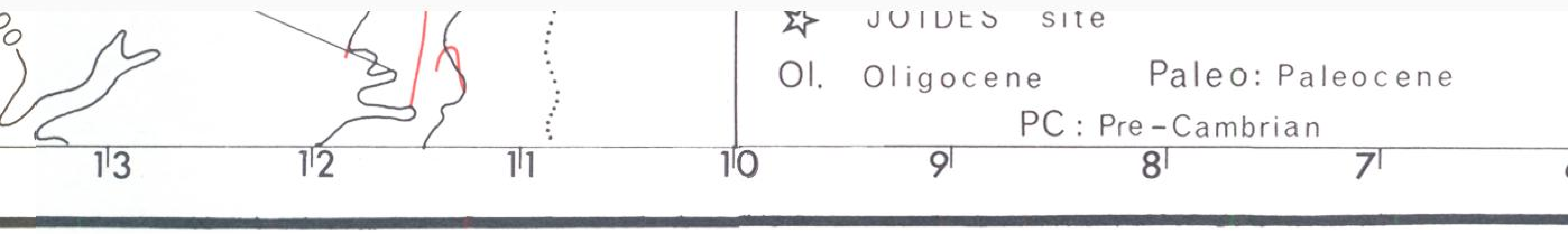
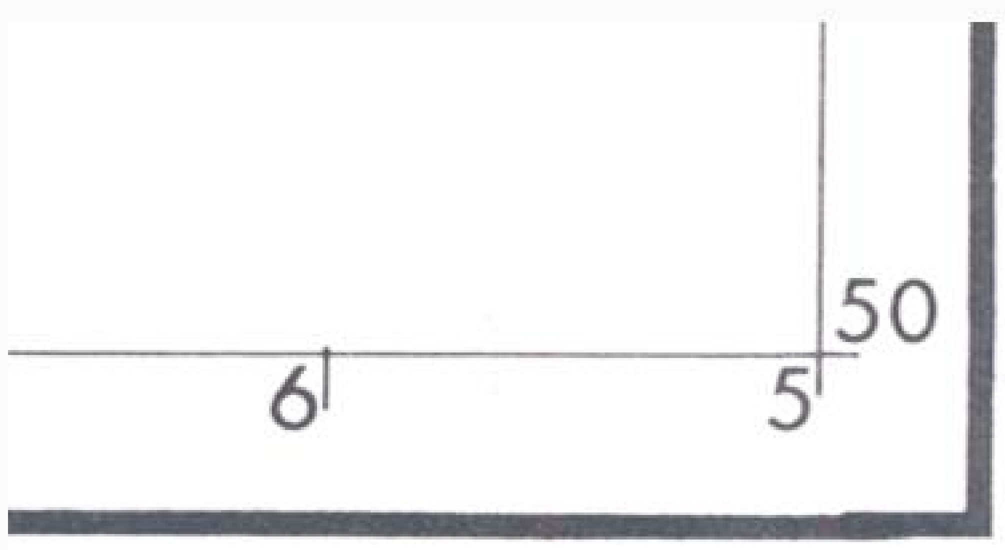


FIGURE 18. Isopachs on basement in two way time below seabed.


JOIDES site
Ol. Oligocene
Paleo: Paleocene
PC: Pre-Cambrian
13
12
11
10
9
8
7


50
6
5

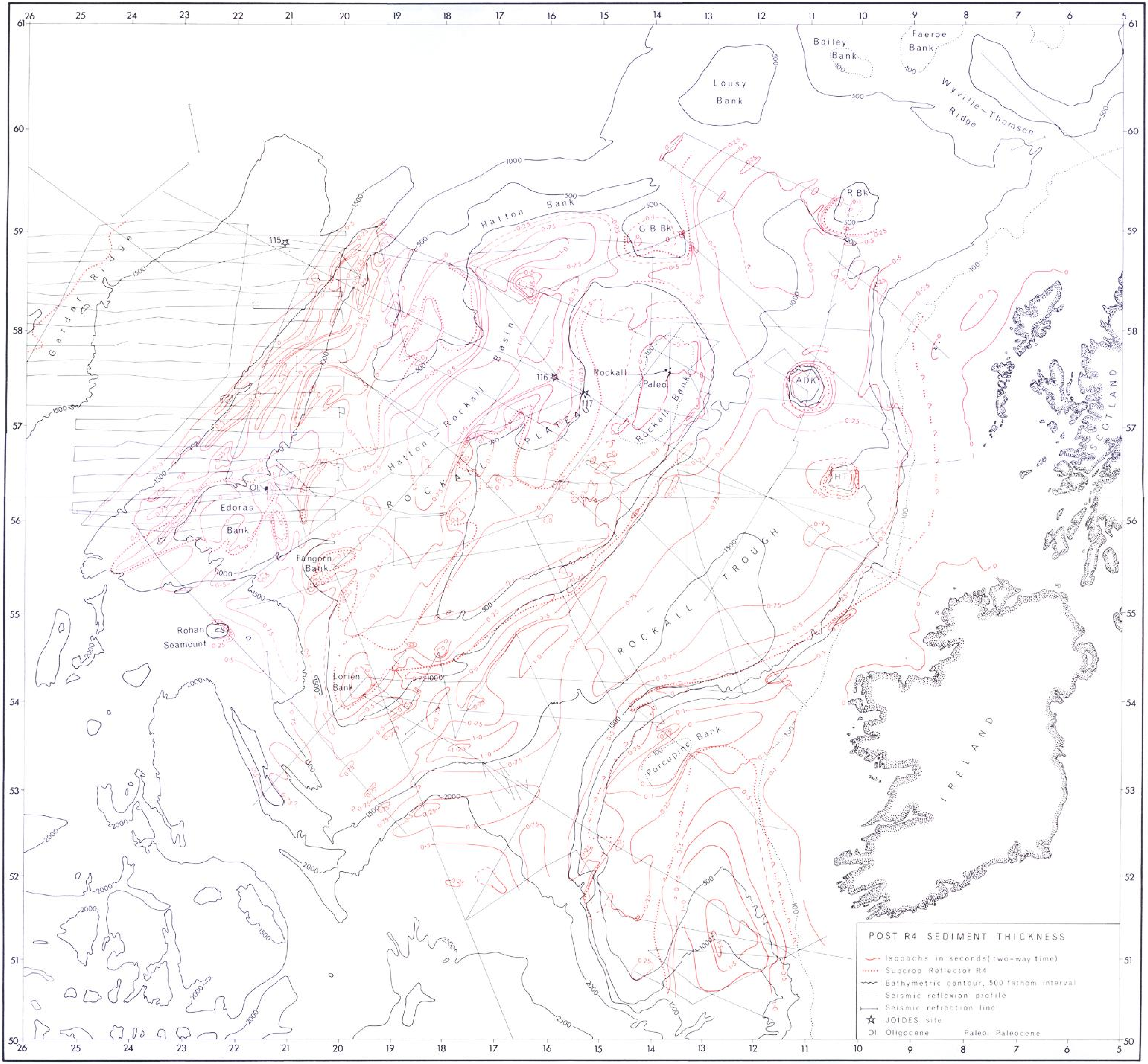


FIGURE 27. Isopachs on R4 in two way time below the seabed.


26
25
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61
60
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58
57
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1500
1000
115
Gardar Ridge
Edoras
Bank
Ol.
0·5
0·25
0·75
1·0
0·1
0


19
18
17
16
15
14
13
Lousy Bank
1000
500
Hatton Bank
G B Bk
Hatton Rockall Basin
Rockall
Paleo.
Rockall Bank
116
117
PLATEAU
Hatton
ROCKALL
100
0
0·1
0·25
0·5
0·75
1·0


12
11
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8
7
6
Bailey
Bank
Faeroe
Bank
ousy
Bank
Wyville–Thomson
Ridge
R Bk
ADK
HT
500
1000
100
0.25
0.5
0.75
0.1
0
0.9
?


5
61
500
60
59
58
SCOTLAND
57
56

Edoras
Bank
Fangorn
Bank
Rohan
Seamount
Lorien
Bank
1000
1500
2000
0·25
0·5
0·75
56
55
54
53
52
51
50
26
25
24
23
22
21
20

ROCKALL TRO
Porcupine Bank
100
500
1000
1500
2000
2500
0·1
0·25
0·5
0·75
1·0
1·25
1·5
19
18
17
16
15
14
13


TROUGH
1500
IRELAND
0.9
0.75
0.5
0.25
0.1
0
100
1.5
1.0
POST R4 SEDIMENT THICKNESS
Isopachs in seconds (two-way time)
Subcrop Reflector R4
Bathymetric contour, 500 fathom interval
Seismic reflexion profile
Seismic refraction line
JOIDES site
Ol. Oligocene
Paleo: Paleocene
12
11
10
9
8
7
6


56
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52
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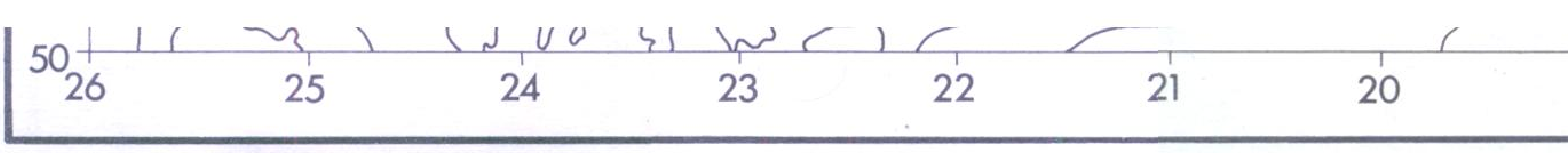
50
5



50
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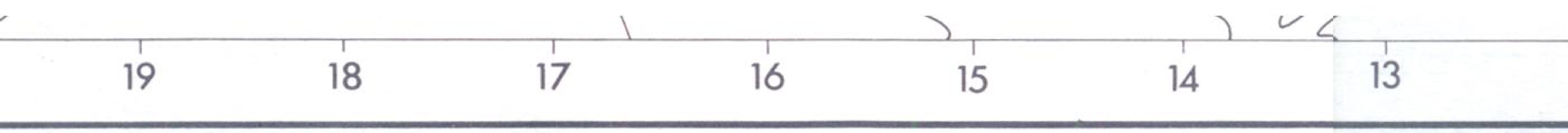
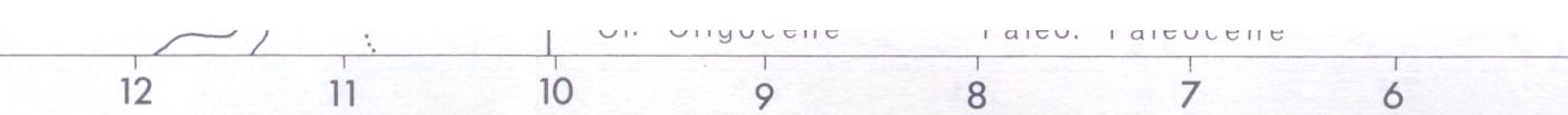
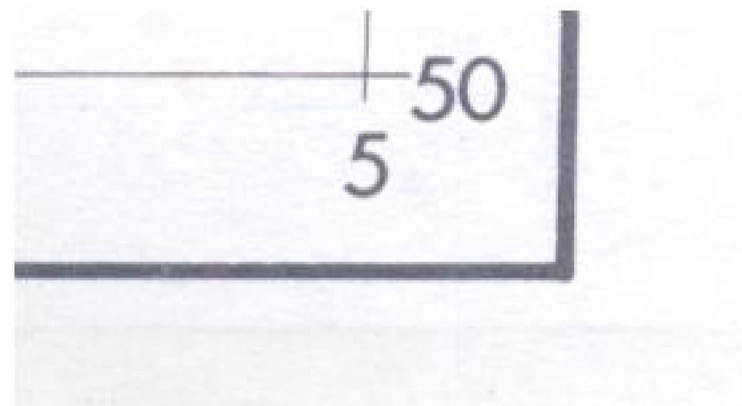


FIGURE 27. Isopachs on R4 in two way time below the seabed.


12
11
10
9
8
7
6


50
5

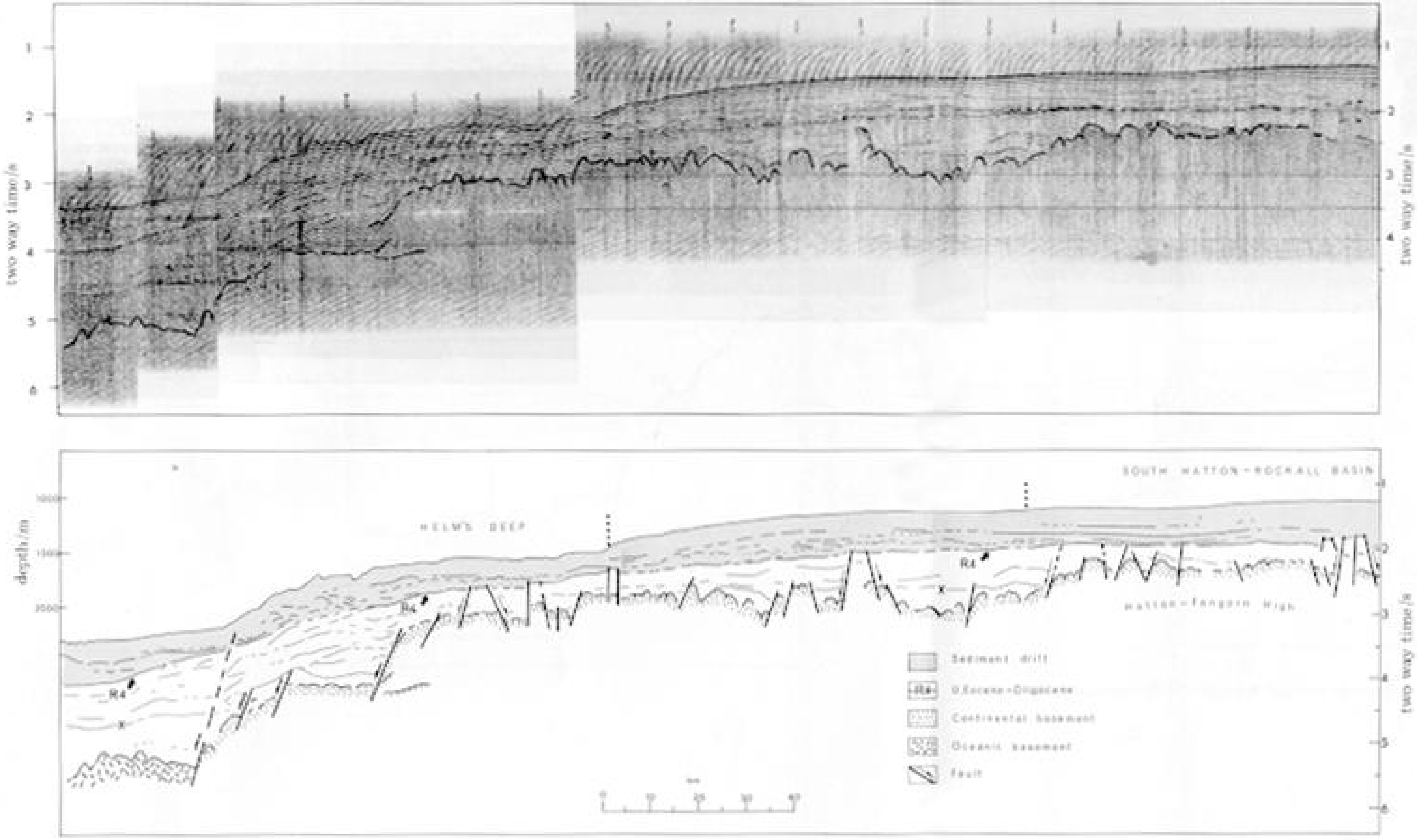


FIGURE 26. Seismic reflexion profile (26) across the southwest margin of the south Hatton-Rockall Basin. Profile is located in figure 12.

