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Marine Geophysical Investigations in the
Norwegian-Greenland Sea between the Latitudes of
62°N and 74°N

by

Alan Gordon Nunns

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A thesis submitted for the degree of
Doctor of Philosophy at the University of Durham

Van Mildert College

October, 1980

ABSTRACT

The Durham University 1977 marine geophysical cruise (R.R.S. Shackleton 977) in the southern Norwegian-Greenland Sea included a survey of the Jan Mayen block between the latitudes of 67°N and 69.5°N and a traverse of the Norway Basin. Reduced bathymetric, gravity and magnetic data are presented. The gravity and magnetic data are interpreted using direct inversion techniques. Most of the multichannel seismic data is not yet processed, but drawings of the monitor records and a short processed section are included, together with a description of the processing system. In addition, a detailed review is given of published geophysical data from the southern Norwegian-Greenland Sea and a compilation of marine magnetic anomalies is presented.

The principal results are as follows. The Jan Mayen microcontinental fragment is about 90 km wide and extends from about 65°N to 70°N . Contrary to previous inferences, there is little evidence of pre-Tertiary sediment beneath the block. The crust thickens (to about 16-20 km) passing from the Norway Basin to the Jan Mayen block.

A revision of anomaly correlations and identifications in the Norway Basin is presented and the nature of the magnetic source layer is analysed. The central fan shaped anomaly pattern was formed from the end of anomaly 20 time (-43 Ma) to about anomaly 7 time (-27 Ma) as the Jan Mayen block rotated 27.4° anticlockwise relative to Norway about a near pole (64.9°N , 12.3°W). Complementary spreading took place about the newly formed Kolbeinsey axis, such that Jan Mayen rotated 29° anticlockwise relative to Greenland (about a pole at 71.1°N , 16.4°W). Bathymetric and structural evidence is adduced in favour of this scheme and the implications with respect to the structure and the evolution of the Jan Mayen block are discussed.

It is suggested that the anticlockwise reorientation of spreading between Greenland and Norway, which occurred when spreading ceased in the Labrador Sea, caused compressional stresses across a major transform fault at the southern end of the Norway Basin and led to the fission of Jan Mayen from Greenland.

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Chapter 1

INTRODUCTION

The Norwegian-Greenland Sea is the northeastern extension of the Atlantic Ocean beyond Iceland (Figure 1.1). The area investigated in this study stretches across the Norwegian-Greenland Sea from the Greenland shelf to the Norwegian shelf and is bounded to the northeast by the compound Jan Mayen fracture zone and to the southwest by the Greenland-Iceland Ridge, the Icelandic shelf and the Iceland-Faeroe Ridge (Figure 1.2). The term southern Norwegian-Greenland Sea will be used for this area, based on a division of the Norwegian-Greenland Sea into three parts by the major fracture zones (Figure 1.2).

It is generally accepted that most of the southern Norwegian-Greenland Sea was formed during Tertiary times by the processes of sea-floor spreading, but despite intense geophysical investigation for over a decade, a number of important unsolved problems remain. The aim of the present study is to contribute towards an understanding of the structure and evolution of the southern Norwegian-Greenland Sea. It is based partly on a synthesis and reinterpretation of previously published data and partly on new gravity, magnetic and seismic data acquired on board the R.R.S. Shackleton during the 1977 Durham University geophysical cruise (SHACK977, Part II; Jan Mayen Ridge and Norway Basin).

This chapter begins with a review of the spreading history of the North Atlantic and Arctic Oceans, an appreciation of which is critical to an understanding of the evolution of the southern Norwegian-Greenland Sea. The basic hypothesis for the origin of the southern Norwegian-Greenland Sea, which was proposed in 1967 by Johnson and Heezen is then described. This is followed by a more detailed description of the various physiographic provinces of the southern Norwegian-Greenland Sea, in which the major subsequent results are introduced. In the case of the Jan Mayen block in particular, the emphasis is on a critical synthesis of

Figure 1.1 Polar stereographic chart of the North Atlantic and Arctic Oceans redrawn after Phillips et al. (in press). Continental shelf areas are outlined by the 1000 m contour (solid lines) and deeper areas of interest, such as the Rockall Plateau, by the 2000 m contour (dash-dot line). Within the ocean basins the active spreading ridges and major fracture zones are shown as solid lines and extinct spreading axis as broken lines. Ocean-continent boundaries, where they have been defined, are shown dotted. The key magnetic anomalies which are shown are assumed to have the following ages

Anomaly	Age/Ma	Reference
5	10	Hailwood et al. (1979)
13	35	
20	43	
24	52	
28	63	La Brecque et al. (1977)
31	67	
33	74	
34	82	
M0	111	Vogt and Einwich (1979)

Anomalies M0 and 34 bracket the Cretaceous normal polarity period. Crust generated during this period is characterised by a rough magnetic field without recognisable anomalies.

The lineations shown have been transferred by projector from a number of charts. Anomalies younger than 34 for most of the region (excluding the Labrador Sea) are given by Phillips et al. on the same base map, enabling locations to be checked.

South of the Charlie Gibbs Fracture Zone, lineations are mainly from Kristoffersen (1978) with M0 from Sibuet and Ryan (1979) and the northern Biscay ocean-continent boundary from Montadert et al. (1979). Kristoffersen's anomaly 33 is the anomalies 31 of previous workers (for example Williams and McKenzie 1971, Pitman and Talwani 1972). This revision appears to be generally accepted.

In the vicinity of the extinct triple junction at the mouth of the Labrador Sea lineations are from Kristoffersen (1978), Kristoffersen and Talwani (1977) and Srivastava (1978). The magnetic lineations further north in the Labrador Sea and the fracture zones and ocean-continent boundaries throughout the Labrador Sea and Baffin Bay are from Srivastava alone. Srivastava's older identifications are shown bracketed because they differ from those of Kristoffersen (1978) where they overlap. Srivastava's anomalies 32, 31 and 28 correspond to Kristoffersen's anomalies 34, 33 and 31 respectively. The location of the extinct axis in Baffin Bay is from Jackson (1978) based on gravity minimum. The magnetic identifications which exist in the Bay have not been identified.

Magnetic lineations about the Reykjanes Ridge are mainly from Voppel et al. (1979), based on a number of earlier sources. The closely spaced fracture zones shown to the east of the Ridge are from Vogt and Avery (1974). Those to the west are from Voppel and Rudloff (1980). In all probability similar fracture zones occur further to the north. Positions of ocean-continent boundaries between the Icelandic transverse ridge and the Charlie-Gibbs Fracture Zone are from Bott (1978). There is no direct evidence for the location of the extinct axis in the Rockall trough.

In the Norwegian-Greenland Sea, lineations about the Kolbeinsey Ridge are from Vogt et al. (1980) and in the Norway Basin, from the results of the present study. Lineations about the Mohns Ridge are from Phillips et al. (in press) with the ocean-continent boundary of Talwani and Eldholm (1977) which is coincident with the Voring Plateau Escarpment. The oldest anomaly definitely identifiable about the Knipovich Ridge is 5, although Phillips et al. (in press) have identified anomalies as old as 22 west of the Ridge. These identifications are tentative because an eastward axis jump may have occurred (Vogt et al. 1978).

Lineations about the Nansen Ridge in the Eurasia Basin are from Phillips et al. (in press). The position of the ocean-continent boundary is not known in the Eurasia Basin. Anomaly 24 generally lies 50-100 km from the physiographic margins of Eurasia and the Lomonosov Ridge (Vogt et al. 1979), but there is no evidence of older anomalies.

The origin of the Amerasia Basin, which lies between the Lomonosov Ridge and Alaska, and includes the Canada Basin, the Alpha-Mendeleev Ridge and the Toll Basin, is not known, but present evidence (Vogt et al. 1979) suggests that any sea-floor spreading in the Basin ceased by anomaly 34 time.

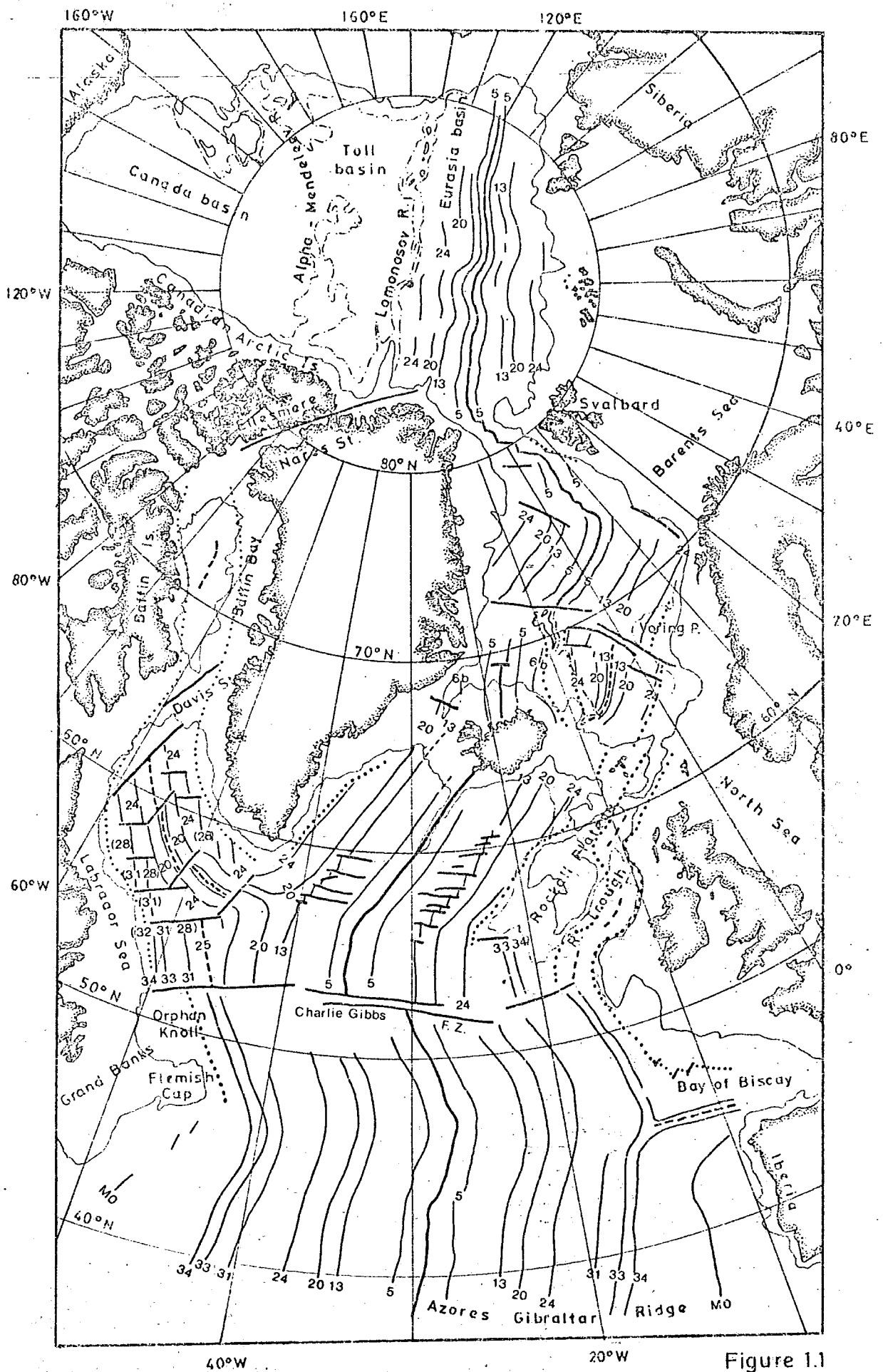


Figure 1.1

Figure 1.2. The Norwegian-Greenland Sea, showing the division into three parts (northern, central, southern) by the sets of major fracture zones. Redrawn, with slight alterations from various charts in the Geophysical Atlas of the Norwegian-Greenland Sea (Gronlie and Talwani 1978).

Bathymetry in nominal fathoms with a 500 m contour interval

Spreading axes, fracture zones and marginal escarpments are shown in bold. The configuration of the Kolbeinsey axis north of 70.5 N and the proposed central Jan Mayen Fracture Zone are based on considerations given in the present study. The Voring Plateau Escarpment (VPE) and the Faeroe-Shetland Escarpment (FSE) are considered by Talwani and Eldholm (1977) to mark the site of initial rifting in the central and southern regions respectively.

The extinct Aegir axis is shown dashed, in bold.

Earthquake epicentres and fault plane solutions, illustrating the present tectonic configuration of the area are originally from Husebye et al. (1975), Sykes (1967), Conant (1972) and Zobin (1972).

In the Norway Basin the two-ship refraction profiles of Ewing and Ewing (1959) and Hinz and Moe (1971) are shown (F9 and II&III respectively).

The locations of Deep Sea Drilling Project Leg 38 sites are indicated (Talwani, Udinstev et al. 1976).

Not included on the original charts, but redrawn from Einarsson (1979) are the outlines of the neovolcanic zones in Iceland.