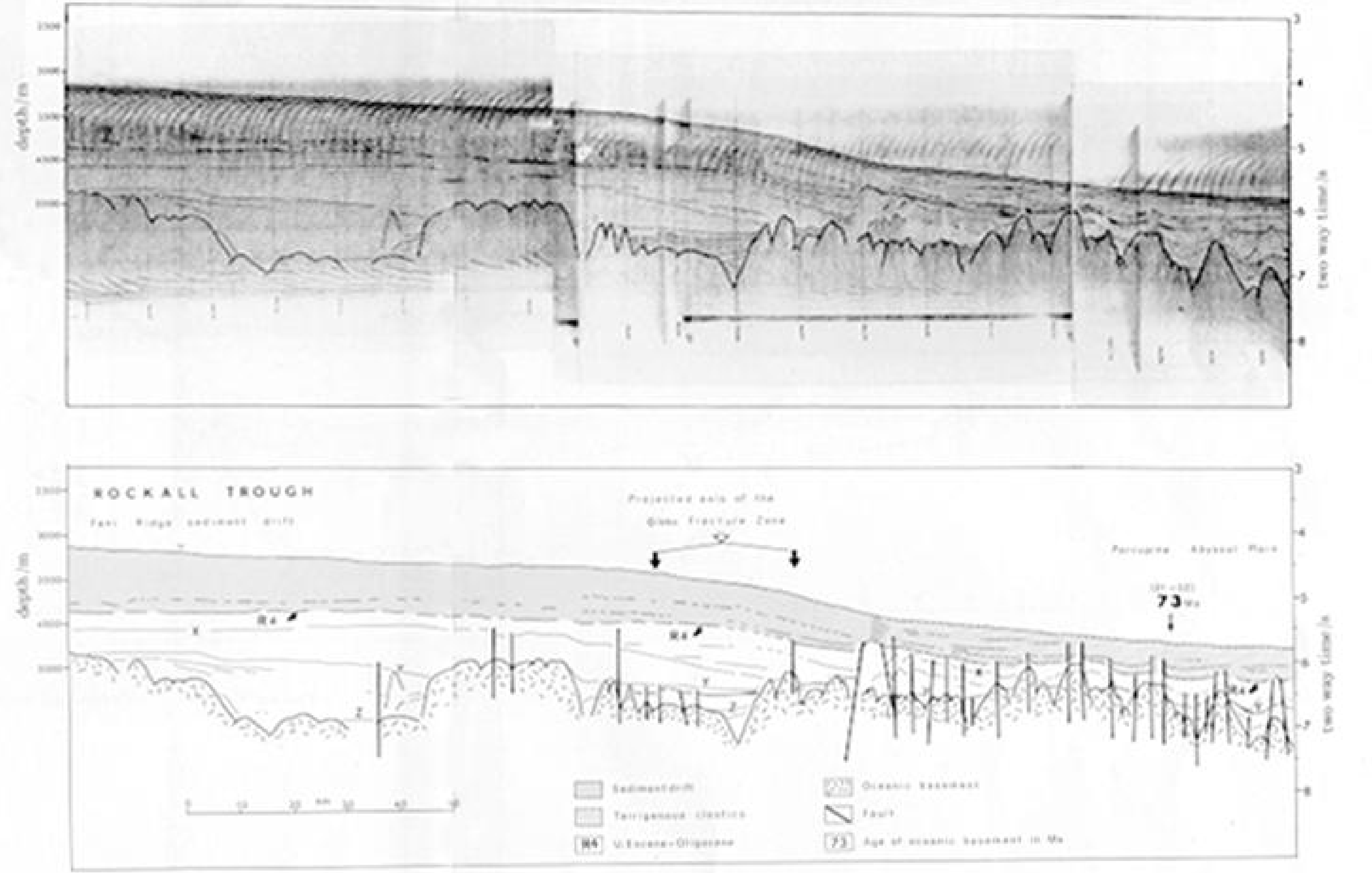
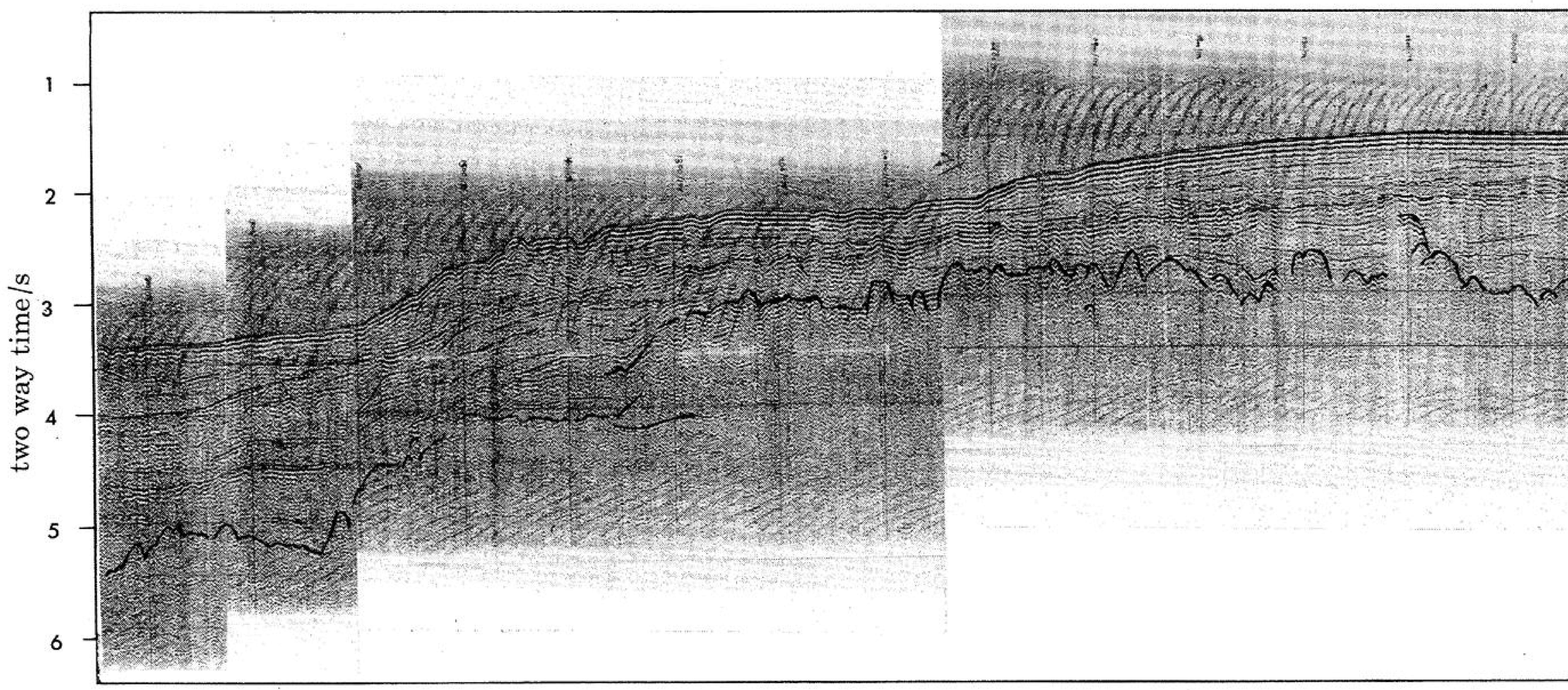


FIGURE 34. Longitudinal seismic profile in the entrance to the Rockall Trough showing relation to anomaly 32 and the Gibbs Fracture Zone. Profile is located in figure 31.



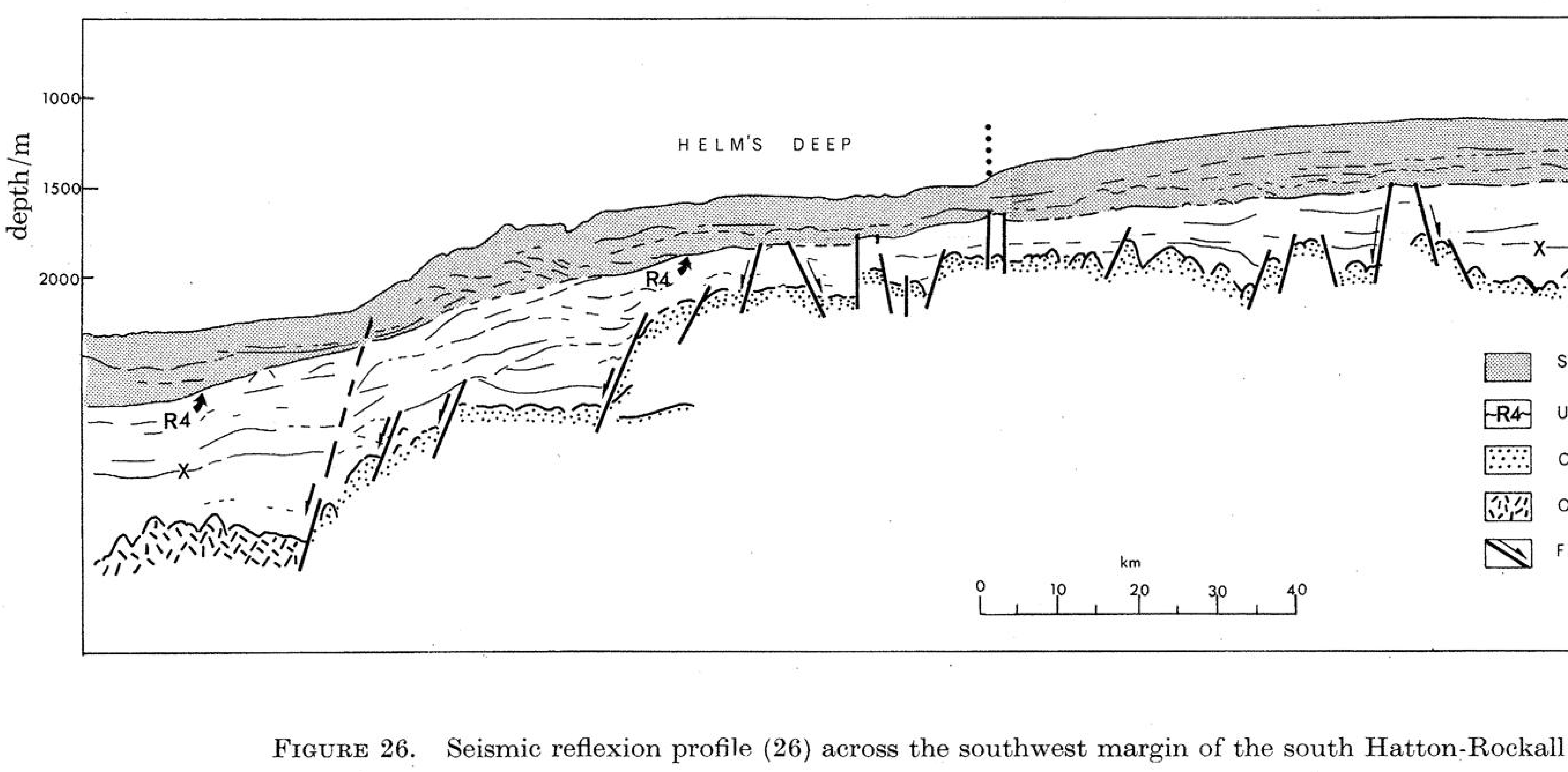
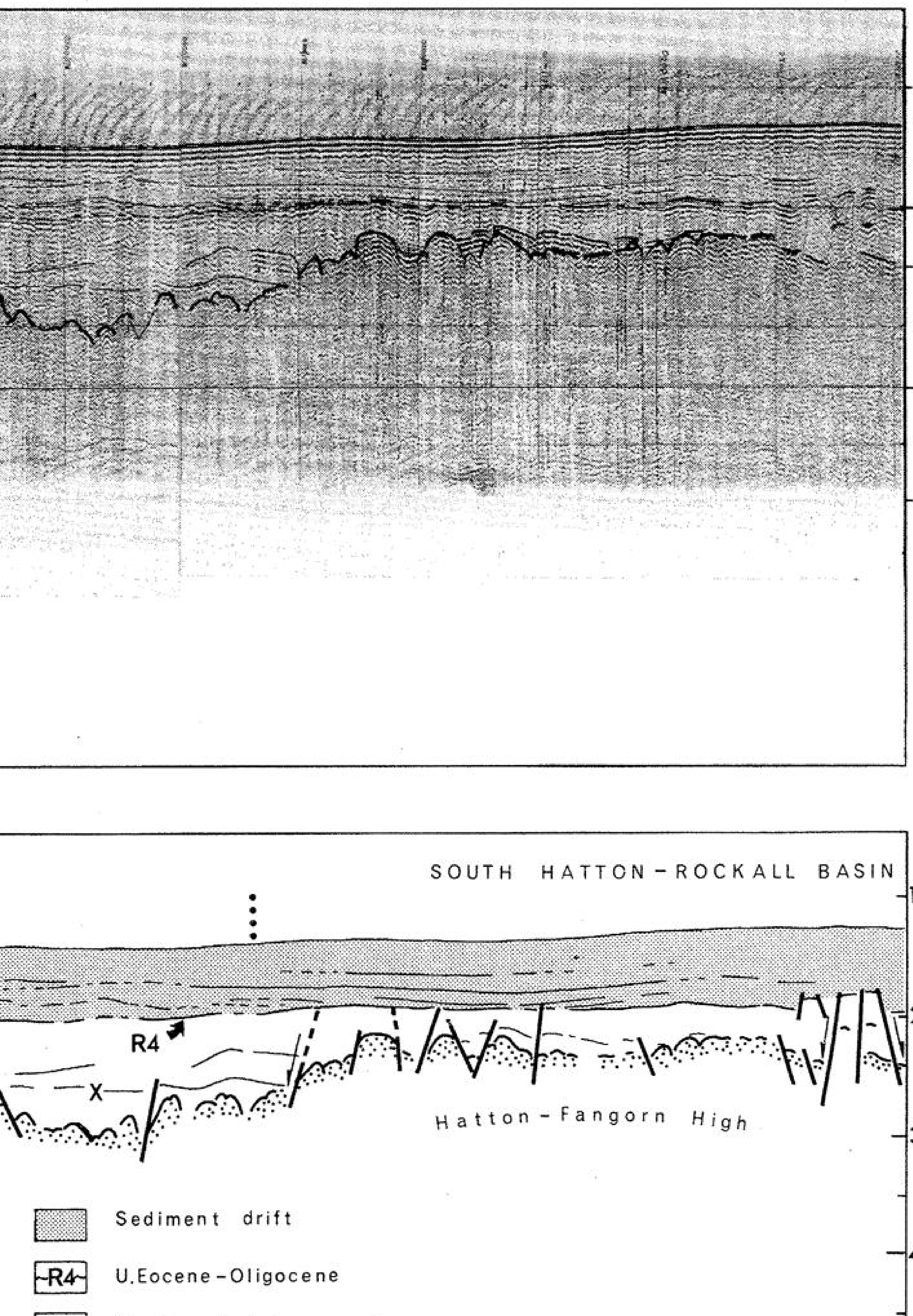


FIGURE 26. Seismic reflexion profile (26) across the southwest margin of the south Hatton-Rockall



-Rockall Basin. Profile is located in figure 12.

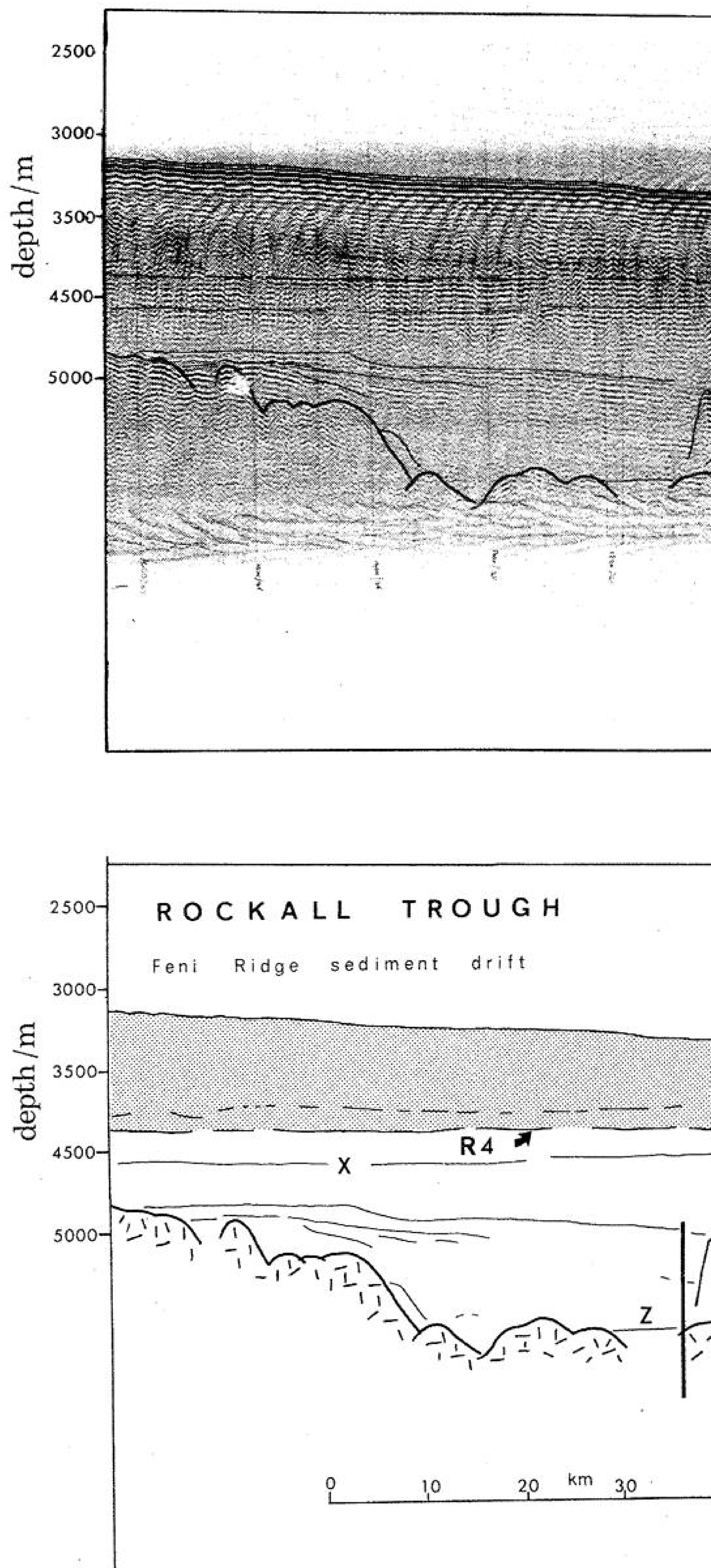
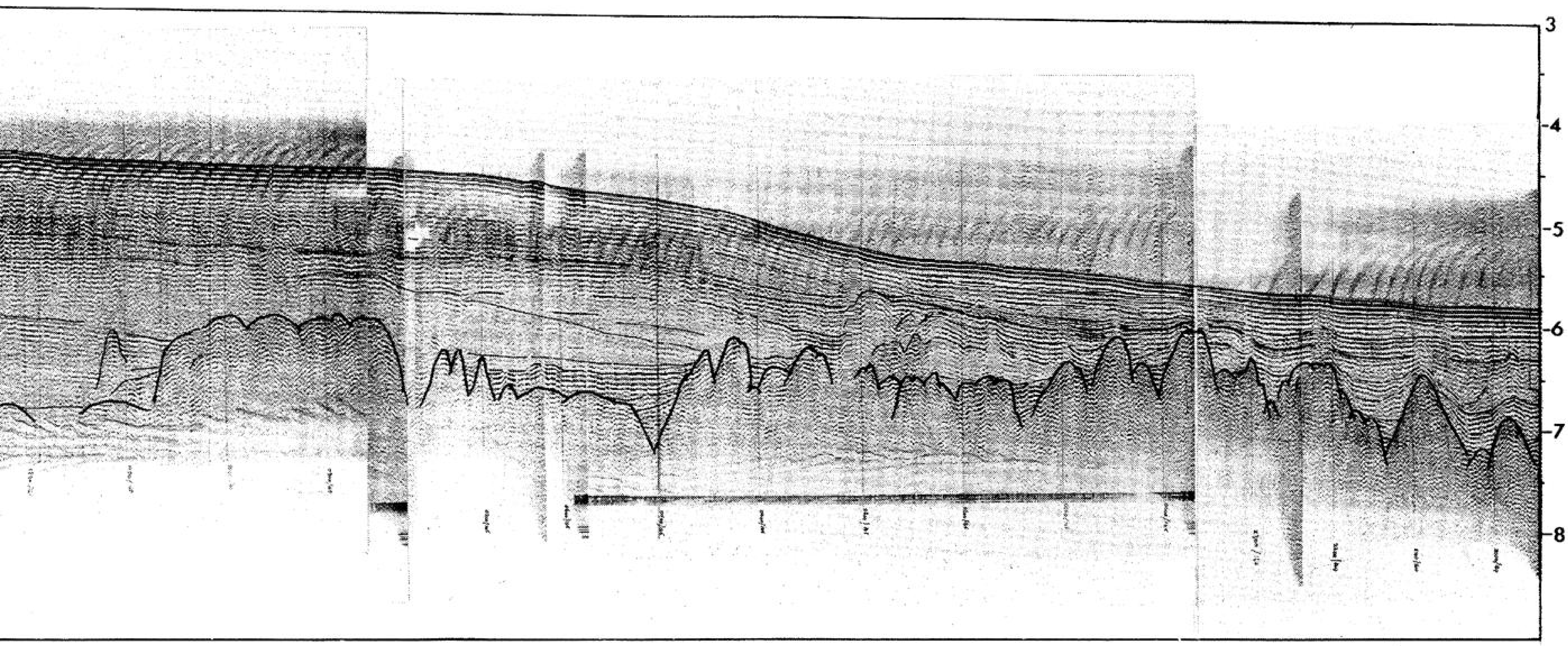
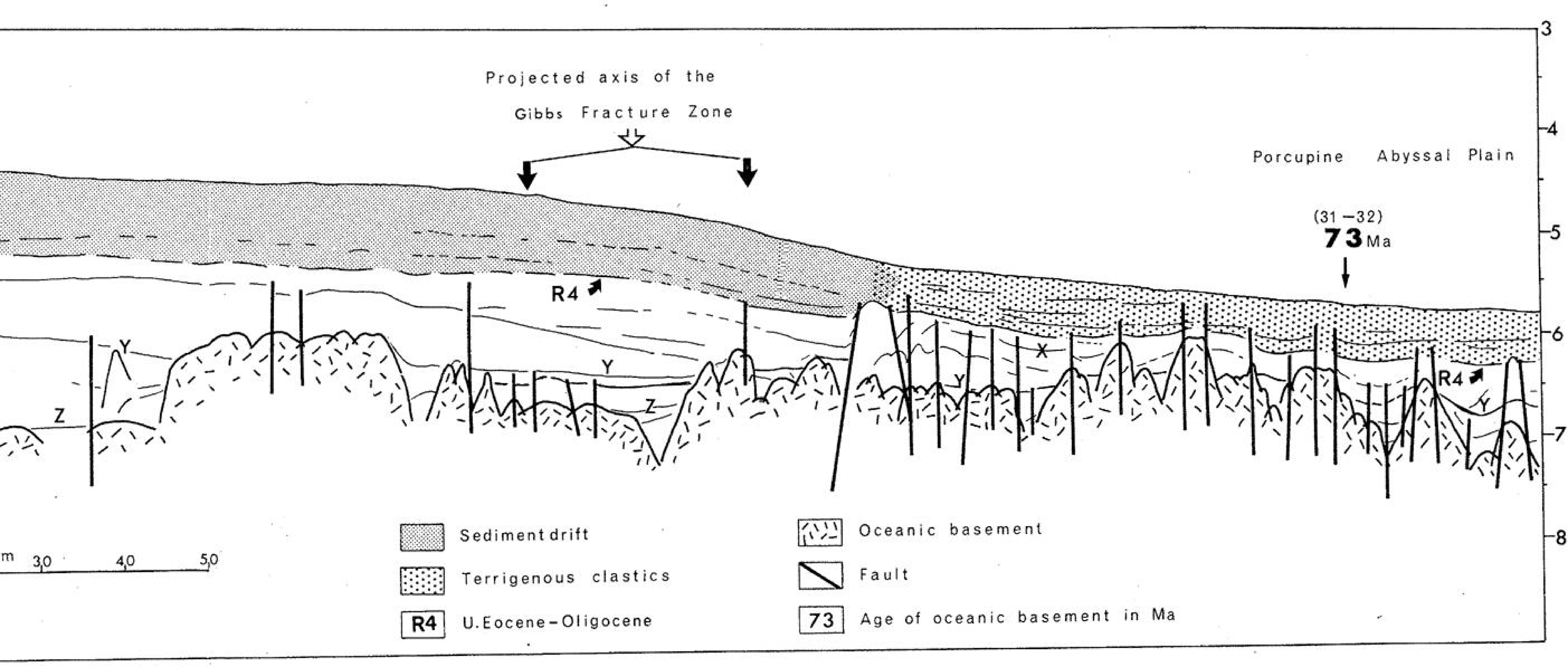
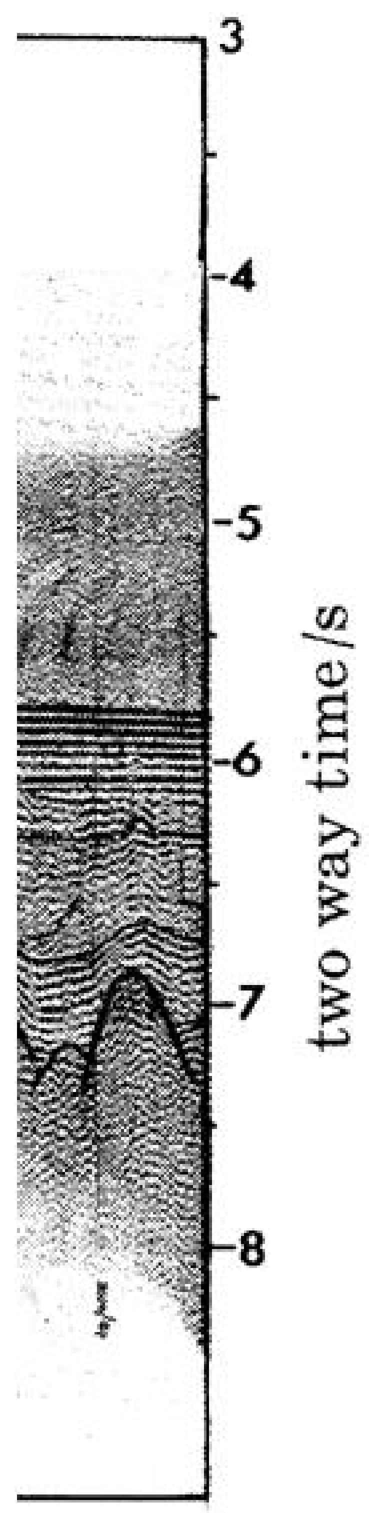