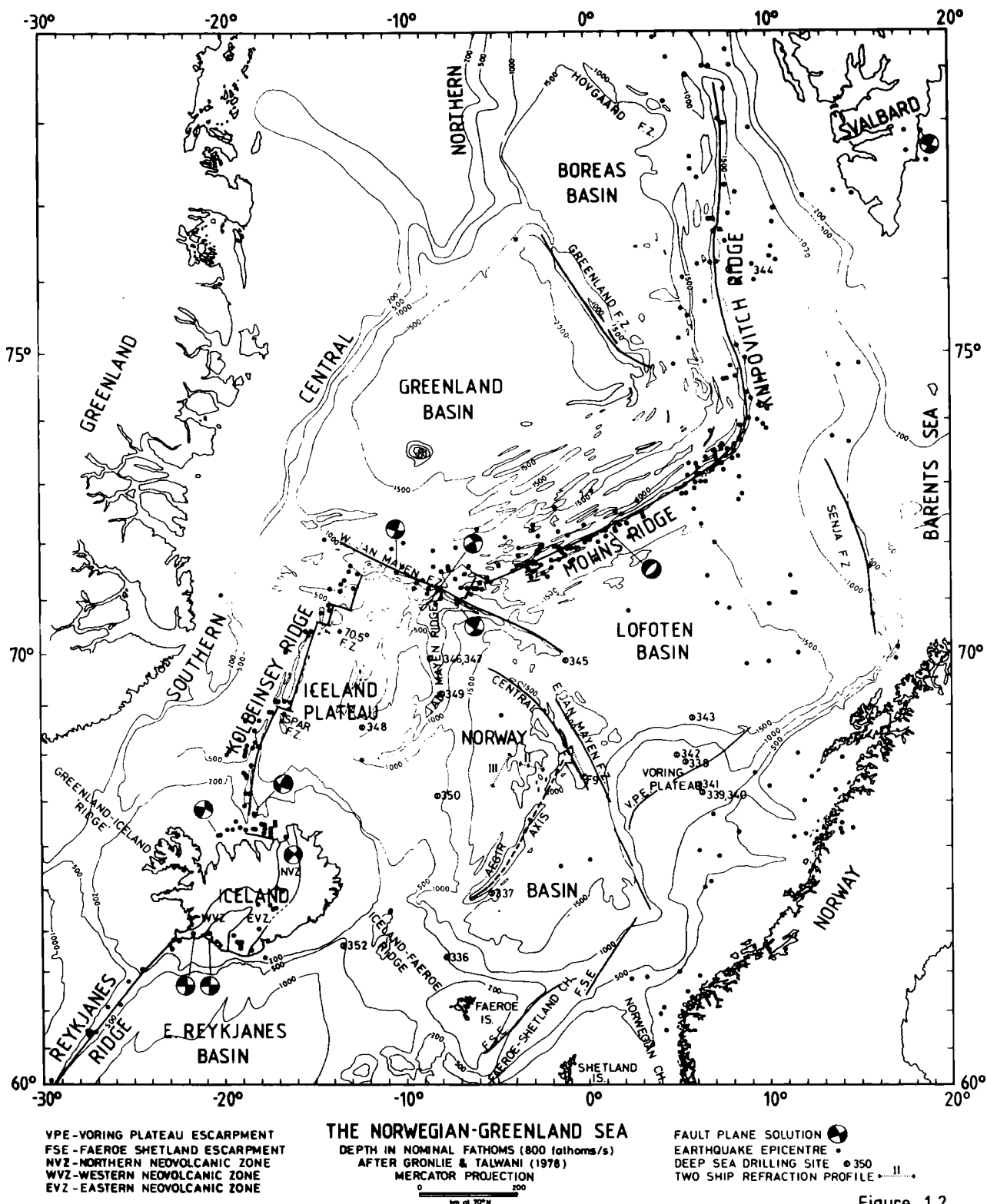


Figure 1.2

recently published geophysical data. In the conclusion to Chapter 1 the outstanding geophysical problems of the region are summarised and a brief description is given of the contents of the remainder of the thesis.

Geographical nomenclature follows Gronlie and Talwani (1979) with the following main exceptions. In accordance with earlier usage (for example Johnson et al. 1972) the names "Kolbeinsey Ridge" and "Aegir Axis" are used instead of the names "Iceland-Jan Mayen Ridge" and "Extinct Axis". The name "Iceland Plateau" is used in the original broad sense of Johnson and Heezen (1967) rather than in the restricted sense of Gronlie and Talwani (1979). Because of the paucity of geographical names in the region, some new names are introduced in this thesis. These have been conservatively adopted from existing names and are underlined where first used. Terms such as "southern Norwegian-Greenland Sea" should be taken to indicate structural rather than oceanographic entities.

Most of the maps in this thesis were based on the Mercator and polar stereographic charts in the "Geophysical Atlas of the Norwegian-Greenland Sea" (Gronlie and Talwani 1978). The Mercator charts were drawn at an original scale of 1:4 000 000 at 63°N and the polar stereographic charts at a scale of 1:3 500 000 at 70°N. When information from other charts at different scales had to be incorporated this was done using a Grant Projector, degree square by degree square to minimise distortion. Where bathymetric contours have been redrawn from Gronlie and Talwani the depths are given in nominal fathoms, based on assumed sound velocity of 800 fathoms s⁻¹. Correction tables and a map showing echo-sounding regions are given in the Geophysical Atlas.

Figure 1.3 shows the Cenozoic timescale which is used in this thesis (Hailwood et al. 1979), which incorporates a revised magnetic polarity timescale based on the results of DSDP Leg 48 magnetostratigraphy (Section 6.2). The polarity timescale differs most radically from the original Heirtzler et al. (1968) timescale over the range of anomalies 7 to 24, for which the revised anomaly ages (A) are given in terms of

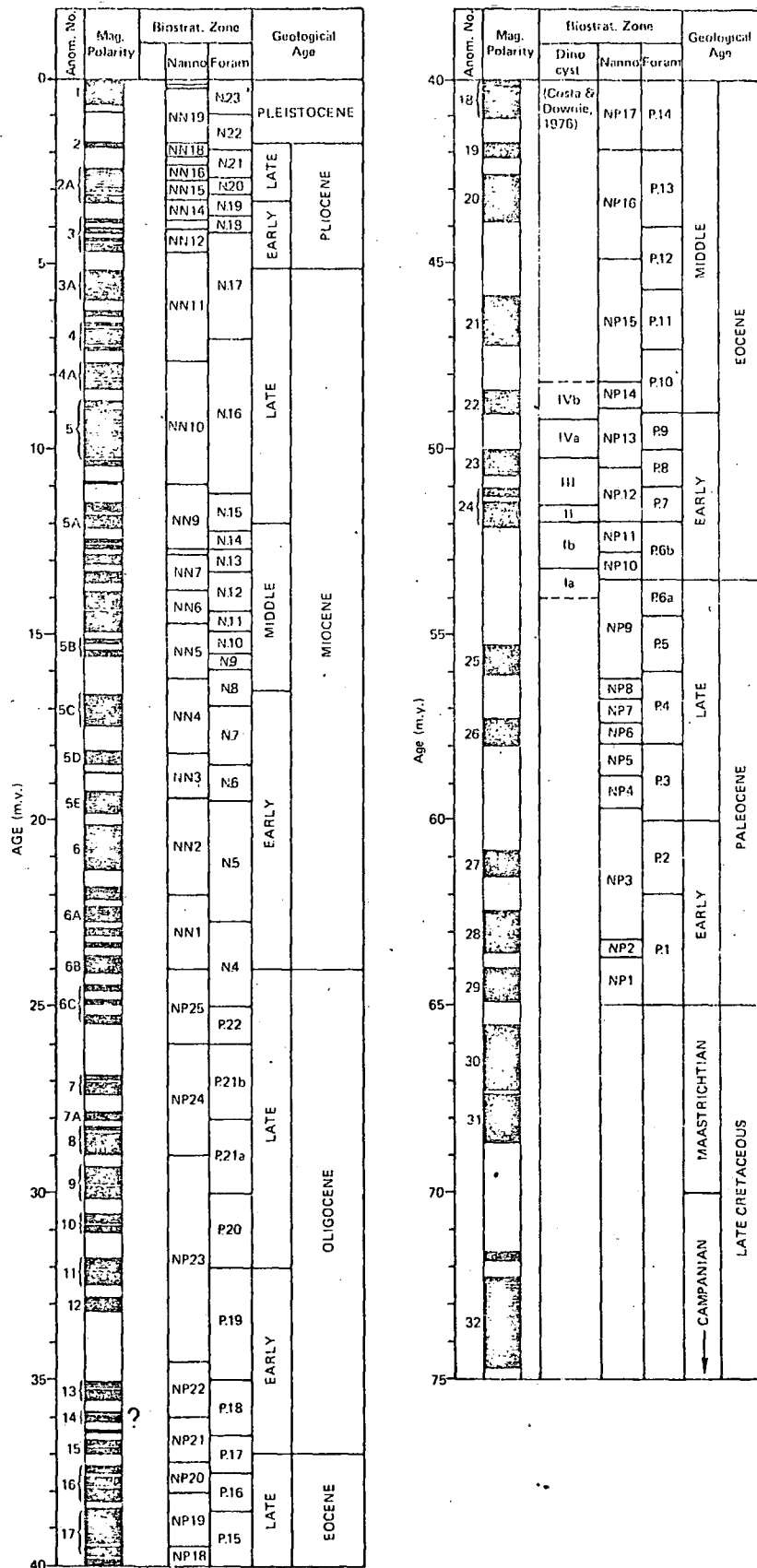


Figure 6. Detailed correlation of revised Cenozoic polarity time scale to biostratigraphic scale of Hardenbol and Berggren (in press).

Figure 1.3. The revised Cenozoic magnetic polarity timescale of Hailwood et al. (1979) based on the results of DSDP Leg 48.

the original ages (H) by the relationship

$$A = 0.73H + 7.47$$

which is expressed in millions of years (Ma). The age of anomaly 24 is revised downwards by about 8 Ma (see section 6.2). Over this interval spreading rates calculated according to the new timescale are 1.37 times larger than those calculated according to the old scale. For the Cretaceous, the polarity timescales of La Brecque et al. (1977) and Vogt and Einwich (1979) are favoured. The ages of some important anomalies are shown in the caption to Figure 1.1. It is also felt that the Cenozoic polarity timescale of Hailwood et al. (1979) should be modified according to the suggestion of La Brecque et al. (1977) to eliminate anomaly 14 (Section 6.1).

1.1 Plate tectonic evolution of the North Atlantic and Arctic Oceans

The following account is based mainly on the references cited in the caption to Figure 1.1, which shows important features of the North Atlantic and Arctic Ocean basins, such as spreading axes, major fracture zones, key magnetic lineations and ocean-continent boundaries. The ideas in published syntheses such as given by Laughton (1975) are also incorporated.

The North Atlantic south of the Azores (Tucholke, Vogt et al. 1979) began to open about 190-170 Ma ago in the early Jurassic. Initial separation was between North America and the bulge of Africa with sinistral shear between North Africa and Europe (see, for illustration, the paleocontinental maps of Smith and Briden 1977).

Prior to Cretaceous times the Atlantic north of the Azores was closed, with the possible exception that Iberia and the Grand Banks of Newfoundland may have separated by about 100 km in the early to middle Jurassic during the period of generation of the Jurassic smooth zone in the central Atlantic (Sibuet and Ryan 1979). The main opening dates from the late Aptian at the beginning of the Cretaceous normal polarity period.

During the early stages of opening the Iberian peninsula separated from North America and also rotated anticlockwise relative to Europe, forming the Bay of Biscay, as if sheared by the easterly moving African plate. A triple junction formed at the mouth of the Bay of Biscay as spreading extended northwestward, separating the Irish shelf from Orphan Knoll-Flemish Cap. Opening may have also occurred in the Rockall trough at this time (Bott 1978). Alternatively Rockall trough may have opened along with Grand Banks-Iberia during an earlier phase of spreading, with transform motion between the Irish shelf and Orphan Knoll-Flemish Cap. All that is known for certain is that spreading in the trough must have ceased by anomaly 34 time, as this anomaly cuts across the mouth of the trough, without evidence of a triple junction.

Just prior to anomaly 34 time, spreading commenced between the southern Rockall Plateau and Labrador, with the formation of the Charlie Gibbs fracture zone. Shortly afterwards, spreading ceased in the Bay of Biscay. The northwesterly rift propagated into the Labrador Sea, with spreading in the northern Labrador Sea by anomaly 31 time (adopting the identification scheme of Kristoffersen (1978) rather than that of Srivastava (1978)).

Up until just before anomaly 24 time the Eurasia Basin, the Norwegian-Greenland Sea and the Reykjanes Basin remained closed and Greenland was effectively part of the Eurasia plate, which was moving in an easterly direction away from North America. Kristoffersen (1978) considers that the pole of opening lay near the Mendeleev Ridge in the Arctic Ocean while Srivastava (1978) considers that it was situated near the northwestern part of Baffin Island (for convenience pole positions are given roughly in terms of present day geography, choosing the more westerly reference frame, that is, North America in this case). Similar motion south of Davis Strait would result from either of these poles. However, assuming that the plate boundary between Greenland and America followed Baffin Bay and Nares Strait, then Kristoffersen's pole predicts opening in Baffin Bay and sinistral slip along Nares Strait while Srivastava's pole

predicts a small amount of sinistral slip in Baffin Bay and considerable compression across Nares Strait.

Immediately prior to anomaly 24 time the spreading pattern changed dramatically with the initiation of spreading between Greenland and Eurasia. Spreading began on the Reykjanes Ridge, the Aegir axis in the Norway Basin and on the Mohns and Knipovich Ridges further north. The motion of Eurasia relative to Greenland after anomaly 24 time was parallel to the Greenland, Senja and eastern Jan Mayen fracture zones, in a southeasterly direction. The northeast margin of Greenland and the margin west of Svalbard are parallel to this trend, and it is likely that the early motion between Greenland and Svalbard was dextral slip. Further north, spreading also began in the Eurasia Basin, where the Lomonosov Ridge was detached from the Eurasia margin.

The direction of relative motion between North America and Eurasia did not alter appreciably at anomaly 24 time as can be seen from the straight trend of the western extension of the Charlie Gibbs fracture zone. A triple junction came into existence south of Greenland and spreading continued in the Labrador Sea, although the direction of spreading changed in a manner consistent with Greenland's newly acquired component of northerly motion relative to Eurasia and North America and a new set of north-northeasterly fracture zones formed. According to Kristoffersen (1978) the pole of rotation between Greenland and North America was situated on the western side of Hudson Bay (56°N , 98°W), giving rise to very oblique spreading in Baffin Bay and compression across Nares Strait. On the other hand Srivastava (1978) considers that it was situated south of Spain, giving opening in Baffin Bay parallel to the Davis Strait lineament and sinistral slip along Nares Strait. The solutions of Kristoffersen and Srivastava are perfectly conjugate with respect to motion along the Nares Strait and in Baffin Bay, before and after anomaly 24 time. Unfortunately the age of Baffin Bay is not independently known.