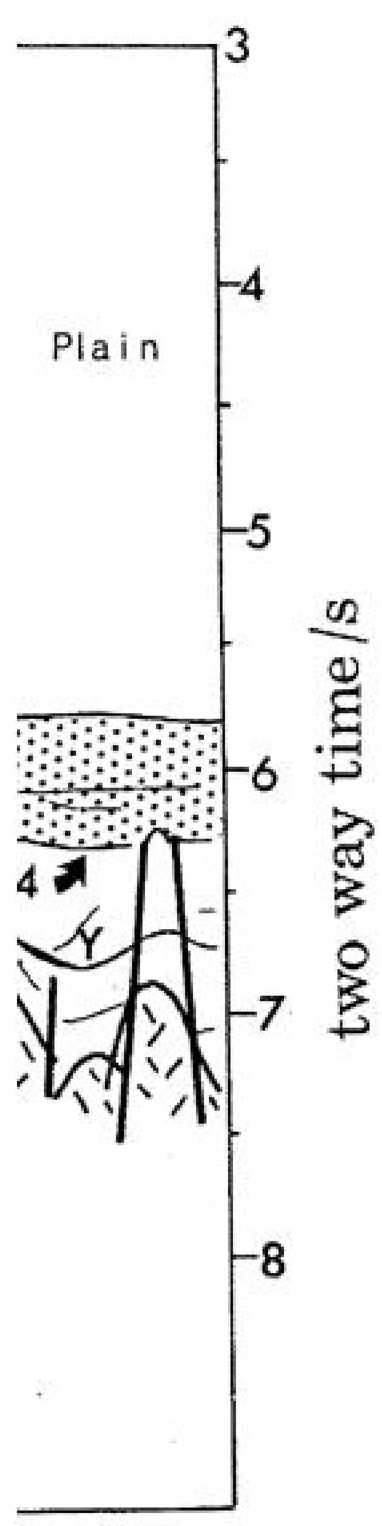
FIGURE 34. Longitudinal seismic profile in the en





in the entrance to the Rockall Trough showing relation to anomaly 32 and the Gibbs Fracture Zone. Profile is located in figu





ed in figure 31.

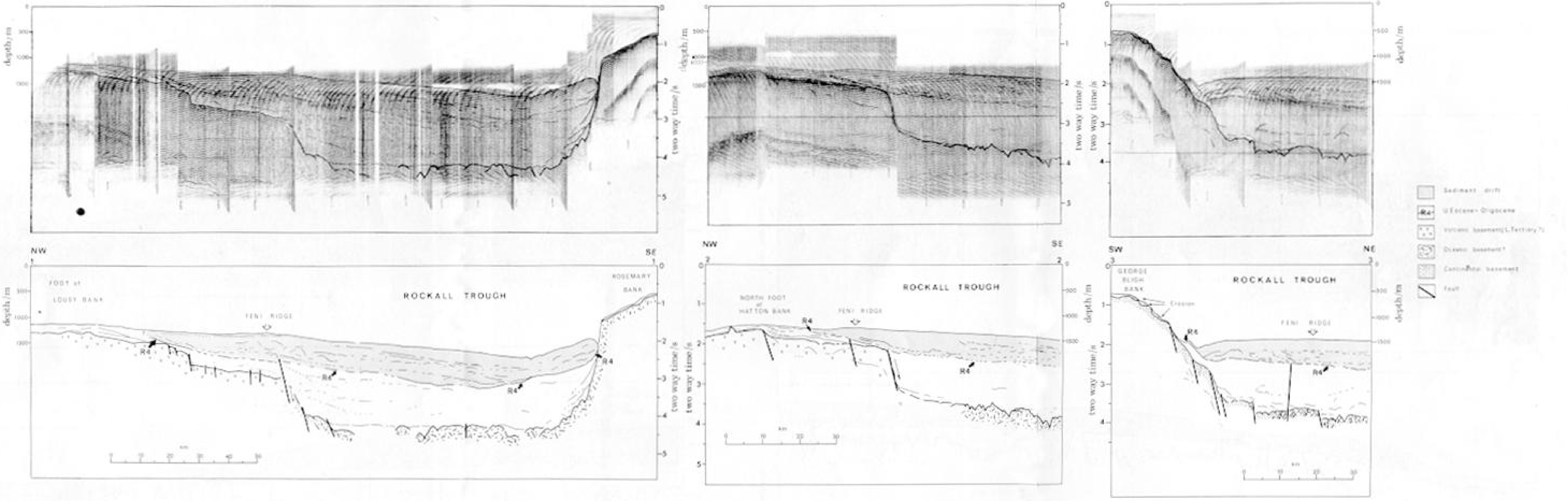
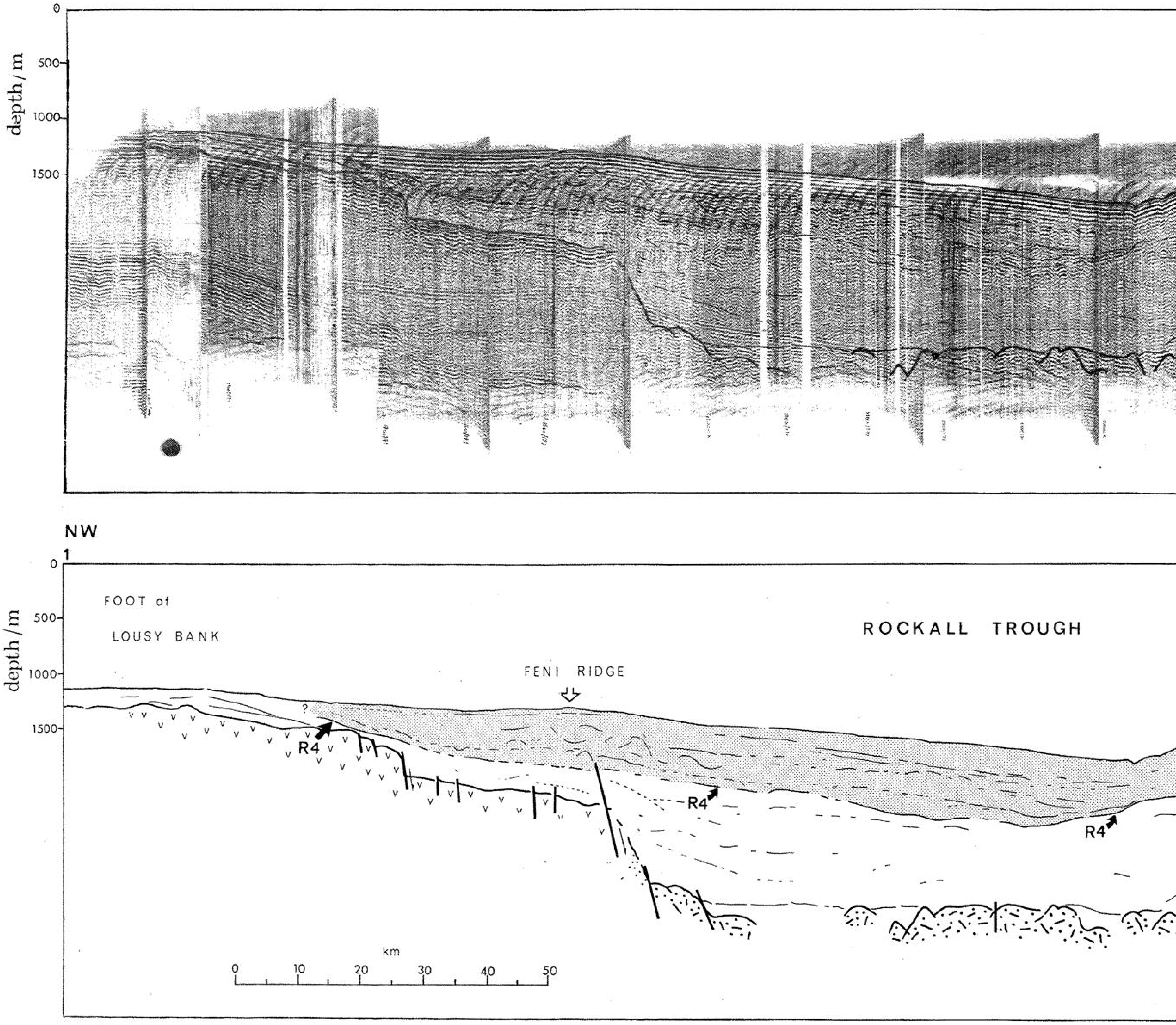


FIGURE 35. Seismic reflexion profiles (35) in the northern Rockall Trough. Profiles located in figure 12.



0
500
1000
1500
depth/m
NW
1
FOOT of
LOUSY BANK
ROCKALL TROUGH
FENI RIDGE
?
R4
R4
R4
km
0
10
20
30
40
50

Fr

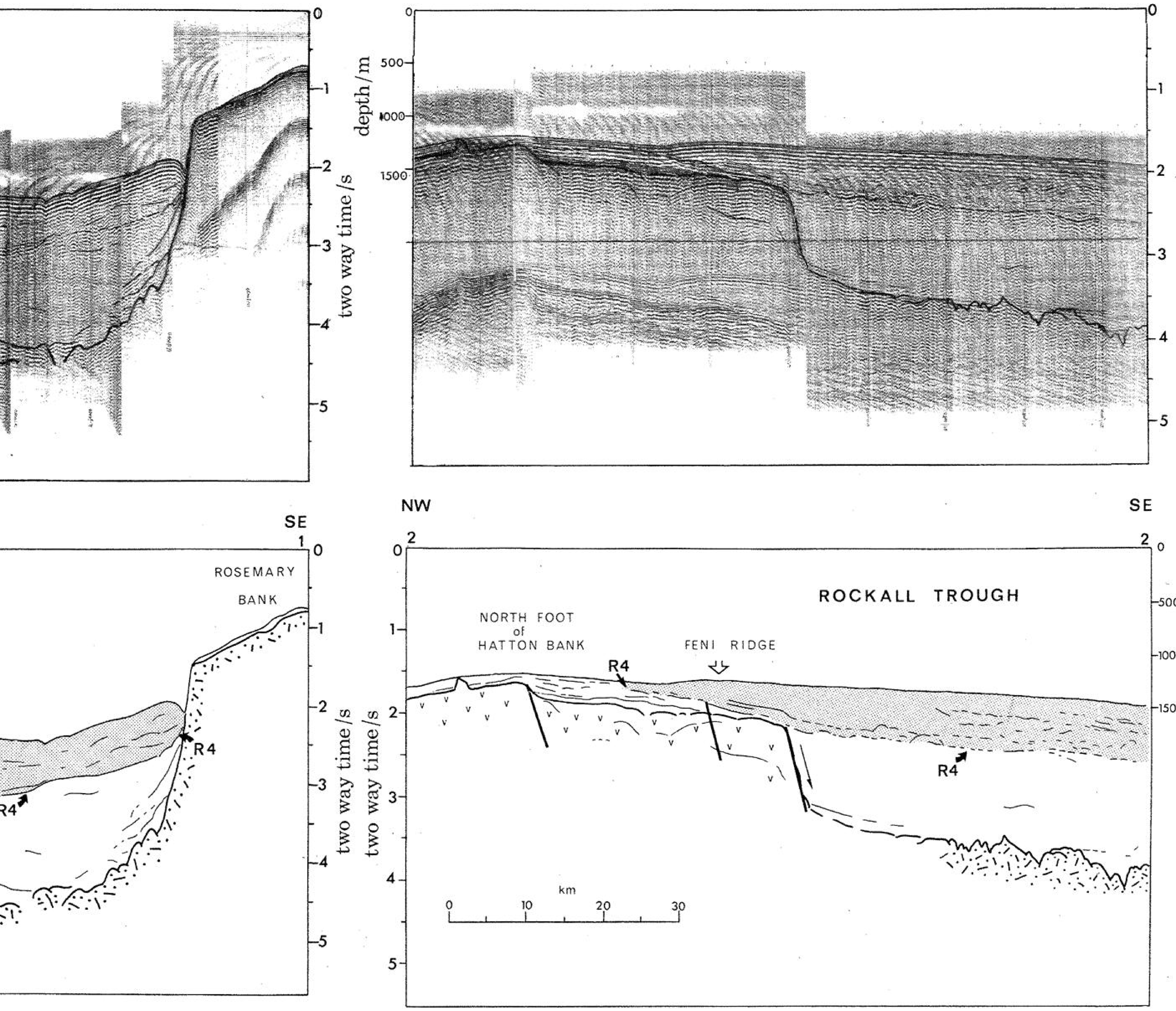
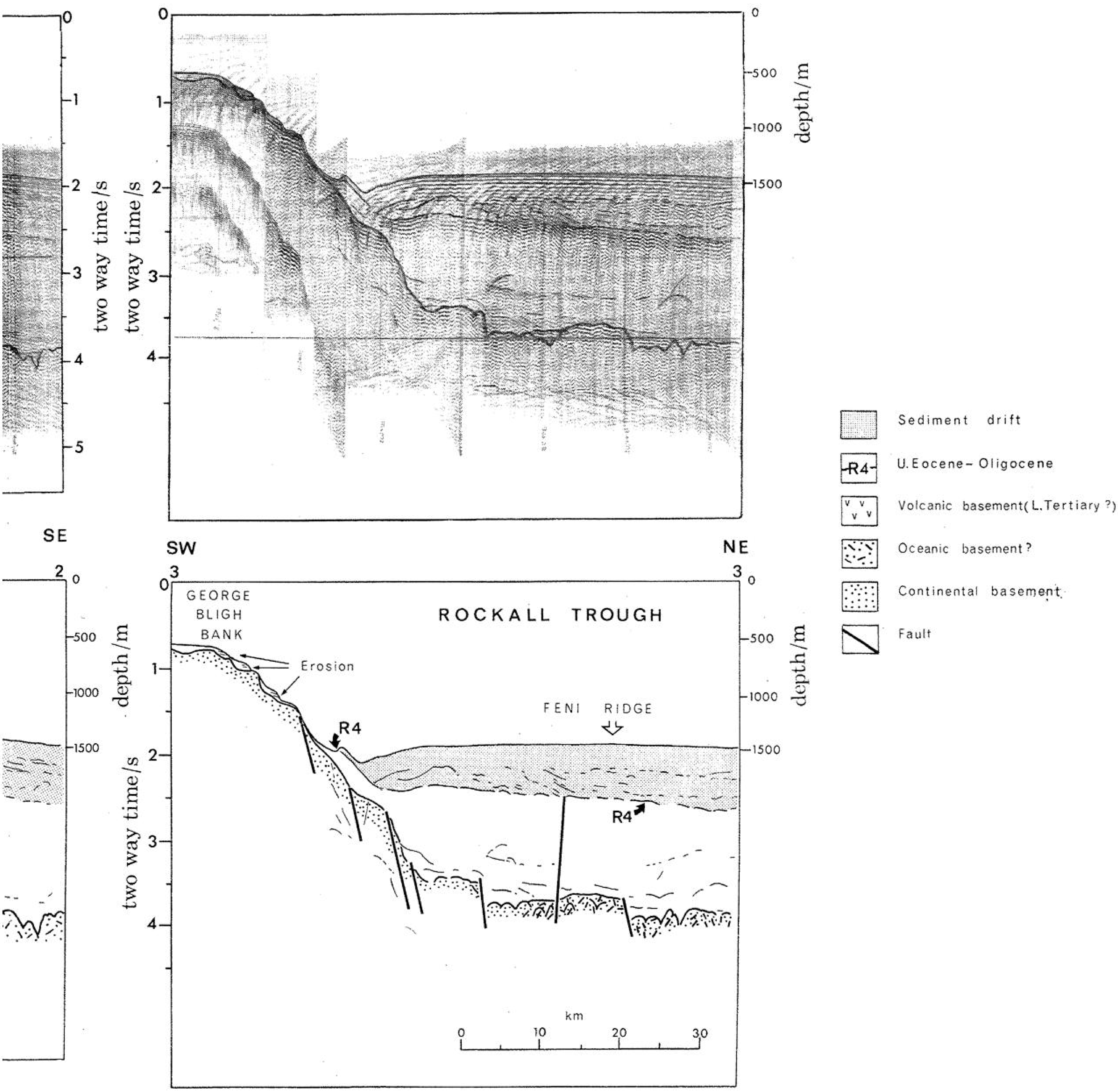


FIGURE 35. Seismic reflexion profiles (35) in the northern Rockall Trough. Profiles located in figure 12.



two way time/s
depth/m
SE
SW
NE
GEORGE
BLIGH
BANK
ROCKALL TROUGH
Erosion
R4
FENI RIDGE
R4
km
0 10 20 30
Sediment drift
R4
U. Eocene – Oligocene
Volcanic basement (L. Tertiary ?)
Oceanic basement?
Continental basement
Fault

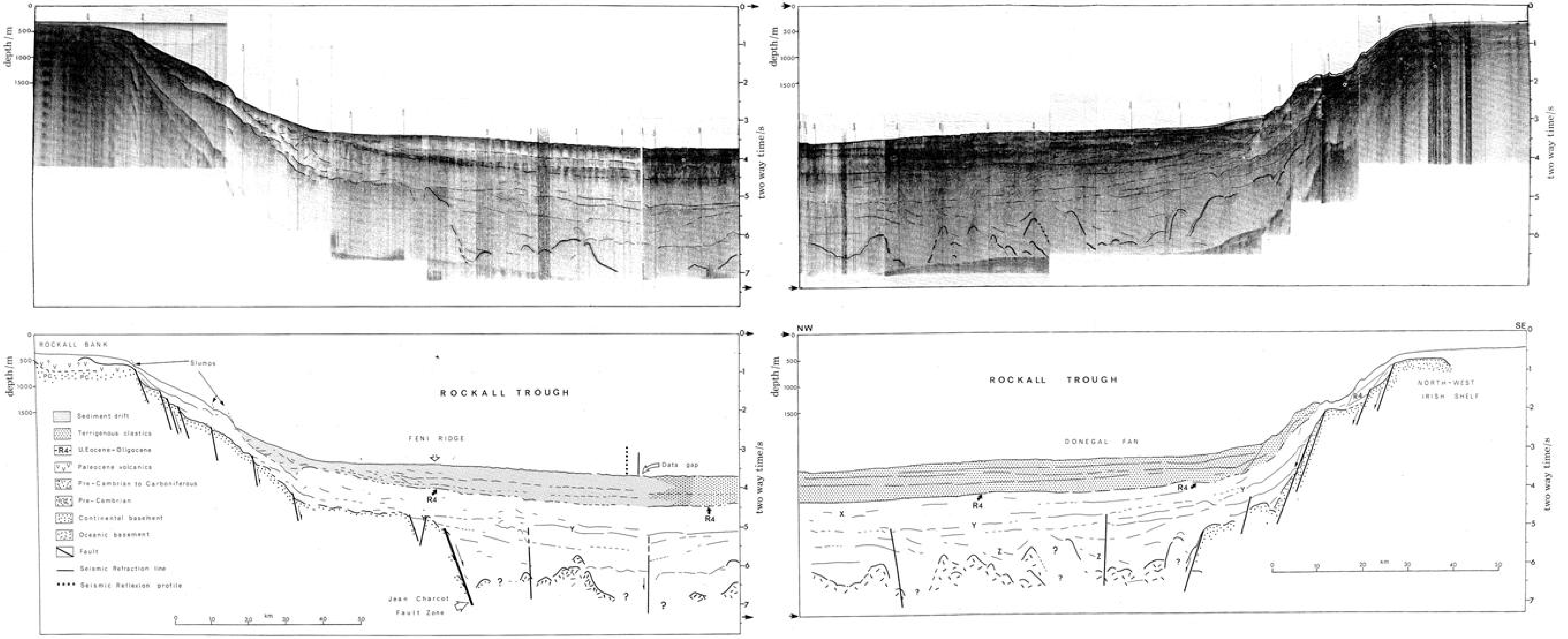


FIGURE 29. Seismic reflexion profile (29) between Rockall Bank and Donegal. Profile is located in figure 12.

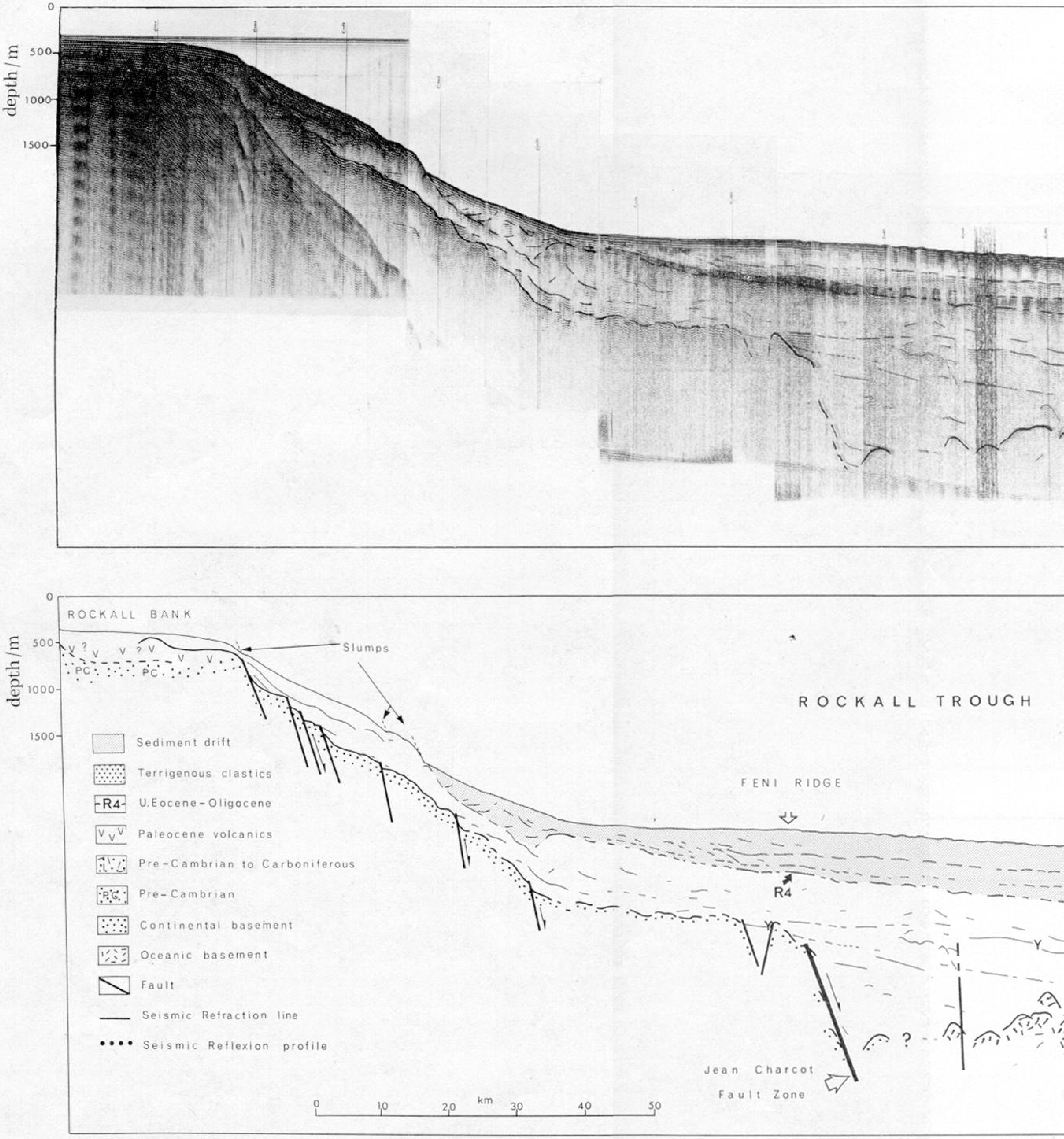
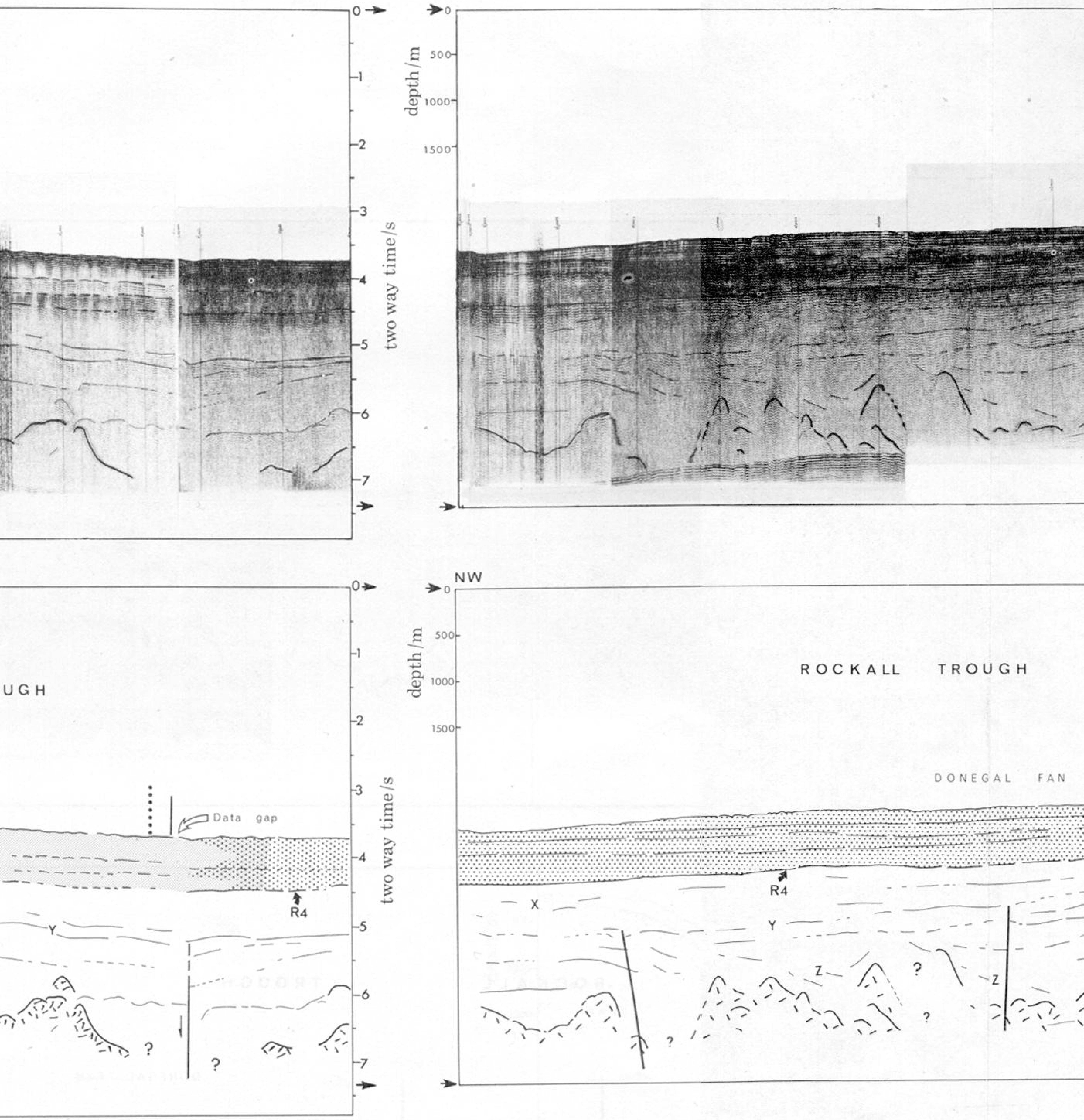
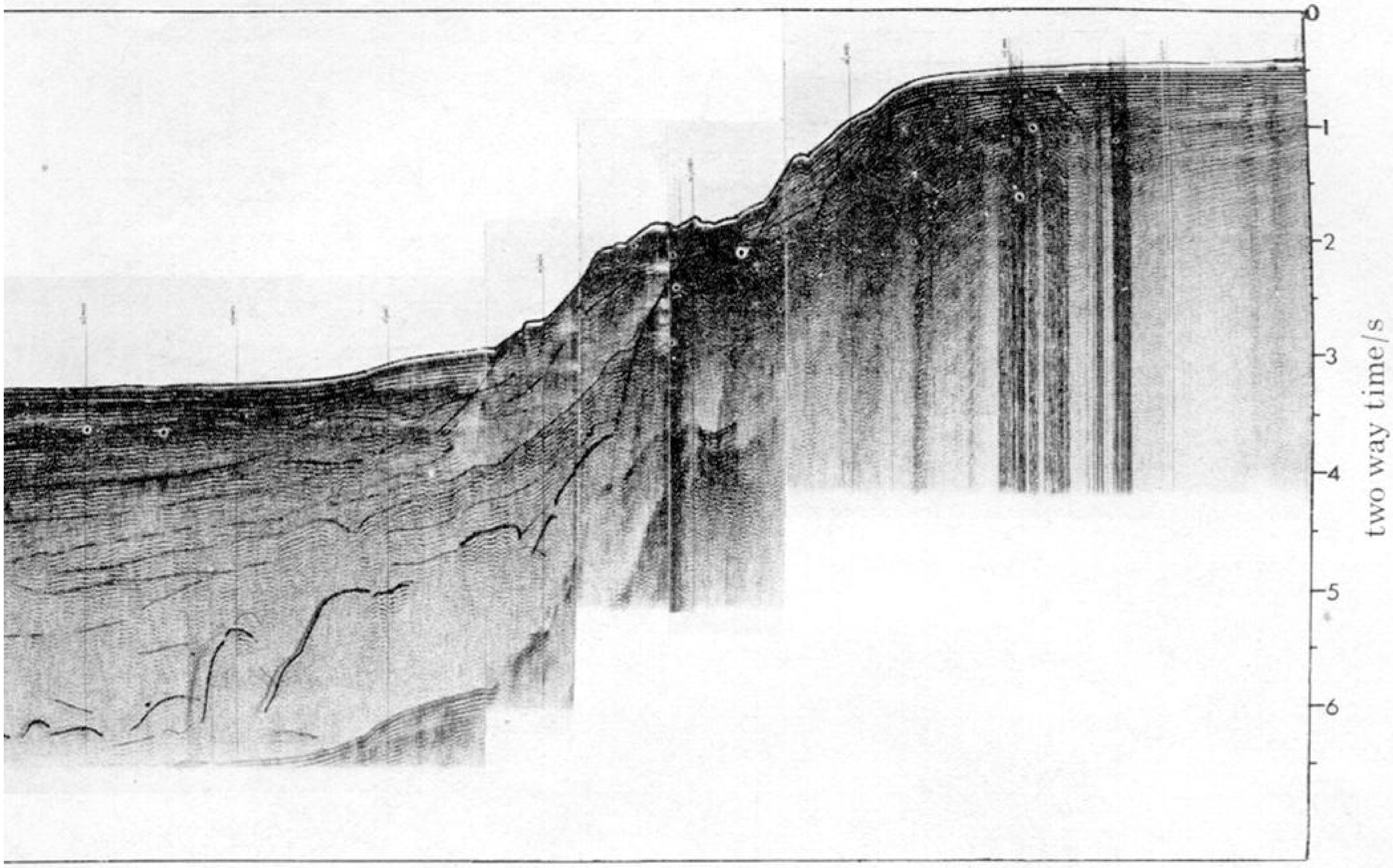