The next major event in the plate tectonic evolution of

the Atlantic north of the Azores occurred some time between anomaly 20 and anomaly 13 time when spreading in the Labrador Sea stopped. Once again the direction of relative motion between Eurasia and North America remained approximately constant. In consequence of the fact that Greenland became fixed to the North American plate when spreading ceased in the Labrador Sea, the direction of spreading between Greenland and Eurasia changed to become more nearly east-west. South of present day Iceland closely spaced fracture zones formed, striking about 105° (Vogt and Avery 1974). In the Norway Basin spreading continued, but in a confused manner. Further north the Greenland and Senja fracture zones became inactive and spreading probably began between Greenland and Svalbard.

At about anomaly 7 time the closely spaced fracture zones in the Reykjanes Ridge disappeared, although the direction of spreading did not change, so that the Reykjanes Ridge became an oblique spreading axis, which it has remained to the present. At about the same time the spreading in the Norway Basin ceased and the spreading axis in the southern Norwegian-Greenland Sea shifted westward to the Kolbeinsey Ridge, forming the western Jan Mayen fracture zone and possibly splitting off a microcontinental fragment (the Jan Mayen Ridge) from the Greenland margin in the process. Iceland in its present form, probably also began to form at about this time.

Since about anomaly 5 time spreading north of the Azores has proceeded in a relatively uncomplicated manner about the present axes.

1.2 The hypothesis of Johnson and Heezen for the origin of the southern Norwegian-Greenland Sea

Although some indication has already been given of the plate tectonic evolution of the southern Norwegian-Greenland Sea, it is convenient to begin a more detailed description with a review of the work of Johnson and Heezen (1967), who were the first authors to consider the geophysical data from the region in the light of the sea floor spreading

hypothesis.

The conclusions of Johnson and Heezen (1967) were based on a relatively small amount of information; namely bathymetric data from a variety of sources, geophysical data from the cruises of U.S.C.G.C. Spar (1966) and R.V. Vema (1966) and the study of Arctic seismicity by Sykes (1965). Nevertheless they have proved substantially correct in the light of later researches. The Kolbeinsey Ridge was identified as the present spreading axis in the southern Norwegian-Greenland Sea, on the evidence of morphology, the distribution of earthquake epicentres and the presence of a positive axial magnetic anomaly. Similarly the Mohns Ridge was identified as the spreading axis in the central Norwegian-Greenland Sea. The Kolbeinsey Ridge is asymmetrically situated with respect to Greenland and Norway, being much closer to Greenland and is offset from the symmetrically situated Mohns Ridge by the seismically active section of the Jan Mayen fracture zone.

The Kolbeinsey Ridge rises out of the western part of the Iceland Plateau, which Johnson and Heezen defined as the entire western half of the southern Norwegian-Greenland Sea (west of about 8°W); an area which is considerably elevated with respect to the Norway Basin to the east and the Greenland Basin to the north. Along the eastern border of the plateau lies the flat-topped Jan Mayen Ridge, which is continuous southward from the Tertiary volcanic island of Jan Mayen to about 68.5°N , being replaced further south by isolated peaks and ridges. A seismic profile across the Jan Mayen Ridge at 69°N showed a thin blanket of horizontal sediments overlying truncated eastward dipping layers. The Jan Mayen Ridge and its southern extension are separated from the Norway Basin by the Jan Mayen rise which has the typical morphology of a continental rise, despite being situated in mid-ocean. Johnson and Heezen considered this "unique rise" to be intermediate in development between the broad, mature Norwegian rise and the Greenland rise, which in the southern Norwegian-Greenland Sea has a narrow juvenile appearance.

The Norway Basin, which lies at a similar depth to the

deep basins flanking the Mohns Ridge, is characterised by "abyssal plains interfingering between sea-mounts and flat intermontane basins" and by sediment thicknesses which are generally greater than those of the central Iceland Plateau.

Johnson and Heezen satisfactorily explained all of these observations (with the exception of the anomalous elevation of the Iceland Plateau) with the following hypothesis. In the central Norwegian-Greenland Sea, spreading has always taken place about the same axis, represented by the symmetrically situated Mohns Ridge. In contrast, in the southern Norwegian-Greenland Sea spreading began in the Norway Basin and the axis later shifted to the Iceland Plateau. During the first phase of spreading the Jan Mayen Ridge was part of the Greenland continental shelf and the Jan Mayen rise was indeed a normal continental rise. When the locus of spreading shifted west (probably in mid-Tertiary times) the Jan Mayen Ridge was detached. In the process the Ridge was uplifted and eroded, but as it drifted further from Greenland it subsided, leading to the deposition of the horizontally stratified sedimentary cap.

Subsequent work

The following authors have dealt specifically with the southern Norwegian-Greenland Sea since 1967 and prior to 1977;

- 1968 Avery, Burton and Heirtzler (magnetic anomalies over the Norway Basin)
- 1970 Vogt, Ostenso and Johnson (evolution)
- 1971 Hinz and Moe (two ship seismic refraction study in the Norway Basin)
- 1972 Johnson, Southall, Young and Vogt (magnetic anomalies over the Iceland Plateau)
- Meyer, Voppel, Fleischer, Closs and Gerke (detailed magnetic anomaly study of the Iceland Plateau)
- Talwani and Eldholm (the Norwegian margin)
- 1973 Eldholm and Talwani (the Jan Mayen Ridge)
- Johnson and Vogt (structure and evolution)
- 1974 Eldholm and Windisch (sediment distribution)
- Talwani (R.V.Vema collated data)

- Talwani and Eldholm (continental margins)
- 1975 Hinz (description of surveys)
- Husebye, Gjoystdal, Bungum and Eldholm (seismicity)
- Johnson (the Jan Mayen Ridge)
- Johnson, Egloff and McMillan (the Greenland margin)
- 1976 Johnson and Campsie (the western Jan Mayen fracture zone)
- Mirlin and Melikhov (magnetic anomalies over the Iceland Plateau)
- Talwani, Udinstev et al. (DSDP Leg 38 results)

In 1977, Talwani and Eldholm published an evolutionary synthesis for the entire Norwegian-Greenland Sea. This was based on many of the publications cited above, on data collected aboard the R.V. Vema (1966 to 1973) and on the results of Leg 38 of the Deep Sea Drilling Project (Talwani, Udinstev et al. 1976, 1979). Talwani and Eldholm's synthesis confirmed and extended the hypothesis of Johnson and Heezen (1967) for the origin of the southern Norwegian-Greenland Sea. A concordant view of the evolution of the region has been presented in a number of related publications (including the Geophysical Atlas) by workers from Lamont-Doherty Geological Observatory and the University of Oslo, for example Eldholm (1979), Gronlie (1979), Gronlie, Chapman and Talwani (1979), Gronlie and Talwani (1978, 1979) and Talwani (1978). Damuth (1978) has summarised echo character data from R.V. Vema cruises.

Prior to 1978, published material dealt with single channel seismic reflection data only. Since then, the results of a number of multichannel surveys have been published, adding greatly to the knowledge of the Greenland margin and the Jan Mayen Ridge (Gairaud, Jacquart, Aubertin and Beuzart, 1978; Hinz and Schluter, 1979a,b; Sundvor, Gidskehaug, Myhre and Eldholm, 1979).

Detailed magnetic surveys have been reported by Larsen (1978), Navrestad and Jorgensen (1979) and Vogt, Johnson and Kristjansson (1980). Multichannel seismic and magnetic data south of the Jan Mayen Ridge have been presented by Garde (1978, Ph.D. dissertation, University of Clausthal) but this

work has not been seen by the present author.

Bungum and Husebye (1977) have studied the seismicity of the Jan Mayen fracture zone in detail. Evans and Sacks (1979) have investigated surface wave dispersion across the Iceland Plateau.

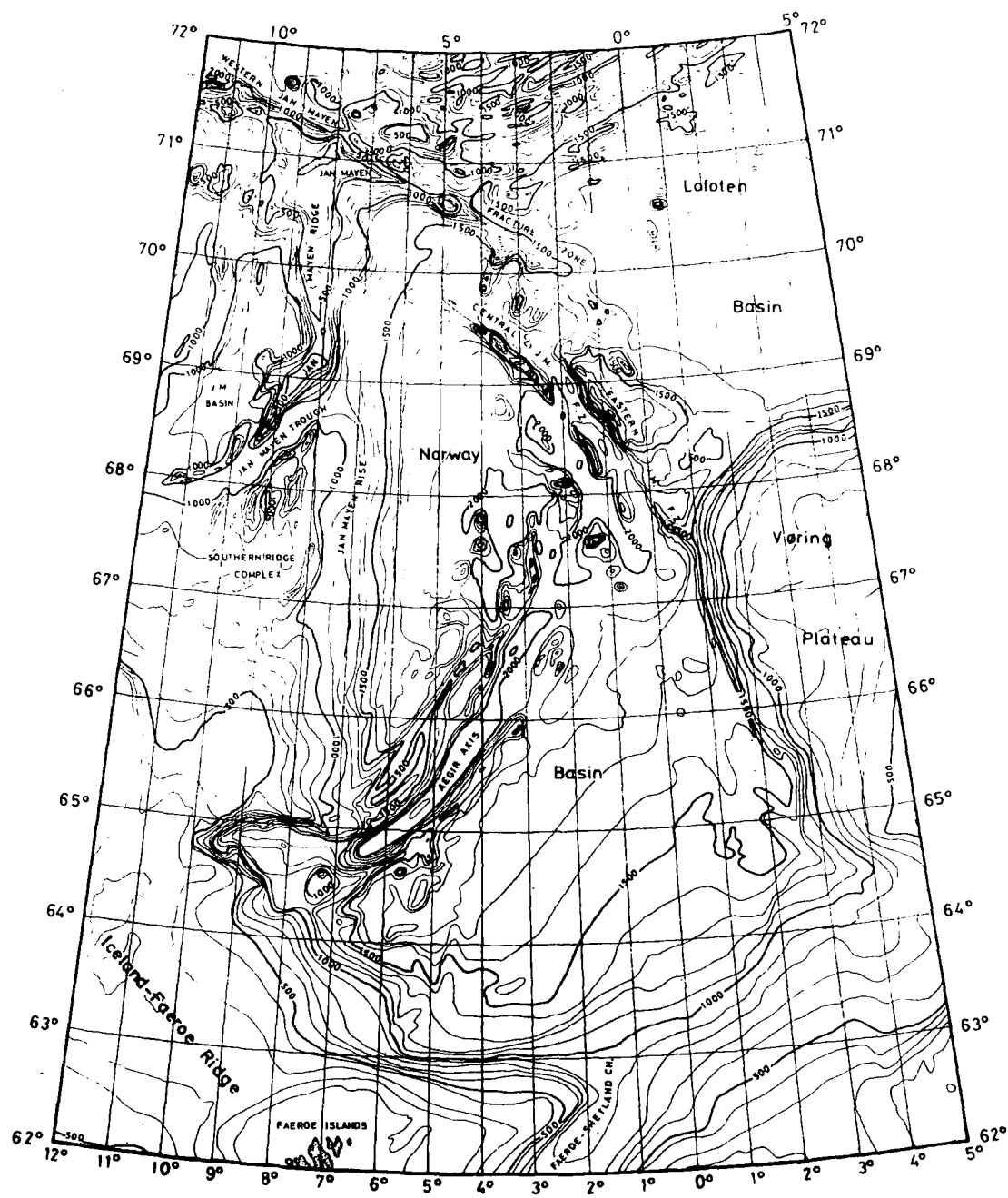
Since 1977, most workers have used Talwani and Eldholm's (1977) synthesis as a basic framework. However Talwani and Eldholm's ocean-continent boundaries in the vicinity of the Jan Mayen Ridge have been disputed by Gairaud et al. (1978) and Vogt et al. (1980) have challenged the concept of the Iceland Plateau intermediate axis (a modification of Johnson and Heezen's hypothesis which was originally suggested by Vogt et al. (1970) and was adopted by Talwani and Eldholm (1977)).

In the following sections the various provinces of the southern Norwegian-Greenland Sea will be discussed in detail and the important recent results will be introduced.

1.3 The Norway Basin

Bathymetry and basement relief

Most of the Norway Basin (Figure 1.4) lies deeper than 2500 m below sea level, with maximum depths of about 3800 m in the central region. The main feature of the Norway Basin is the Aegir axis, which is a buried basement rift structure with associated marginal ridges. The Aegir axis has been identified by Talwani and Eldholm (1977) with the extinct spreading axis which was proposed by Johnson and Heezen (1967). It gives rise to a prominent linear gravity anomaly (Figure 1.5) which trends southwest from 67.7°N , 1.8°W to 64.8°N , 6.3°W , but is less evident in the bathymetry as the basement relief is subdued by a considerable thickness of sediments. The apparent sinistral offset of the axis (as defined by the 2000 fathom contour in Figure 1.4) at 66.3°N is not reflected in the gravity anomaly chart. The Durham University 1977 Norway Basin traverse (Chapter 2) confirms that the ridge continues northeast without offset. It is



POLAR-STEREOGRAPHIC PROJECTION
 Depth in nominal fathoms after
 Granlie and Talwani (1978)
 Scale at 70°N
 0 100 200 300
 km

Figure 1.4 Bathymetry of
 the Norway Basin
 (100 nom. fathoms contour int.)

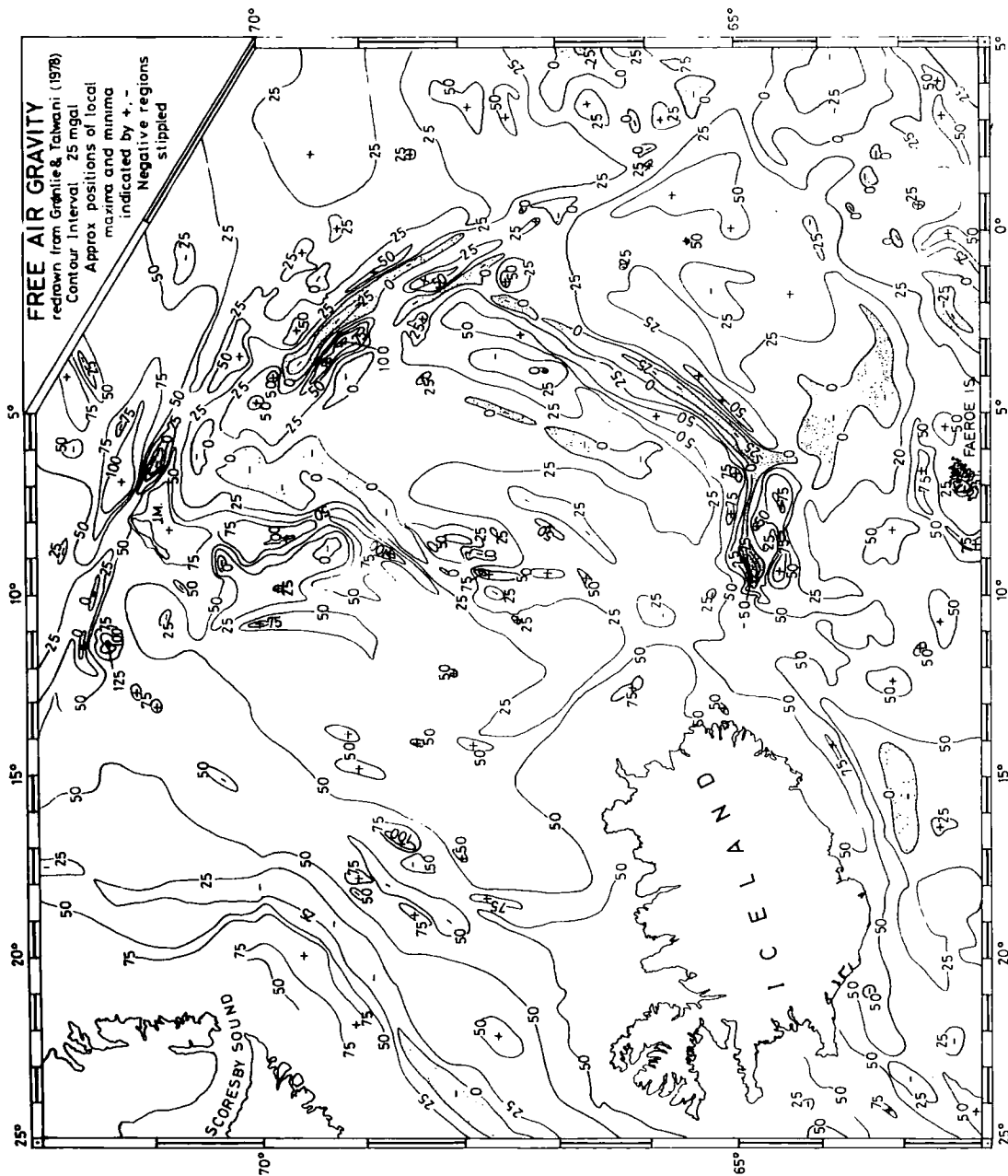


Figure 15 Free air gravity anomalies over the southern Norwegian-Greenland Sea
 International Reference Ellipsoid (Flattening 1/297)

noted also that the substantial sinistral offset of the axis at 66.6°N which was postulated by Gairaud et al. (1978) appears to be based upon the misidentification of a sub-parallel depression (67°N , 3.8°W to 68°N to 3.3°W) as the axis. At its southern end the Aegir axis joins the prominent bathymetric bight which is formed between the southern end of the Jan Mayen block and the Iceland-Faeroe Ridge.

The distribution of sediment in the Norway Basin (Eldholm and Windisch 1974, Gronlie and Talwani 1978) is in broad accordance with the hypothesis of Johnson and Heezen (1967). In general the sediment thickness increases away from the Aegir axis, as would be expected. However, the majority of the sediment within the basin appears to have been deposited since the cessation of spreading. In particular there is up to 1.4 s thickness (two way time) within the rift itself. Eldholm and Windisch (1974) consider that the basement relief in the vicinity of the Aegir axis is physiographically similar to that observed in mid-ocean ridges (presumably referring to slow spreading ridges (Laughton and Searle in press) which have median valleys). An obvious difference is that the Aegir axis lies deeper than active spreading ridges. Sclater et al. (1971) have demonstrated that in the event of an accreting margin becoming inactive the crust formed at the margin should continue to subside with age just as if spreading were still taking place. This is because the heat flow within the lithosphere, which determines the temperature profile, density distribution and hence crustal elevation, is predominantly vertical, so that spreading rate has no first order effect on the variation of elevation with age. From this point of view the sea floor in the Norway Basin is actually anomalously shallow when compared with the world-wide empirical depth-age relationship of Sclater et al. (1971), as is all of the sea floor in the Atlantic north of about 30°N (Cochran and Talwani, 1977). After correction for sediment loading, the basement depth in the Norway Basin is between 700 and 2000 m shallower than would be predicted (Cochran and Talwani, 1977, 1978; Gronlie and Talwani, 1978), with the smaller depth anomalies in the centre of the basin.

Seismic refraction studies

There have been a number of sonobuoy refraction profiles shot in the Norway Basin (Gronlie and Talwani, 1978) but these have not afforded much information on the basement structure. On the other hand the two ship refraction study of Hinz and Moe (1971) attained deep penetration (see Figure 1.2 for location of their profiles II and III). The interpretation of profile II is complicated by travel time anomalies which may be caused by an intrusive body; nevertheless the models for both profiles can be characterised as follows. Underlying the sediment layer there is an upper crustal layer in which the seismic velocity increases downwards from $3-4 \text{ km s}^{-1}$ to 5 km s^{-1} . The base of this layer at 6-7 km depth is marked by a velocity increase to about 7 km s^{-1} , below which the velocity increases to 8 km s^{-1} at about 10 km depth. At the time that these results were published they were remarkable as they showed velocity gradients within oceanic layers 2 and 3, which were previously assumed to have uniform velocities (for example Raitt, 1963). However the results do not differ markedly from those which have been subsequently obtained in other oceanic areas (see Section 4.3).

The earlier refraction profile F9 (Figure 1.2) of Ewing and Ewing (1959) which was shot parallel to the central Jan Mayen fracture zone (Section 1.10) in the northern Norway Basin showed a layer of velocity 5.4 km s^{-1} beneath the sediment layer, extending from about 4 to 7 km depth. This is underlain by a layer with an apparent velocity of 8.0 km s^{-1} . However on account of the topographic variations this velocity is unreliable and it was considered that the true velocity at 7 km depth could be as low as 7.4 km s^{-1} .

Magnetic anomaly studies

The aeromagnetic chart of Avery et al. (1968) shows a pattern of positive and negative magnetic anomaly stripes over the Norway Basin (trending approximately northeast), indicative of underlying oceanic crust (Vine and Matthews,

1963). However this chart is not accurate enough to allow confident anomaly identification according to the scheme of Heirtzler et al. (1968). Marine magnetic profiles across the basin exhibit symmetry about the Aegir axis, over which there is a consistent positive anomaly (Vogt et al., 1970; Talwani and Eldholm, 1977) confirming that the Aegir axis is the inferred spreading axis in the Norway Basin.

Vogt et al. (1970) considered anomaly 24 to be the oldest identifiable anomaly in the Norway Basin, as it is in the basins to the north and south. Talwani and Eldholm (1977), while agreeing that the Norway Basin began to form at or slightly before anomaly 24 time, proposed that a westward shift of the spreading axis occurred just before anomaly 23 time. The anomaly lineations in the Norway Basin, as interpreted by Talwani and Eldholm (1977) are shown in Figure 6.9. The oldest anomaly which they identified is anomaly 23, on both sides of the Norway Basin and no identifiable anomalies were associated with the inferred zone of early spreading, which lies immediately east of the Faeroe-Shetland escarpment (Talwani and Eldholm, 1972).

The major problem associated with the interpretation of the magnetic anomalies in the Norway Basin, as pointed out by Talwani and Eldholm (1977), is that "the magnetic anomalies lying between the extinct axis and anomaly 20 are not parallel, as magnetic anomalies generated by sea-floor spreading usually are; rather they form a fan-shaped pattern which is nearly 300 km wide at its northern end but decreases to about 150 km wide near the southern end of the extinct axis".

This fan-shaped anomaly pattern is apparently reflected in the pattern of topographic lineaments within the central Norway Basin and also in the southward convergence of the Jan Mayen and Norwegian rises. If this interpretation of the anomaly sequence is correct then simultaneous fan-shaped spreading (convergent to the north) must have also taken place elsewhere in the southern Norwegian-Greenland Sea, in order to maintain an overall parallel displacement of Norway and Greenland. Talwani and Eldholm (1977) suggested that

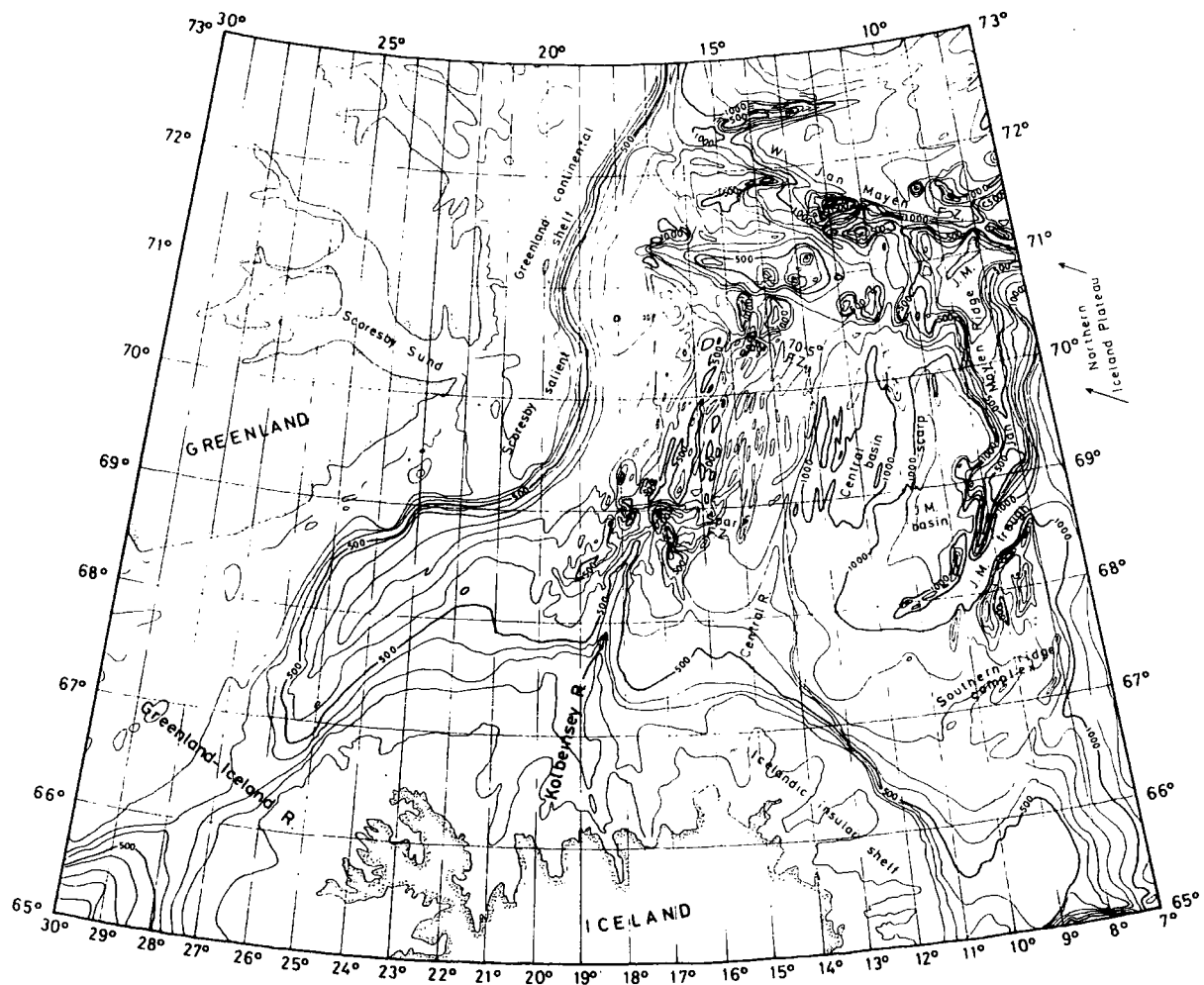
this complementary spreading took place south of the Jan Mayen Ridge, thereby creating the eastern part of the southern ridge complex (Section 1.8, Figure 6.14).

Talwani and Eldholm (1977) proposed that spreading in the Norway Basin ceased at anomaly 7 time on the basis of "the identification of anomalies as well as from geometrical considerations". The precise reasons for this judgement were not given. No anomalies younger than anomaly 20 were explicitly identified by Talwani and Eldholm (1977) and while cessation of spreading at anomaly 7 time is consistent with a qualitative interpretation of the anomaly sequence, it does not appear to be directly deducible from the sequence. The deep sea drilling evidence is not decisive. The oldest sediments recovered from DSDP site 337 (20 km east of the Aegir axis, see Figure 1.2) yielded a faunal age of Middle Oligocene-Late Eocene (29-43 Ma). The underlying basalt, which is definitely not intrusive, gave radiometric ages of 18 ± 1.5 and 25.5 ± 2.4 Ma (Kharin et al., 1976). Previous authors have adopted the younger faunal age limit for the time of formation of the basement (on the grounds of greater consistency with the radiometric ages) and making a 2 Ma allowance for spreading from the axis to site 337, have given an age of 27 Ma for the cessation of spreading (Talwani and Eldholm, 1977). However, once the probability of substantial argon loss in the basalt is accepted (Kharin et al. 1976), an older age appears possible. One of the primary aims of the present study is the clarification and re-interpretation of the magnetic anomaly pattern and spreading history of the Norway Basin.

1.4 The Iceland Plateau

Bathymetry, basement relief and tectonics

The average depth of the Iceland Plateau is about 1500 m below sea level (Gronlie and Talwani, 1978). The morphology of most of the plateau is dominated by topographic elements trending roughly N20°E, perpendicular to the western Jan Mayen fracture zone (Figure 1.6). However, in a zone about 120 km wide, immediately south of the main north facing scarp-



POLAR-STEREOGRAPHIC PROJECTION

Depth in nominal fathoms after
Grönlie and Talwani (1978)

Scale at 70°N

0 100 200 300
km

Figure 1.6 Bathymetry of the Iceland Plateau (100 nom. fathoms contour int.)

of the fracture zone the topography appears to be dominated by transverse, rather than longitudinal elements, and is generally elevated with the rest of the plateau. This zone will be distinguished as the northern Iceland Plateau.