depth/m
0
500
1000
1500
ROCKALL BANK
Slumps
ROCKALL TROUGH
FENI RIDGE
R4
Sediment drift
Terrigenous clastics
U.Eocene-Oligocene
Paleocene volcanics
Pre-Cambrian to Carboniferous
Pre-Cambrian
Continental basement
Oceanic basement
Fault
Seismic Refraction line
Seismic Reflexion profile
Jean Charcot
Fault Zone
km
0
10
20
30
40
50

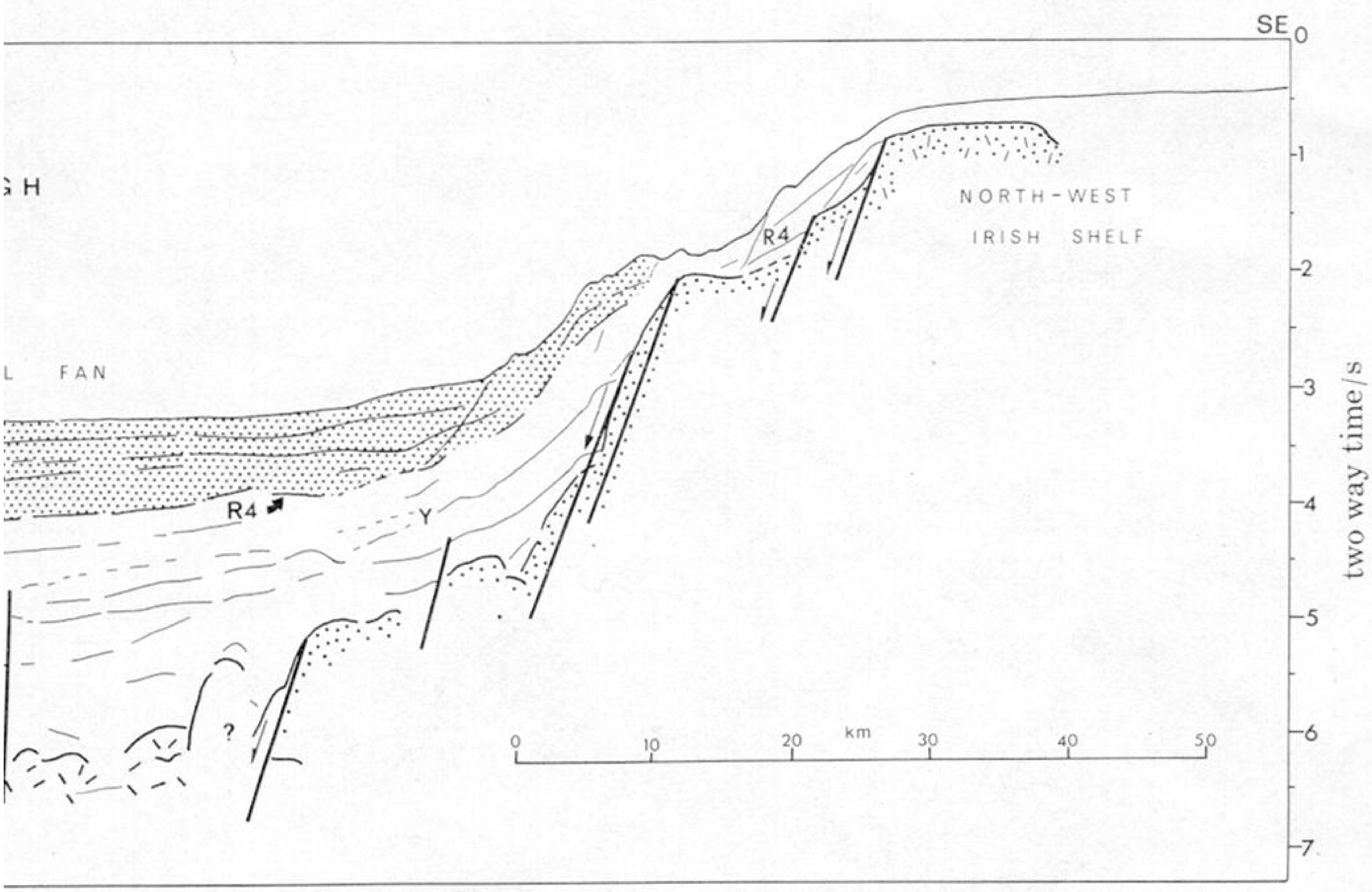
FIGURE 29. S



RE 29. Seismic reflexion profile (29) between Rockall Bank and Donegal. Profile is located in figure 12.