The Kolbeinsey Ridge is the present axis of spreading within the southern Norwegian-Greenland Sea, as is illustrated by the belt of earthquake epicentres (see Figure 1.2) which follows the ridge crest (Sykes, 1965; Husebye et al., 1975; Bungum and Husebye, 1977). The Kolbeinsey and Mohns Ridges are connected by a nearly 200 km long seismically active section of the western Jan Mayen fracture zone, along which there is sinistral transform motion. In the south the spreading axis shifts eastward from the Kolbeinsey Ridge to the northern neovolcanic zone in Iceland, via the Tjornes fracture zone (Sykes 1967), along which dextral transform motion is taken up by several faults (Saemundsson, 1974; McMaster et al., 1977; Einarsson, 1979).

The morphology of the Kolbeinsey Ridge changes along its length (Meyer et al. 1972). The ridge is first recognisable at about 66.5°N where its crest is less than 200 m deep and is flanked on each side by a valley which embays the Icelandic shelf. The ridge crest deepens to about 1000 m approaching 69°N , where it is offset about 30 km east by the Spar fracture zone (Johnson and Heezen, 1967; Meyer et al., 1972). Between the Spar fracture zone and 70.5°N a shallow axial valley is present and the ridge flanks have a rugged morphology dominated by linear scarps and slopes (Meyer et al., 1972). In this region the depth to the ridge is about 1000 m. Talwani and Eldholm (1977) have postulated a fracture zone at 70.5°N , offsetting the ridge east. On the basis of the distribution of earthquake epicentres Bungum and Husebye (1977) locate this fracture zone further north at 70.75°N , coincident with the southern margin of the northern Iceland Plateau. The topography suggests that two fracture zones may actually be present, as shown in Figure 1.2. Within the northern Iceland Plateau Bungum and Husebye (1977) consider that the axis trends from 70.7°N , 13°W to 71.7°N , 12.2°W . Seamounts rising to within 200 m of the surface may represent the ridge in this region.

Another important physiographic feature of the Iceland Plateau is the elongate Iceland Plateau central ridge (the western ridge of Gronlie and Talwani (1979) and the lineament EK' of Vogt et al. (1980)) which is well defined in the south (68°N , 15°W) where it has about 300 m of relief, but becomes less distinct in the north (69.5°N , 13°W) and does not persist into the northern Iceland Plateau. This ridge forms the western boundary of the central Iceland Plateau, in which the central basin (68.5°N to 70.3°N) is situated. The floor of the central basin at 1800 m depth, lies some 300 m deeper than the outer eastern flank of the Kolbeinsey Ridge. Between 60°N and 70.3°N , a small ridge rises at the eastern edge of the central basin, east of which there is a sharp drop in depth to the floor of the Jan Mayen basin, which is 200-400 m deeper than the central basin. This scarp forms the eastern boundary of the central Iceland Plateau between 69°N and 70.3°N .

The Iceland Plateau is anomalously elevated. North of the Spar fracture zone the depth anomaly (Sections 1.2 and 4.5) over the axial zone of the Kolbeinsey Ridge is about 1800 m (Gronlie and Talwani, 1978). The depth anomaly increases moving closer to Iceland and also further away from the ridge and has an average value of about 2000 m over the whole plateau.

Seismic reflection studies of the Iceland Plateau reveal some interesting results. South of the Spar fracture zone the axial zone of the Kolbeinsey Ridge is narrow with steep flanks. On either side of the axial zone the acoustic basement is exceptionally smooth and highly reflective. In some regions peaks appear to penetrate the smooth surface of the opaque layer, as it is called (Eldholm and Windisch, 1974) and in others the level of the acoustic basement changes abruptly. North of the Spar fracture zone the axial zone is wider. On the western flank the rough basement topography persists under the sediment cover at least as far west as the foot of the Greenland slope. On the eastern flank there is a basement transition (from rough to smooth) approximately coincident with the central ridge. As far as can be determined, from the limited amount of seismic data

available, most of the central Iceland Plateau region is characterised by a smooth acoustic basement. DSDP hole 348 showed that the opaque layer in the central Iceland Plateau region consists of basalt (Talwani, Udinstev et al., 1976) rather than volcanic ash or turbidite (Eldholm and Windisch, 1974). Talwani (1978) concluded that the opaque layer was emplaced as a submarine flood basalt, possibly some time after the creation of the bulk of the oceanic crust. Johnson and Palmason (1980) have suggested the same origin for the "ultra-flat opaque reflector" which is observed on the flanks of the Reykjanes Ridge.

Seismic refraction and surface wave studies

Sonobuoy seismic refraction studies (Gronlie and Talwani, 1978) indicate that the acoustic basement in the Iceland Plateau has a variable seismic velocity; sometimes as low as 3.5 km s^{-1} and rarely exceeding 5 km s^{-1} . Shallow sub-basement reflectors (1-3 km below basement) showing a velocity increase of the order of 1 km s^{-1} are sometimes observed, but deep subcrustal refractors are not. Talwani and Eldholm (1977) consider the basement velocities to be similar to those of subaerial flood basalts.

Evans and Sacks (1979) have investigated the deep seismic structure of the Iceland Plateau by modelling the dispersion of surface wave trains which were generated by earthquakes on the Jan Mayen fracture zone and recorded in northern Iceland. Typical oceanic dispersion curves are not observed. The shear velocity-depth models are characterised by an increase of crustal velocity with depth from an initial (sub-sediment) value of about 2.6 km s^{-1} . Evans and Sacks have defined the depth of the crust-mantle boundary as the depth at which the shear velocity attains a value of 4.35 km s^{-1} . The crust, so defined, has an anomalously large thickness of 15-20 km and overlies a nearly constant velocity "lid" (maximum velocity $4.5\text{--}4.6 \text{ km s}^{-1}$). The depth to the top of the low velocity channel underlying the lid (that is the thickness of the "seismic lithosphere") is found to be 40-50 km, a value which is consistent with other determinations of thicknesses of young oceanic lithosphere.

Evans and Sacks suggest that most of the anomalous elevation of the Iceland Plateau may be isostatically compensated for within the lithosphere by the thick crust (Section 4.5).

Magnetic anomaly studies

Correlatable sea-floor spreading magnetic anomalies exist over the Iceland Plateau west of the scarp dividing the central Iceland Plateau from the Jan Mayen Basin. Anomaly 5 is prominent on both sides of the Kolbeinsey Ridge (Vogt et al. 1970). The trend of anomaly 5 is not offset by the Spar fracture zone. Meyer et al. (1972) concluded from a detailed magnetic survey (R.V. Komet, 1971) that the Spar fracture zone was formed by an eastward jump of the axis north of 69°N at about anomaly 2A time. From a limited amount of data Talwani and Eldholm (1977) inferred that the 70.5°N fracture zone must also have been formed since anomaly 5 time. Vogt et al. (1980) have noted from Project Magnet data (south of 70°N) that "numerous minor fracture zones and bends have formed at various times subsequent to anomaly 5 time" and that "fractures of small offset tend to be ephemeral features". Their identifications seem to indicate that the main shift north of the Spar fracture zone occurred at about anomaly 3A time (Figure 6.7). When the shifts of axis are taken into account, it can be inferred that accretion at the Kolbeinsey Ridge axis has been symmetrical since anomaly 5 time with a half-spreading rate of 1 cm/yr (Vogt et al., 1980).

On the western side of the Kolbeinsey Ridge anomaly 5 lies near the Greenland shelf break in the vicinity of the Scoresby Salient; on the other side, it lies just west of the central ridge. The central Iceland Plateau is characterised by linear magnetic anomalies striking parallel to the Kolbeinsey Ridge. Vogt et al. (1970) suggested that an intermediate Iceland Plateau axis existed after the cessation of spreading on the Aegir axis and prior to the start of spreading on the Kolbeinsey axis. Talwani and Eldholm (1977) adopted this idea and Gronlie et al. (1979) defined an axis of symmetry of the magnetic anomalies, which had, however, no expression in the basement topography or gravity field. Vogt

et al. (1980) have convincingly demonstrated that this proposed spreading axis did not exist. Instead, the Kolbeinsey Ridge lineation pattern simply continues eastward, from anomaly 5A, which is coincident with the central ridge through to anomaly 6B-6C which is coincident with the scarp bounding the Jan Mayen Basin (see Figure 6.7). The same anomalies, albeit with diminished amplitudes, can also be identified west of the axis, where the older oceanic crust has been partly buried under the sediments of the Greenland shelf, which has prograded dramatically in late Tertiary times (Vogt et al., 1980). DSDP hole 348 (Talwani, Udinstev et al., 1976) in the Central Iceland Plateau (Figure 1.2) did not test the intermediate axis hypothesis satisfactorily. The hole was drilled east of the proposed axis, into crust of anomaly 6 age (20.7 Ma, Hailwood et al. (1979)) according to the identification of Gronlie et al. (1979). The simpler hypothesis of Vogt et al. (1980) predicts that the crust is of anomaly 6A age (22.5 Ma, Hailwood et al 1979), less than 2 Ma different. A radiometric age of 18-19 Ma (Kharin et al., 1976) and a sediment age of Oligocene-Lower Miocene was obtained. Bearing in mind the likely errors of these age determinations, both hypotheses are supported equally well.

Outside of the latitude range 66.5°N to 70°N the magnetic anomaly pattern of the Iceland Plateau is not well understood. Isolated profiles across the northern Iceland Plateau (Mirlin and Melikhov, 1975; Gronlie and Talwani, 1978) suggest that a linear anomaly pattern persists through this region; an axial positive anomaly is certainly present. In the southern part of the Iceland Plateau, there is a substantial attenuation of anomaly amplitude as Iceland is approached. There are a number of possible explanations for the magnetic smooth zone that is currently being formed where the Kolbeinsey ridge enters the Iceland Platform (Vogt et al., 1980).

The results of Vogt et al. (1980) are in agreement with the original hypothesis of Johnson and Heezen (1967) and allow a simple view of the tectonic evolution of the southern Norwegian-Greenland Sea, without the complication of considering an intermediate axis. In particular, it can be

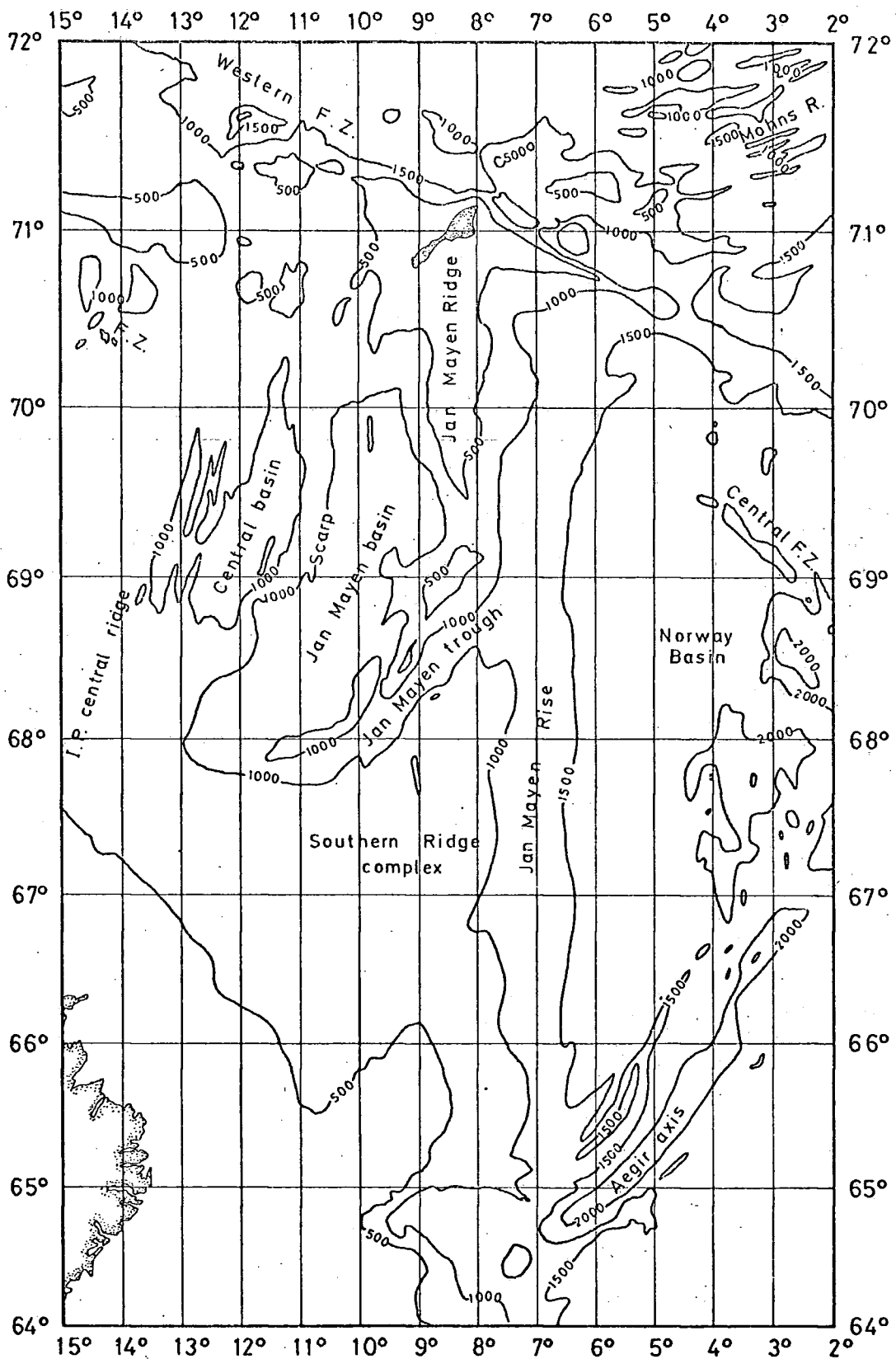
hypothesised that the Kolbeinsey axis first came into being when the fan-shaped spreading was taking place in the Norway Basin, and that the first stage of spreading on the new axis consisted of compensatory fan-shaped spreading convergent to the north. Vogt et al. (1980) consider that oceanic crust of anomaly 7 age or older could exist near the Greenland coast north of the Greenland-Iceland Ridge or in the western part of the Jan Mayen block.

1.5 The Jan Mayen block - general description

The term "Jan Mayen block" is used to denote a region which lies roughly between the following limits; 11°W and 7°W , 65°N and 71°N . It is composed of a number of main physiographic elements (see Figure 1.7) which are described below. The block as a whole is characterised by the absence of recognised sea-floor spreading magnetic anomalies and by relatively complex subsurface structures. However the use of the name "Jan Mayen block" is not meant to imply a necessary common geological origin for the various subregions.

The flat-topped Jan Mayen Ridge is broad and shallow in the north (corresponding to the northern Iceland Plateau zone) and narrower and deeper south of 70.4°N , terminating at 68.25°N . Out of the northern part, or Jan Mayen bank, rises the Tertiary volcanic island of Jan Mayen. Fitch et al. (1965) consider that the emergent part of the island has formed in the late Pleistocene. Initial activity probably formed the southwestern part of the island, followed by a shift to the presently active (Gjelsvik, 1970; Zobin, 1972) Mt Beerenberg which is located at the northeastern end of the island, near the termination of the ridge by the Jan Mayen fracture zone and adjacent to the southern end of the Mohns Ridge.

The Jan Mayen basin, which has a smooth floor everywhere deeper than 2000 m below sea level, is separated from the Jan Mayen trough by a small southwestern extension to the Jan Mayen Ridge which is depicted variously on different bathymetric charts (Perry et al, 1977; Gronlie and Talwani, 1978; Gairaud et al, 1978). The Jan Mayen trough is a major



MERCATOR PROJECTION
 Depth in nominal fathoms after
 Grønlie and Talwani (1978)

Scale at 70°N

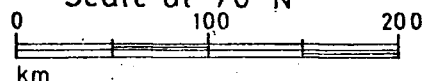


Figure 1.7 Bathymetry of the Jan Mayen block (500 nom. fathoms contour interval)

southwest trending depression about 30 km wide and 2000 m deep, south of which is the southern ridge complex. Within the ridge complex a series of narrow discontinuous ridges, trending between north and northwest, rise to within 1000 m of the sea surface from a general topographic level of about 1500 m.

In the following sections (1.6 to 1.9) a summary will be given of published geophysical and geological information about the various structural units which make up the Jan Mayen block. The most well understood region, the Jan Mayen Ridge and rise between the latitudes of 68.5°N and 70.5°N is treated first (Section 1.6), followed by a discussion of the Jan Mayen bank (Section 1.7). The more complex southern region, comprising the Jan Mayen trough, the southern ridge complex and the Jan Mayen rise south of 68.5°N will then be dealt with (Section 1.8). In Sections 1.6 to 1.8, more emphasis is placed on the eastern parts of the Jan Mayen block, so to round the discussion off, the Jan Mayen basin, and the likely western limits of continent crust are considered in Section 1.9.

1.6 The Jan Mayen Ridge and rise between 68.5°N and 70.5°N

In contrast to the central Iceland Plateau to the west, the Jan Mayen bank to the north and the Norway Basin to the east, this region is characterised by a relatively smooth magnetic field (Johnson et al., 1972; Talwani and Eldholm, 1977; Gronlie et al., 1979; Navrestad and Jorgensen, 1979; Vogt et al., 1980) although some weak lineations are observed. The seismic reflection profile across the Jan Mayen Ridge at 69°N which was presented by Johnson and Heezen (1967) provided no firm evidence as to the age of the prominent unconformity or of the underlying sediments. This problem has been resolved unequivocally by drilling (Talwani, Udinstev et al., 1976) and by multichannel seismic reflection profiling.

DSDP holes 346, 347 & 349 - the Oligocene unconformity

DSDP holes 346, 347 and 349 (Figure 1.7a) all penetrated the unconformity. In the following discussion the post- and pre-unconformity sediments will be referred to as sequences 1 and 2 respectively. In order to describe the constraints which the DSDP results place upon the age relationship of the two sequences a certain amount of detail is necessary; firstly because the results presented in DSDP site summaries (Talwani, Udinstev et al., 1976, pages 521-594) differ somewhat from the results of the shore based studies and secondly because care is needed in relating the paleontological ages to the magnetic polarity timescale used in the present study (Hailwood et al., 1979). In all three holes the unconformity was reached at about 120 m. The oldest fossils dated near the base of sequence 1 (holes 346 and 349) were dinocysts from the local zone III (Manum, 1976) of upper middle Oligocene age. Comparison of the Leg 38 timescale (Schrader et al., 1976) with the Leg 48 timescale (Hailwood et al., 1979) suggests that this zone has an absolute age range of 30-27 Ma (anomalies 9-7). The youngest fossil assemblage recovered from sequence 2 was a nannofossil assemblage belonging to the late Eocene - early Oligocene Isthmolithus recurvus zone (Muller 1976) corresponding to the standard zones NP19 to NP22 (40-35 Ma, anomaly 17-13, Hailwood et al., (1979)). The oldest sediments dated from sequence 2 (hole 347) belong to the Reticulofenestrata umbilica nannofossil zone (Muller, 1976) of middle Eocene age (NP15-NP18, 48-40 Ma, anomaly 21-18, Hailwood et al., 1979) although reworked fossils with Lower Cretaceous ages were found.

In all three holes sequence 1 was comprised of several sedimentary units, with the dominant lithologies being unconsolidated muds and sandy muds. The uppermost unit in each hole is a Plio-Pleistocene terrigenous sandy mud of glacial-marine ice-rafted origin (40-60 m thick). This unit is found throughout the Norwegian-Greenland Sea but is not normally distinguishable on seismic sections. In hole 346 (which was the most completely cored) a sedimentary hiatus spanning the Late Miocene is indicated. Schrader and Fenner

JAN MAYEN RIDGE

SITES 346 AND 347

SITE 346 - 69°53.35'N 08°01.14'W
 SITE 347 - 69°52.31'N 08°01.80'W

JAN MAYEN RIDGE; NEAR WESTERN END OF MAIN BLOCK.

Basement was not reached at either site.

Three lithologic units are present at site 346. The lowest (121.8-187.0 m) is of Eocene age and consists of massive terrigenous sandy and pebbly mudstone. Deep water conditions are indicated.

The middle unit (39.7-121.8 m) is Oligocene to Middle Miocene in age and consists of biogenic siliceous mud as well as terrigenous siliceous sandy mud and thin ash layers.

The top unit (0-39.7 m) is of Middle Miocene to Pleistocene age and consists of sandy mud and mud with volcanic ash.

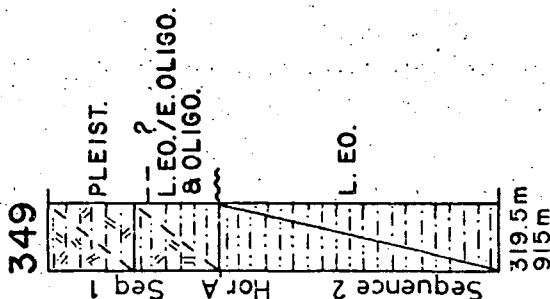
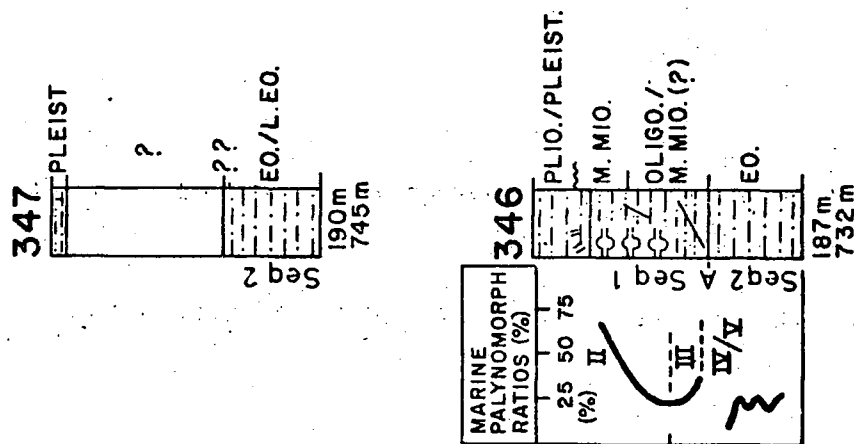
Below 121 m (to 137 and 187-190 m) at site 347 terrigenous sediments similar to the lower unit at site 346 of Eocene age are present.

An unconformity is present at the border between Eocene and Oligocene sediments (121 m).

COMPARISON SITES 346, 347 AND 349

A deep water origin is suggested for the Eocene (largely terrigenous) deposits at all sites based on faunal evidence. An uplift has taken place since that time.

The unconformity with dipping layers below it suggests a tectonic event. A noticeable increase in the biogenic component takes place above one unconformity. The ash layers in the Pleistocene may be associated with the volcanic activity on Jan Mayen Island.



SITE 349
 69°12.41'N 08°05.80'W
 JAN MAYEN RIDGE.

Basement was not reached. Three lithologic units are present. The lowermost (119.5-319.5 m) is of Late Eocene to Early Oligocene age and consists of breccia, sandy mudstone, and sandstone, wholly terrigenous.

The middle unit (59.3-119.5 m) is of Late Eocene or Early Oligocene to Pleistocene age and consists of massive muds and layers of volcanic ash. There is a prominent unconformity between the innermost and the middle units (119.5 m). The top unit (0-59.3 m) consists of massive and sandy mud with ash layers of Pleistocene age.

SITE 350
 67°03.34'N 08°17.68'W
 SOUTHERN EXTENSION OF THE MORPHOLOGICAL JAN MAYEN RIDGE

The radiometrically determined age for basement is approximately 40-44 my.

Three lithologic units overlie basement. The lowermost unit (226-364 m) is dated as Late Eocene to Oligocene and consists of breccia, claystones, mudstones, limestones and volcanic ash layers. Turbidites are prominent in the lower part.

The middle unit (55.5-226 m) is Oligocene to Middle Miocene of age and consists of terrigenous clay/claystones and mud/mudstones.

Volcanic ash layers are present. The top unit (0-55.5 m) consists of sandy mud and volcanic ash of Plio/Pleistocene age.

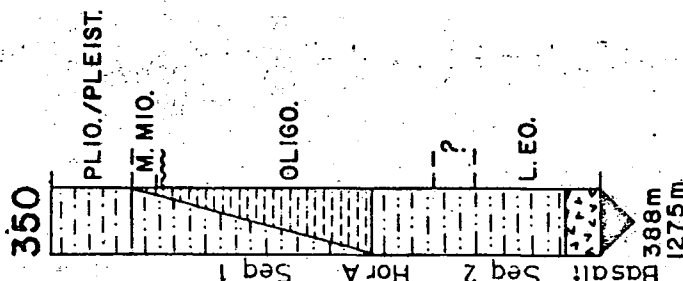


Figure 1.7a Summary of DSDP results. Retouched after Gronlie & Talwani (1978).

(1976) from a study of diatom faunas infer a shallowing in Middle Pliocene times (magnetic epoch 14, 13 Ma, between anomalies 5A and 5B; Opdyke et al. (1974)). The units which underlie the "glacial" unit contain more biogenic material than the latter does, but are still surprisingly rich in terrigenous detritus in view of the drift of the ridge away from Greenland; possibly much of this material is reworked. Basic volcanic ash and glauconite-rich layers are common throughout.

Sequence 2 was found to consist of lithified terrigenous sandy mudstone, sometimes massive and sometimes layered. The nature of the terrigenous clastic material is consistent with derivation from the continental land mass of Greenland (Rateev et al., 1979).

In summary the DSDP results support the hypothesis that the sediments of sequence 2 were deposited adjacent to an ancestral Greenland margin. The oldest sediments sampled post-date the onset of spreading in the Norway Basin. Deposition of this sequence continued until late Eocene (anomaly 17) time at least, and then was interrupted by a period of uplift and erosion. No clear evidence for sub-aerial erosion exists, although Rateev et al. (1979) note the existence of humus at the base of sequence 1 in hole 346. Deposition recommenced by late Oligocene (anomaly 7) time at the latest. The unconformity will be referred to as the Oligocene unconformity. An important change also occurred in the Middle Miocene; an apparent shallowing was followed by a depositional hiatus. It is noted that in hole 348 (central Iceland Plateau) a major change occurs at approximately the same stratigraphic level. The Oligocene - early Miocene basal unit consisting of lithified terrigenous mudstone is overlain by a middle Miocene - Pliocene biogenic mud. This change was explained by Talwani, Udinstev et al. (1976) as being due to the shift of spreading from the central Iceland Plateau axis to the Kolbeinsey axis. If the former axis did not exist (Vogt et al., 1980) an alternative explanation is necessary.

Reflection seismic investigations of the Jan Mayen Ridge

On account of limited penetration of sequence 2 (200 m in hole 349) the nature of the underlying basement was not established by drilling. Single channel seismic profiles (Johnson et al., 1972; Eldholm and Windisch, 1974; Johnson, 1975) indicate a basement culmination along the western edge of the ridge but the basement cannot be traced as it deepens to the east. In contrast, the base of sequence 2 can be recognised throughout on multichannel seismic profiles, allowing certain inferences about the age of the sequence and the nature of the underlying basement.

Figure 1.8 shows the locations of the processed multichannel seismic reflection profiles from the Capan 1 survey which were presented by Gairaud et al. (1978). Line drawing of the sections with geological interpretations are given in Figures 1.9 and 1.10. The structure of the northern Jan Mayen block is illustrated by profiles CP114, 110 and 129. The following discussion draws on the conclusions of Gairaud et al. (1978) and also Hinz and Schluter (1979a), who described the BGR multichannel seismic survey of the Jan Mayen block and Navrestad and Jorgensen (1979), who interpreted the Capan 1 profiles in light of an aeromagnetic survey of the Jan Mayen Ridge.