two way time/s



SE
GH
L FAN
NORTH-WEST
IRISH SHELF
R4
R4
Y
?
km
two way time/s

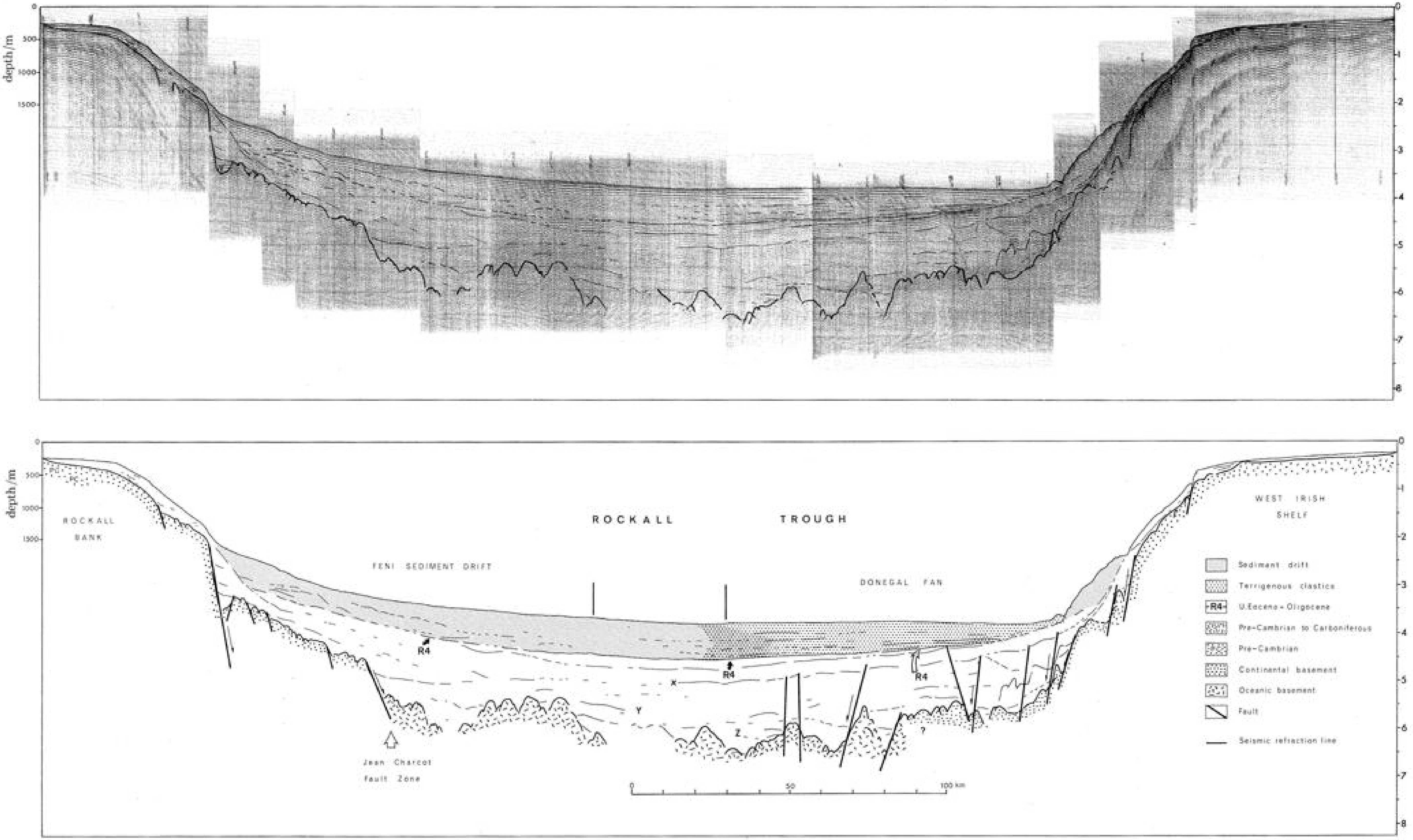
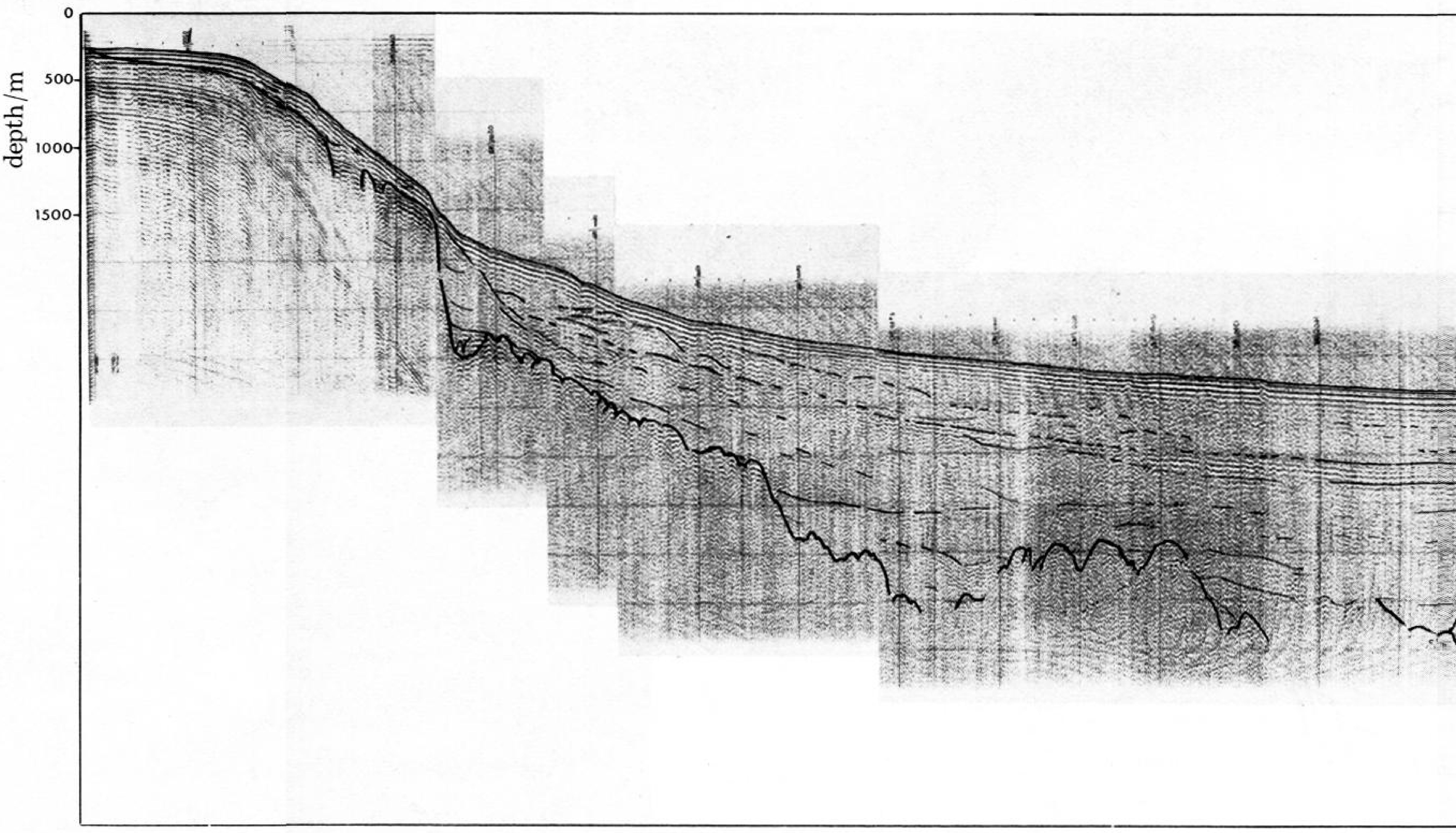
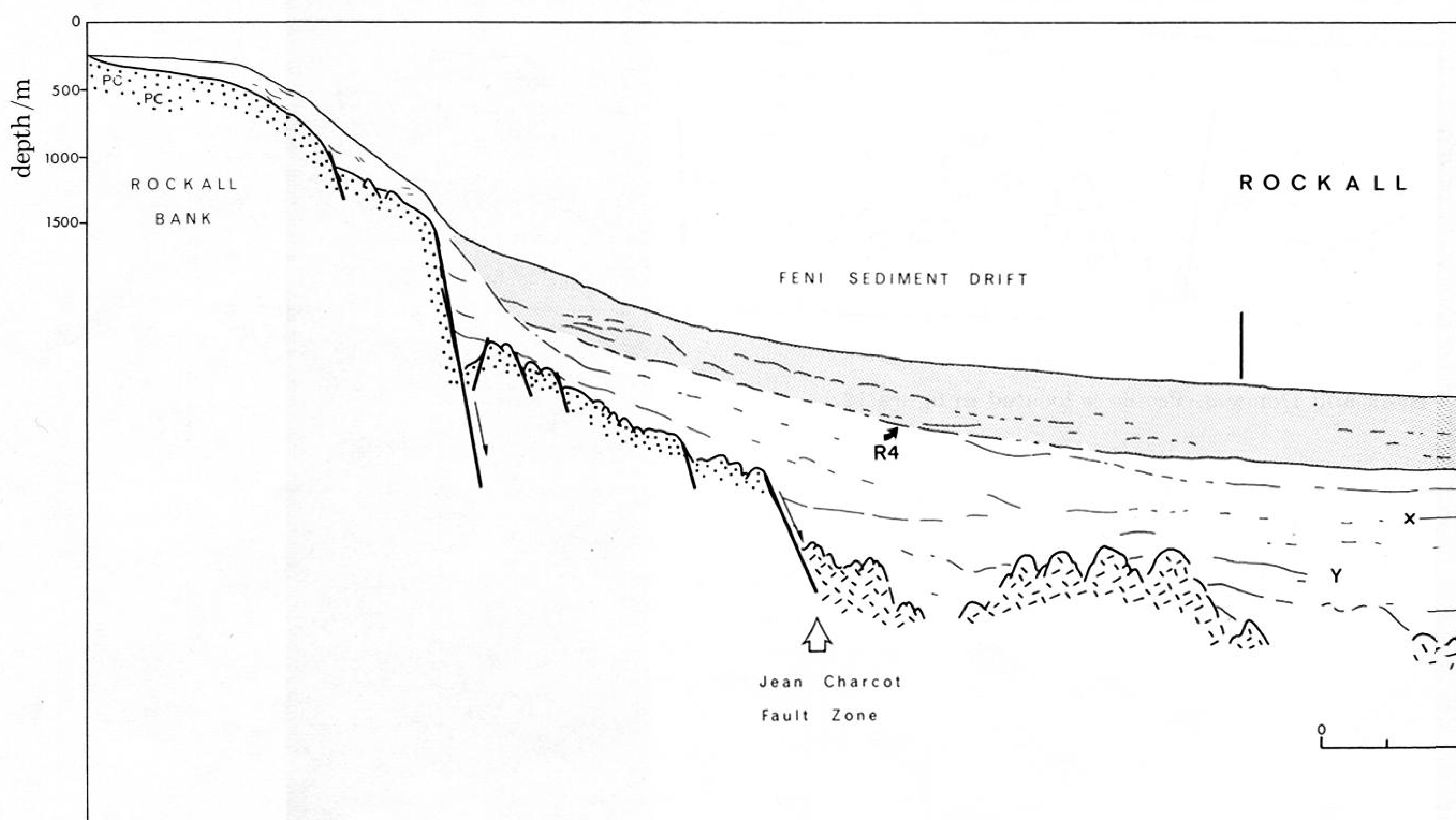
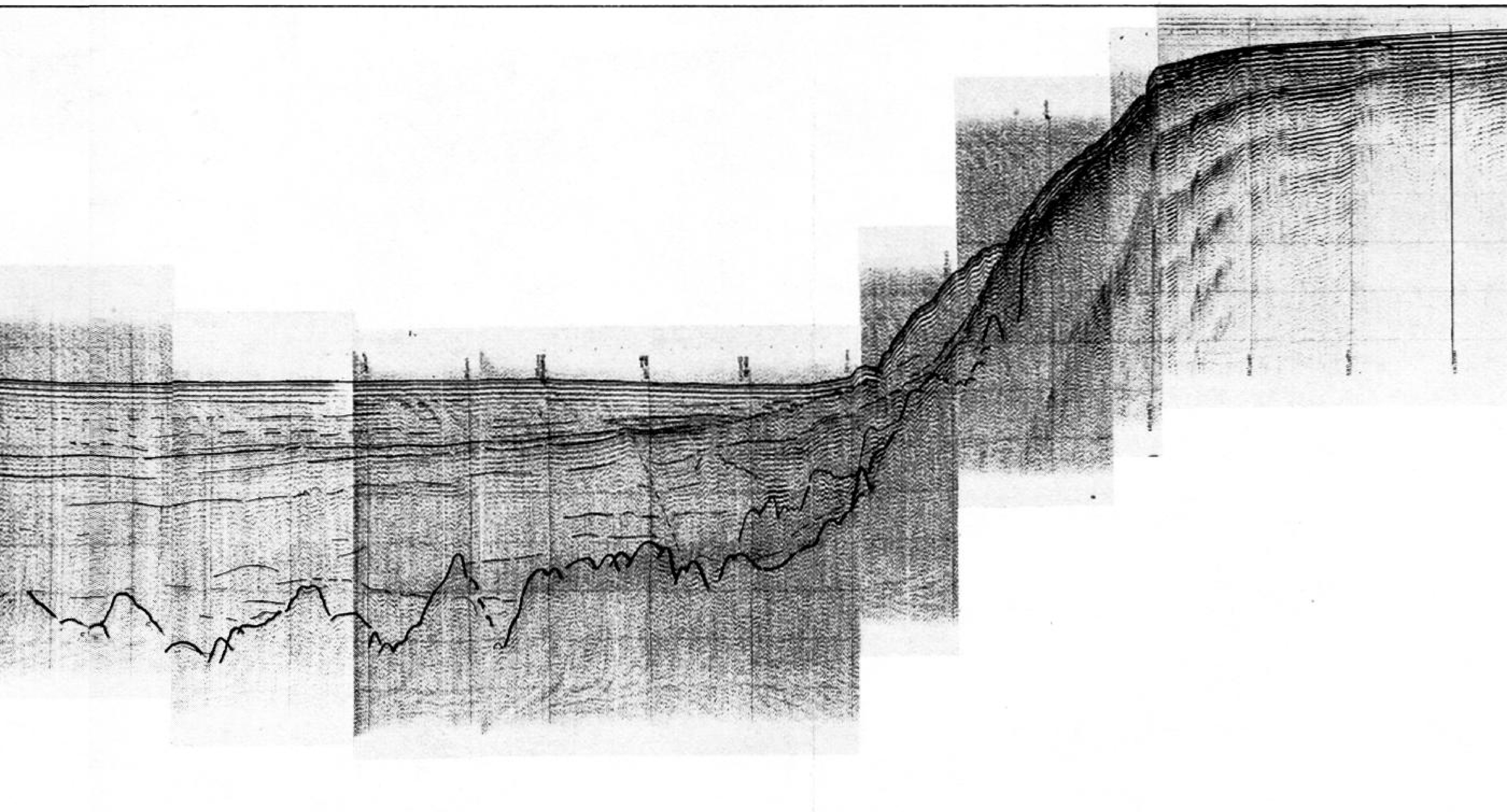
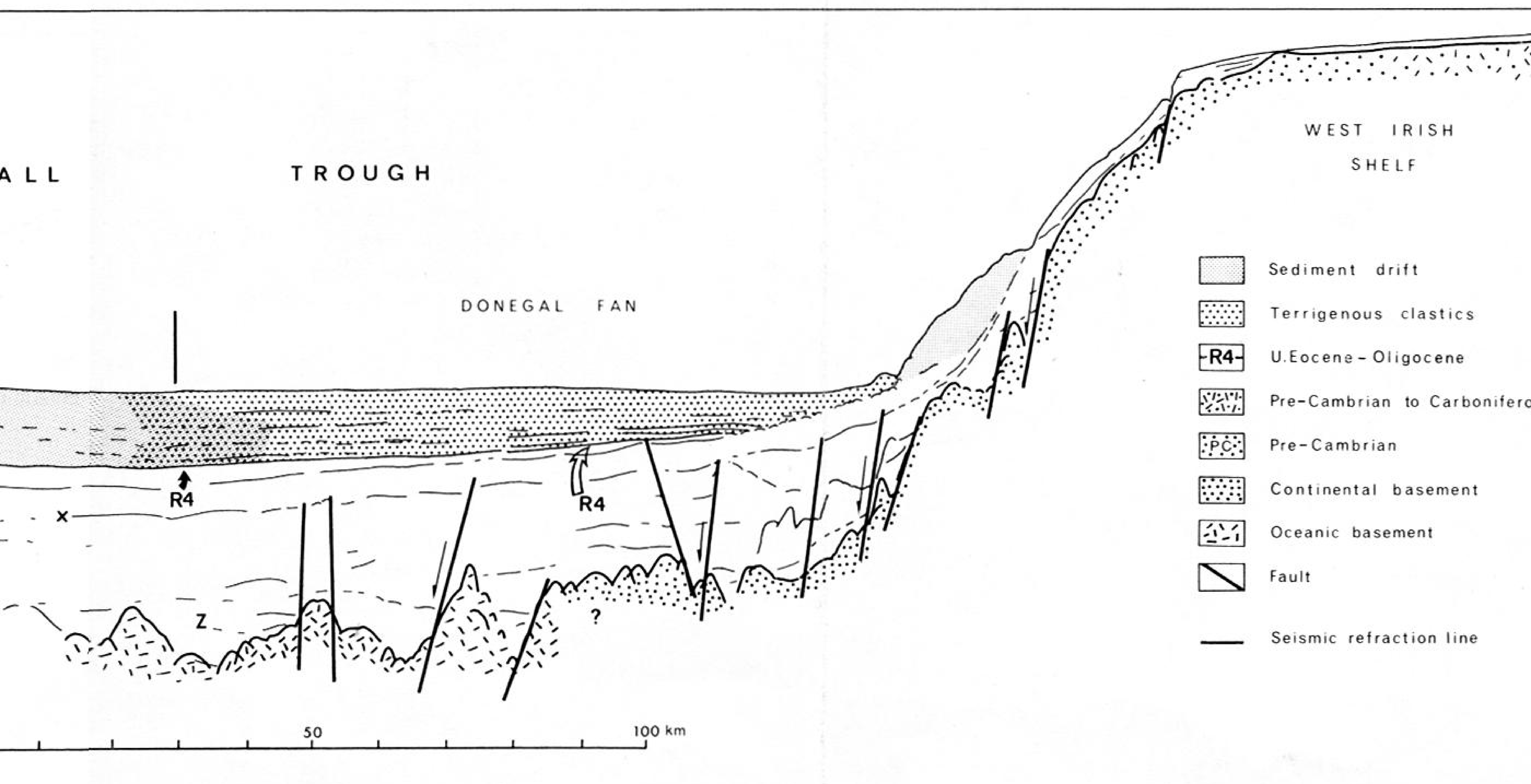