Sequence 1 and Horizon A (the Oligocene unconformity)

The Oligocene unconformity, represented as Horizon A, is horizontal under the ridge top, as observed by Johnson and Heezen (1967). Under the eastern flank of the ridge Horizon A cuts deeply into the underlying sediments and sequence 1 is up to 2 s thick (all "thicknesses" quoted in seconds refer to two way travel times). Referring in particular to CP129, the following points are noted. As shown in Figure 1.9, Horizon A cuts across sequence 2 both under the ridge top and under the ridge flank, eastward to the point where it becomes nearly horizontal, clearly indicating an erosional unconformity. Further east no obvious evidence of erosion is seen, however Horizon A terminates against oceanic crust, which according to the results of the present study is of at

Figure 1.8 Tracklines for the Cepan 1 1975 survey of the Jan Mayen block. Redrawn from Figure 5 of Gairaud et al. (1978). Only those profiles which feature in Figures 1.9 and 1.10 are shown. Tracklines for the SHACK977 survey are also shown as broken lines (with each interval representing 10 km along track) together with bathymetric contours, locations of DSDP holes and positions of basement ridges (as determined from the free air gravity anomaly chart of Figure 1.13).

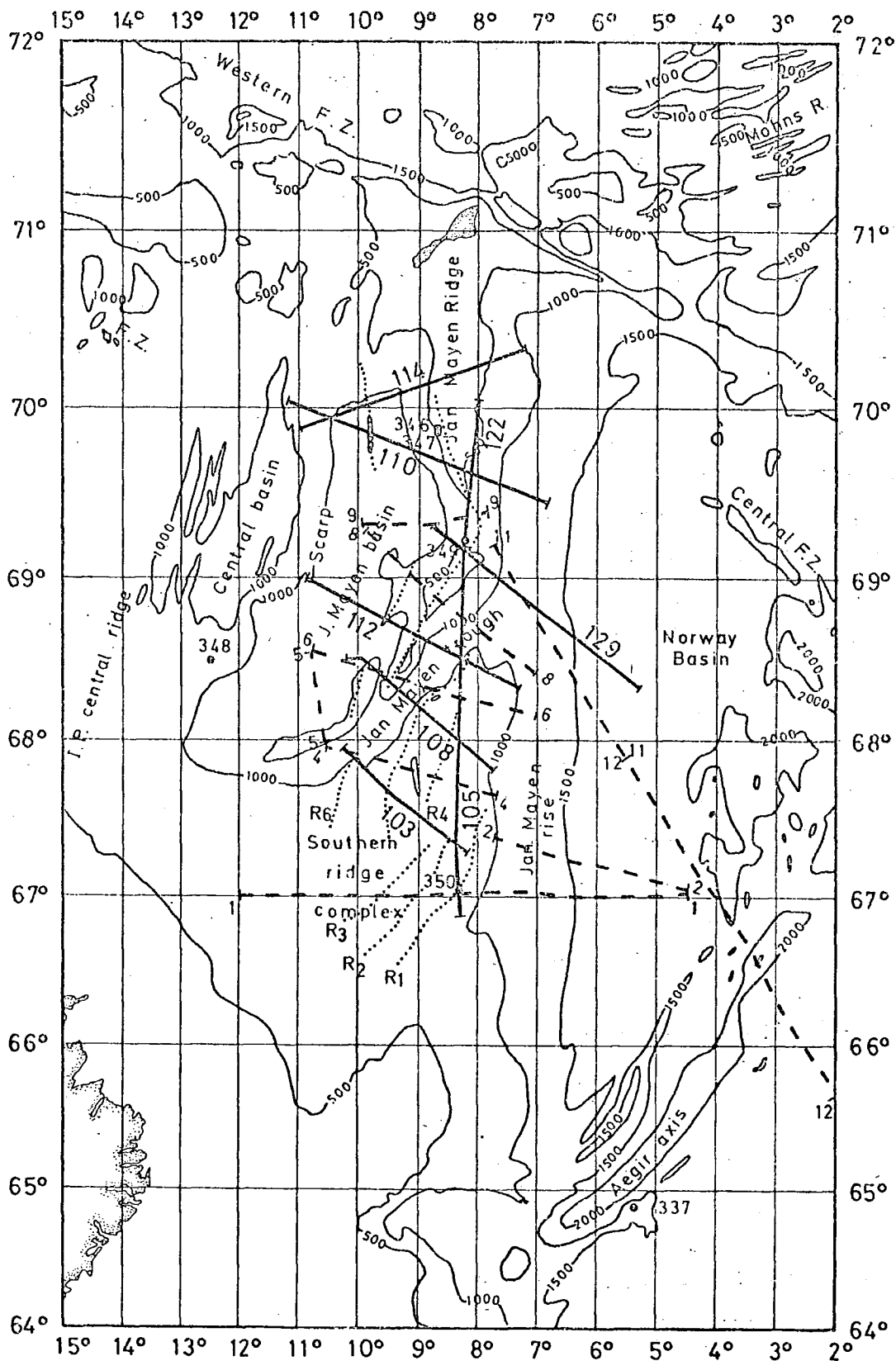
Figure 1.9 Line drawings of Cepan 1 24 channel seismic reflection sections, redrawn from Gairaud et al. (1978). The profiles in this diagram cross the Jan Mayen block and are arranged down the page from north to south. The point at which each profile crosses 8°W is indicated. Crossings with other Cepan profiles are indicated in the manner XCP110, with SHACK977 profiles in the manner XS11.

Also shown are sonobuoy refraction interpretations (Figure 1.11), projected along strike onto the Cepan 1 profiles. The horizontal bars indicate the two way times, relative to the sea bed, at which refractors occur.

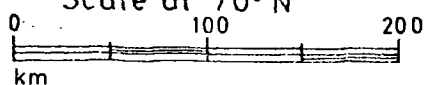
As described in the text the older magnetic anomalies may cross CP129 about 30 km further west than is indicated in the figure, if an important fracture zone (Section 6.4) crosses the eastern end of the line.

The identification of horizons follows that of Gairaud et al. Over the southern ridge complex (CP112, CP108, CP103), the labelling of basement ridges (for example R5 in CP108) follows the nomenclature of Talwani and Eldholm (1977). Gairaud et al. (1978) propose that a major transcurrent fault (the JAG fault) crosses the southern ridge complex between CP108 and CP103 and offsets the Jan Mayen trough eastward to the rift on CP103. However according to the present study, there is structural continuity of R5 between the profiles and the JAG fault does not exist. Instead the median axis of crustal rifting (indicated as M) is considered to step eastward going from north to south.

Figure 1.10 Line drawings of longitudinal Cepan 1 seismic sections, redrawn from Figure 5a of Gairaud et al. (1978).



MERCATOR PROJECTION
 Depth in nominal fathoms after
 Grønlie and Talwani (1978)
 Scale at 70°N



CEPAN 1 PROFILE
 112

SHACK 9/77 PROFILE
 6 0 16
 100 km

DSDP HOLE
 349

BASEMENT RIDGE
 R3

Figure 1.8

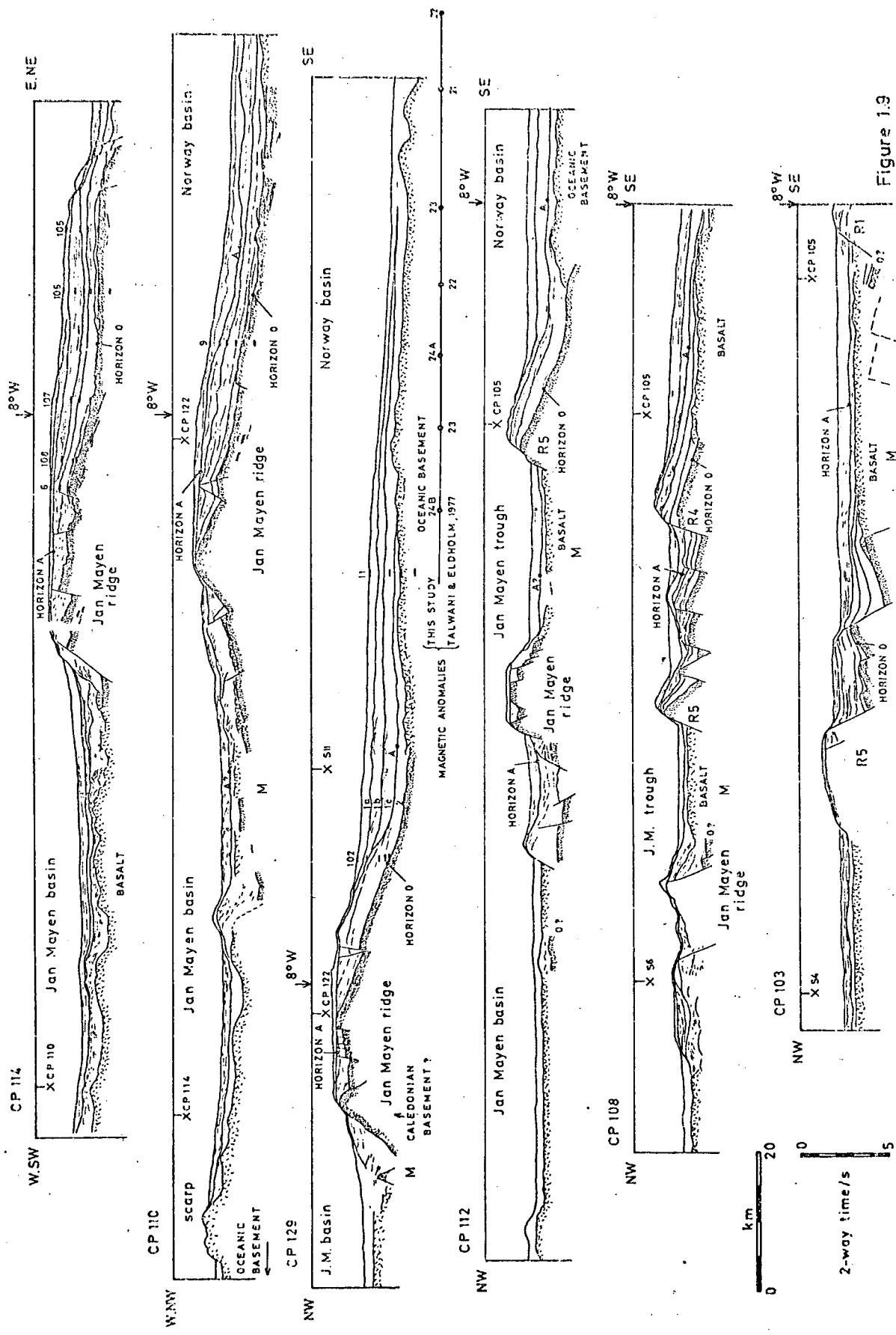


Figure 1.9

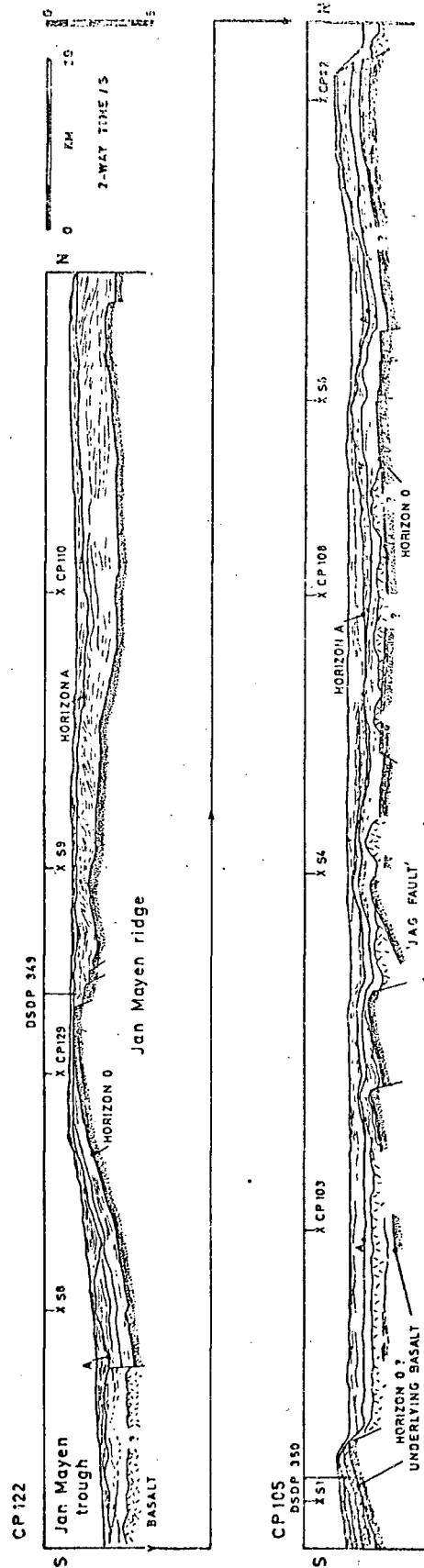


Figure 1.10

least anomaly 24B age. This must be explained either as a consequence of erosion or of prior non-deposition, as it is extremely difficult to see how Horizon A could itself be of anomaly 24B age in the region of the contact. On section CP129 sequence 1 consists of units 1a, 1b and 1c, as shown. Unit 1a is thin and horizontally layered and is apparently continuous with the ridge-top sequence, whereas units 1b and 1c onlap against the erosional unconformity, implying that the latter units were deposited prior to the resumption of deposition on the ridge top, while the ridge was still being actively eroded. Unit 1c is mainly reflection-free whereas unit 1b exhibits what the present author interprets as a chaotic reflection configuration (for example compare the original seismic section of Gairaud et al. (1978) with Figure 11 of Sangree and Widmier (1977)). Unit 1c could represent a massive sandstone or shale (Mitchum et al., 1977) derived from rapid erosion of fairly homogeneous material while the chaotic pattern of unit 1b could have resulted from either high energy turbidity current or mass-transport slump and creep processes (Sangree and Widmier, 1977). The latter explanation appears quite likely, in view of the hummocky top surface of unit 1b which is seen in CP129 and CP110 (see Section 3.1 for evidence of slumping on SHACK977 profile 11, which crosses CP129).