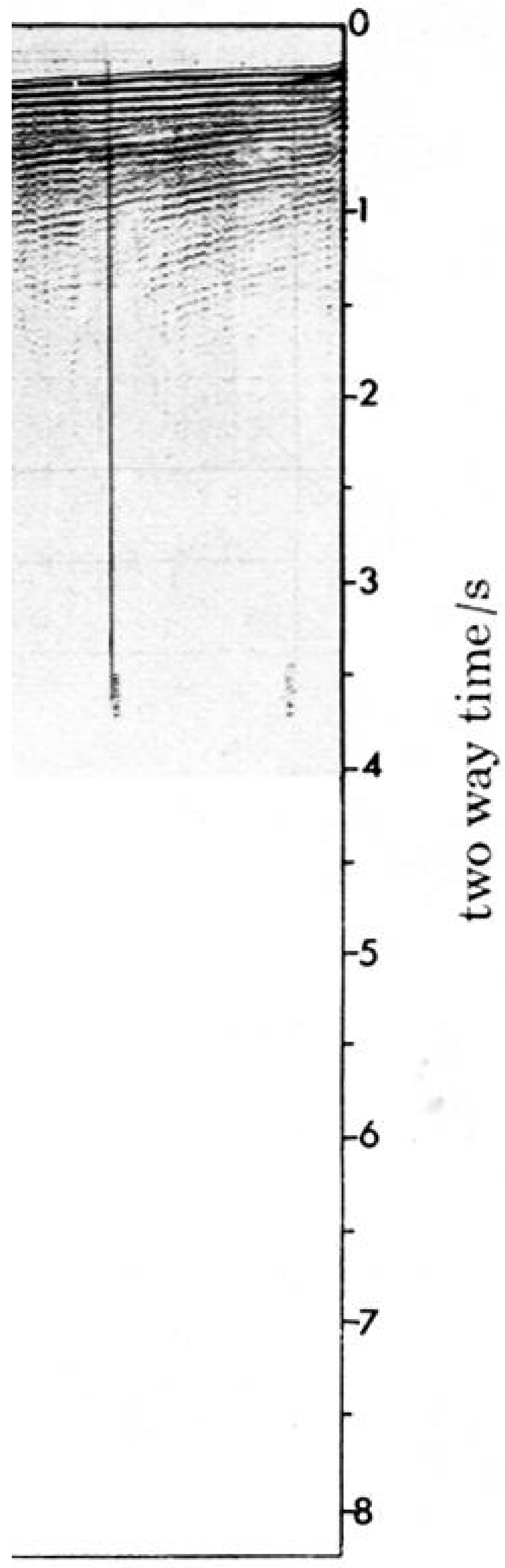
FIGURE 30. Seismic reflexion profile (30) between Rockall Bank and Achill Island. Profile is located in figure 12.

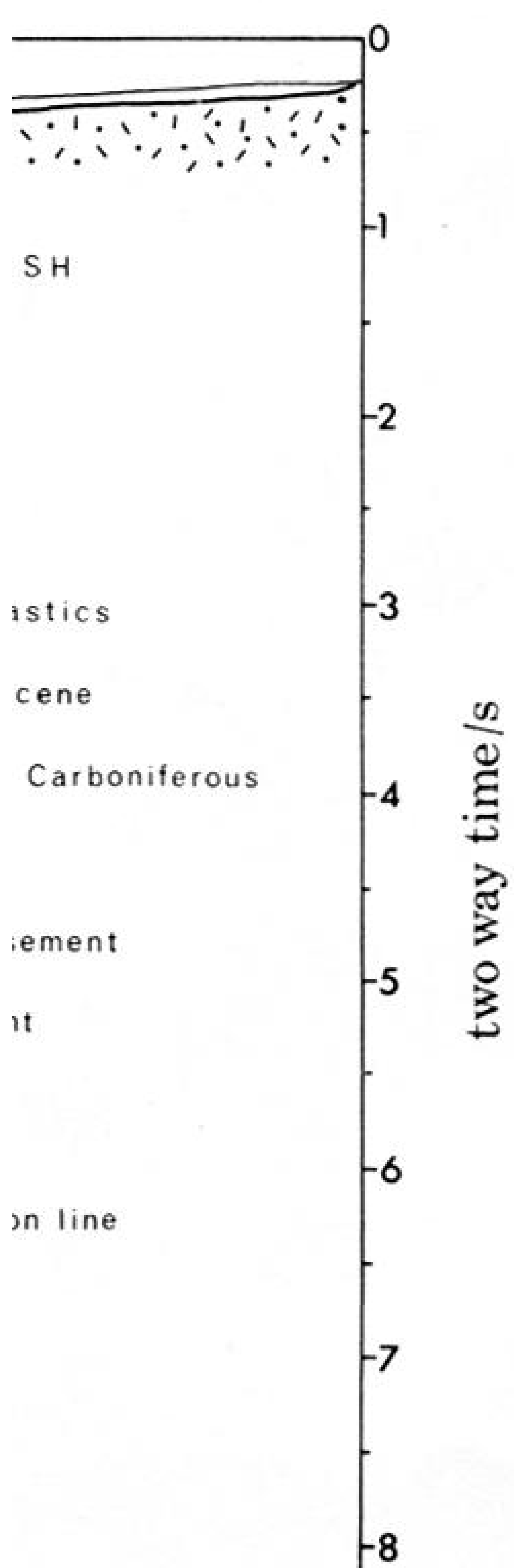

0
500
1000
1500
depth/m


0
500
1000
1500
depth/m
PC
PC
ROCKALL
BANK
ROCKALL
FENI SEDIMENT DRIFT
R4
X
Y
Jean Charcot
Fault Zone
0




ALL
TROUGH
WEST IRISH
SHELF
DONEGAL FAN
R4
R4
X
Z
?
Sediment drift
Terrigenous clastics
R4
U.Eocene - Oligocene
Pre-Cambrian to Carbonifero
PC
Pre-Cambrian
Continental basement
Oceanic basement
Fault
Seismic refraction line
50
100 km


0
1
2
3
4
5
6
7
8
two way time/s


SH
astics
cene
Carboniferous
ement
on line
0
1
2
3
4
5
6
7
8
two way time/s

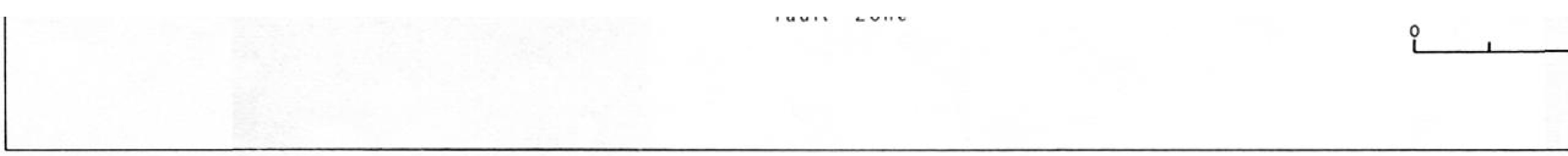
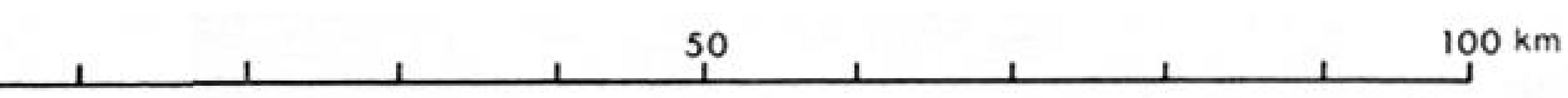
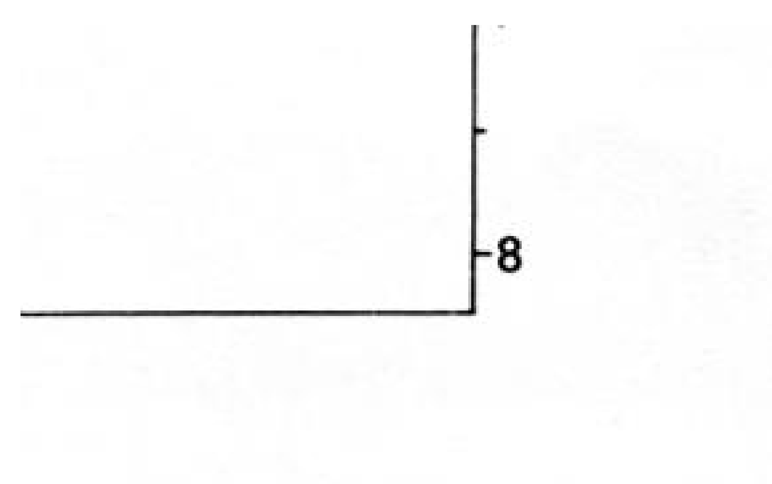
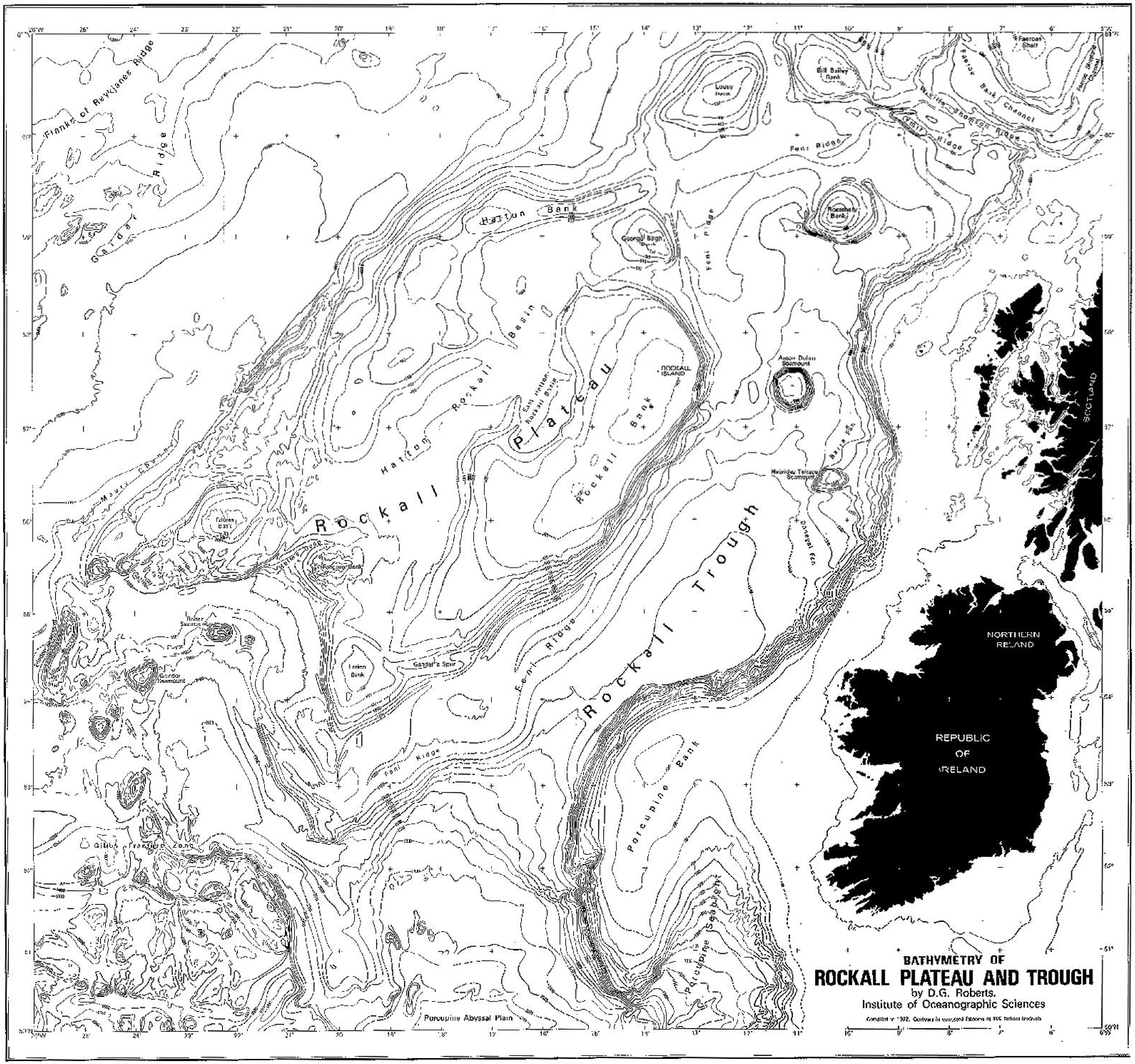


FIGURE 30. Seismic reflexion profile (30) between F



etween Rockall Bank and Achill Island. Profile is located in figure 12.


8


BATHYMETRY OF
ROCKALL PLATEAU AND TROUGH
by D.G. Roberts,
Institute of Oceanographic Sciences
Flanks of Reykjanes Ridge
Gardar Ridge
Hatton Bank
George Bligh Bank
Lousy Bank
Bill Bailey Bank
Faeroes Shelf
Faeroe Bank Channel
Wyville-Thomson Ridge
Feni Ridge
Rosemary Bank
Hatton Rockall Basin
Rockall Plateau
East Hatton Rockall Basin
Rockall Bank
ROCKALL ISLAND
Anton Dohrn Seamount
Hebrides Terrace Seamount
Rockall Trough
Donegal Fan
Maury Channel
Edoras Bank
Fangorn Bank
Lorien Bank
Gandalf's Spur
Porcupine Bank
Porcupine Seabight
Porcupine Abyssal Plain
Gibbs Fracture Zone
SCOTLAND
NORTHERN IRELAND
REPUBLIC OF IRELAND