Sequence 2 and Horizon 0

Under the ridge the basement upon which the sediments of sequence 2 were deposited is represented as seismic Horizon 0, which is a distinct, smooth, locally faulted horizon which dips to the east (Figure 1.9). Sequence 2 is up to 1.8 s thick and in the main is conformable with Horizon 0 although there is weak onlap onto the ridge flank. Sequence 2 is characterised by strong reflectors which are parallel for up to 10-20 km. The line CP122, parallel to the ridge (Figure 1.10) shows structures which are suggestive of channel fill. The reflector configuration is consistent with sequence 2 being composed of alternating sandstones and mudstones deposited in a shelf or slope environment.

Underlying the lower Jan Mayen rise there is an abrupt

change from the smooth Horizon 0 to a rugged reflector having the hummocky appearance of oceanic basement. At the basement transition oceanic basalt appears to overlap Horizon 0 and there is an indication that Horizon 0 might persist underneath the basalt for about 15 km. There is no evidence to suggest that the basal sediments of sequence 2 are not continuous across the basement transition zone, although no reflector can be traced very far. Allowing this and noting that the basal sediments of sequence 2 appear to be nearly synchronous throughout, one can infer that most, or all of sequence 2 is younger than the oceanic basalt at the transition point. Furthermore, if it is assumed that sedimentation started immediately it can be inferred that the crust underlying Horizon 0 must be older than the basalt, a conclusion which is supported by the overlap of the latter onto the former. Anomaly identifications in the vicinity of the basement transition along CPL29 are uncertain. One possibility, based on the identifications of the present study (Section 6.4) is that anomaly 24B lies about 30 km east of the transition (Figure 1.10). However a major fracture zone, offsetting the anomalies sinistrally by about 30 km is known to be present in this region (Section 6.4) but its exact location is uncertain. If this fracture zone lies south of CPL29 then anomaly 24B must lie near the transition (this would bring the younger identifications into line with those of Talwani and Eldholm 1977, who do not identify anomaly 24). Either way the oceanic crust at the basement transition must be of anomaly 24B age or perhaps 1.5 Ma older.

Accepting the validity of these age relationships it can be argued that the Jan Mayen Ridge is not underlain by oceanic crust which was created in the present opening phase of the northeast Atlantic. This is because the present spreading phase began at most 1 or 2 Ma prior to anomaly 24 time. Although this was represented as an accepted fact in Section 1.1, it needs considerable justification, which is given in Section 6.2. The possibility that the Jan Mayen Ridge is underlain by oceanic crust created in an earlier spreading phase cannot be ruled out entirely but is considered unlikely (Section 7.2).

The evidence presented so far indicates that the Jan Mayen Ridge is probably underlain by continental crust, but it should be emphasised that the key evidence is indirect. A direct knowledge of the nature of the rocks underlying Horizon 0 is necessary if the origin of the ridge is to be established without doubt.

Under the western part of the ridge, where the basement culminates, the rock complex underlying Horizon 0 has a diffuse reflection pattern suggestive of some form of igneous or metamorphic basement (Hinz and Schluter, 1979a). Gairaud et al. (1978) have suggested that these rocks represent Caledonian crystalline basement, as shown in Figure 1.9. Under the eastern part of the ridge and ridge flank, discrete eastward dipping reflectors are apparent down to about half a second below Horizon 0, giving interval velocities of about 4 km s^{-1} . Gairaud et al (1978) and Hinz and Schluter (1979a) have proposed that a sedimentary succession underlies Horizon 0 in this region. The present author does not consider this evidence conclusive as similar reflectors have been observed within what must be oceanic crust west of Rockall Plateau (Roberts et al., 1979) and also under the Voring Plateau (Hinz and Schluter, 1979a).

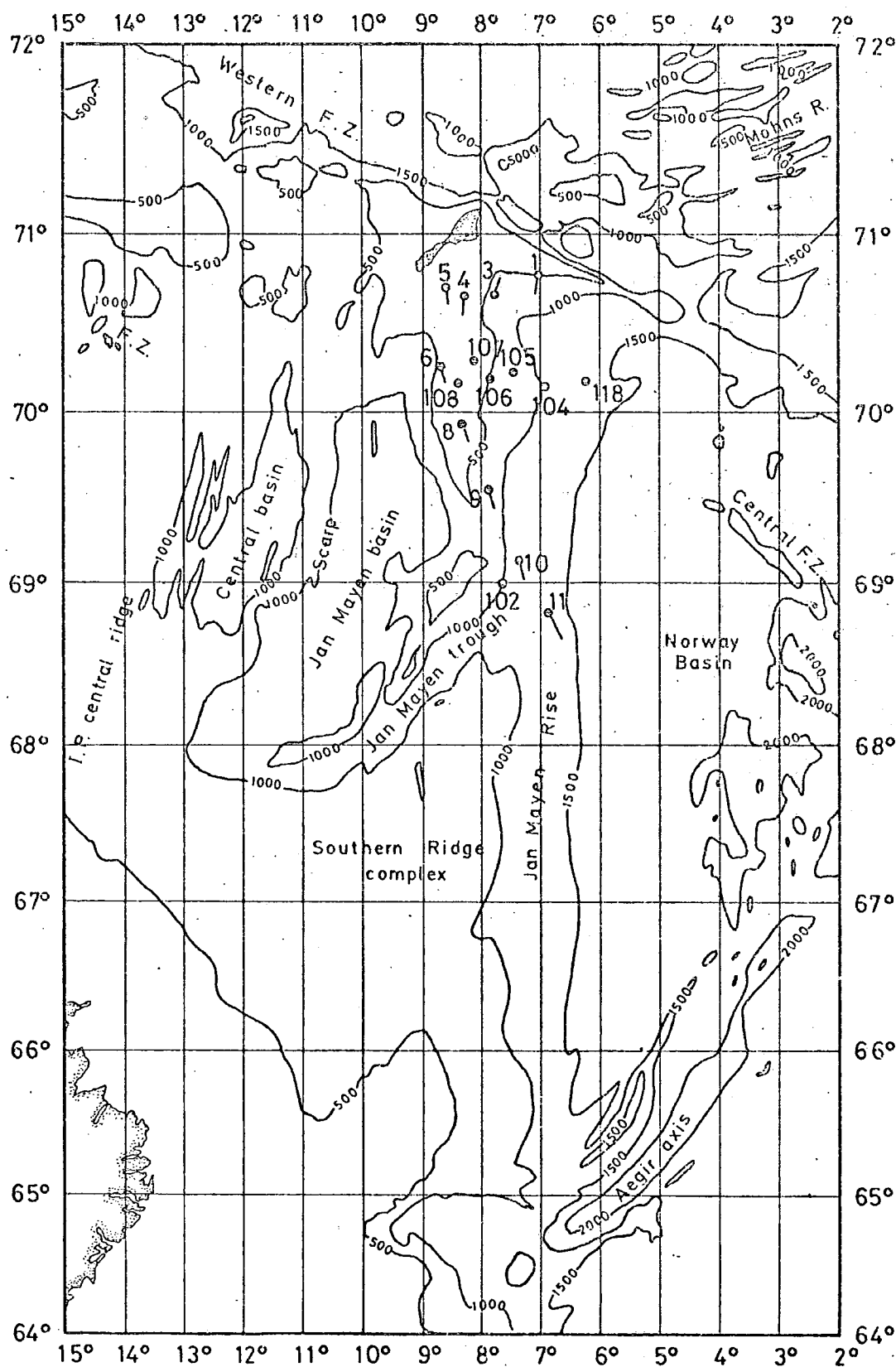
It is natural therefore to turn to refraction studies in order to try and identify the rock types which lie beneath Horizon 0. On the basis of some of the refraction data which are presented in the next section, Talwani and Eldholm (1977) have inferred the presence of a thick sequence of pre-Tertiary sediments beneath the Jan Mayen Ridge.

Seismic refraction studies

Figure 1.11 shows the location of sonobuoy seismic refraction profiles over the Jan Mayen Ridge. The results have been assembled by Sundvor et al (1979) as three seismic structure sections, which are presented in Figure 1.12. It should be noted however that the refraction correlations presented in this figure are quite different to those given by Sundvor et al. (1979) or Talwani and Eldholm (1977), which were based simply on refractor velocities. The

Figure 1.11 Location of sonobuoy refraction profiles over the Jan Mayen Ridge, as summarised by Sundvor et al. (1979). In this diagram (as in Figure 1.12 and the text) the profile labels are abbreviated. Treble figure numbers refer to R.V. Vema 29 sonobuoys (102V29 to 118V29) originally reported by Talwani and Eldholm (1977). The others refer to University of Bergen Seismological Observatory, 1978 sonobuoys (01B78 to 11B78).

Figure 1.12 Seismic structure sections, based on the sonobuoys of Figure 1.11. Redrawn from Sundvor et al. (1979). Velocities are in km s^{-1} . Assumed surface velocities are bracketted. The refractor correlations are different to those presented in the original sections and by Talwani and Eldholm (1977), as explained in the text. They are based on identification of refractors on seismic reflection sections (as in Figure 1.9).



MERCATOR PROJECTION
 Depth in nominal fathoms after
 Grønlie and Talwani (1978)
 Scale at 70° N

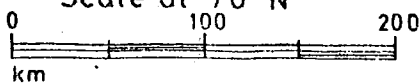


Figure 1.11

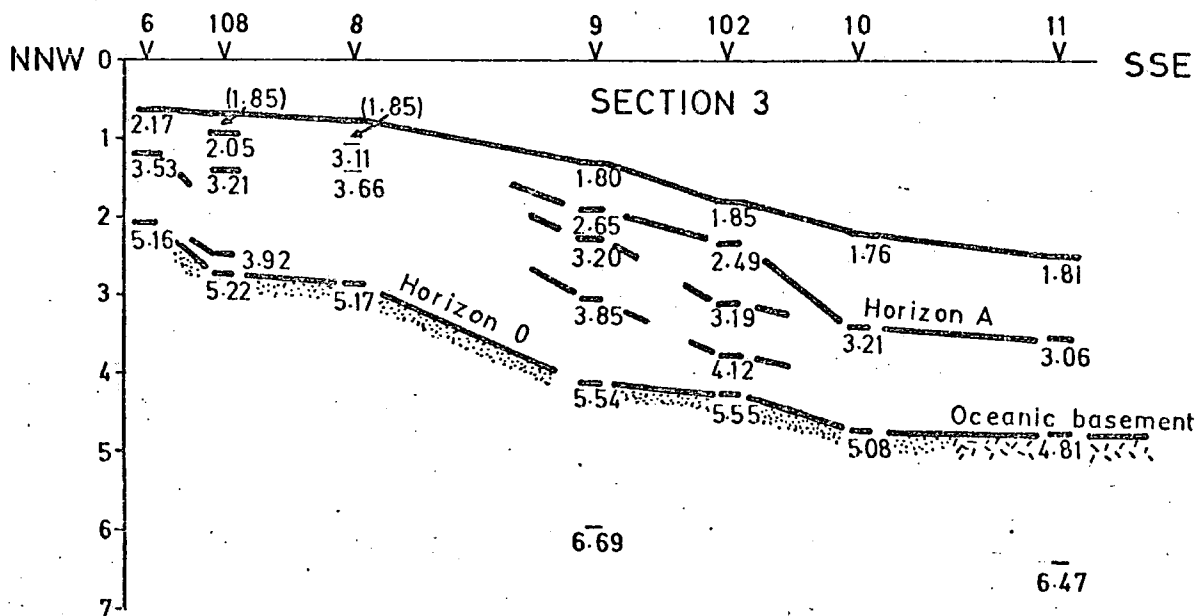
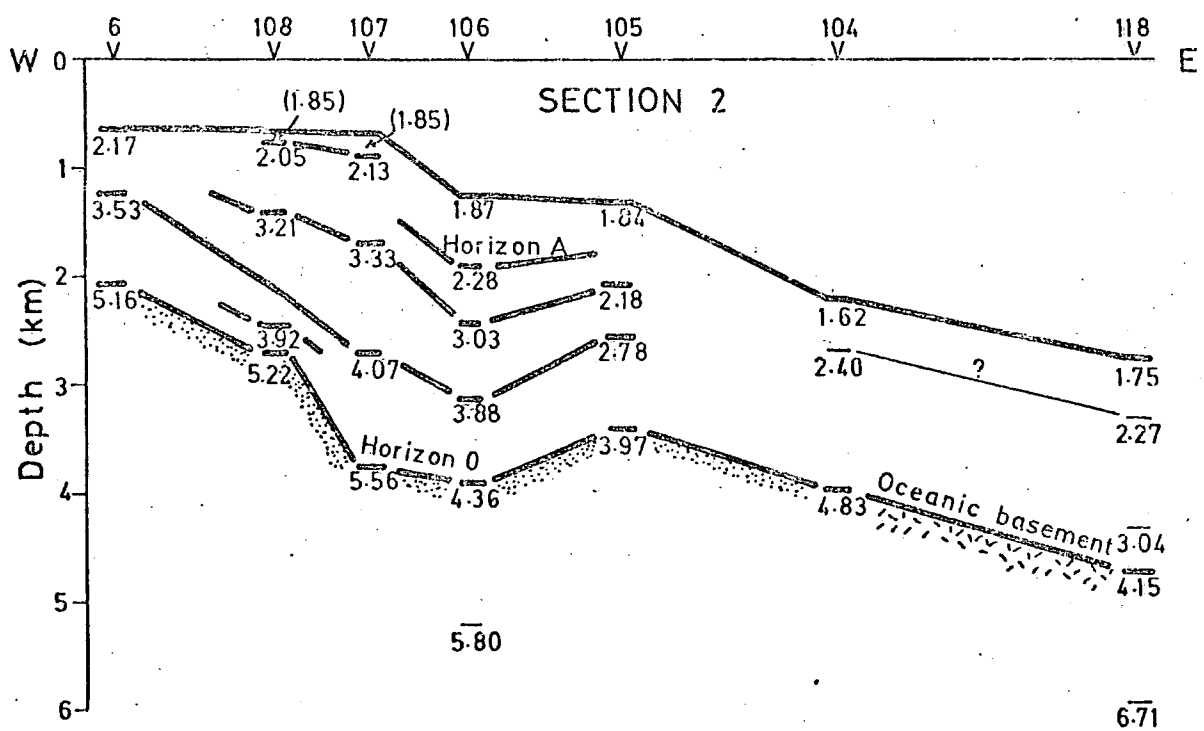
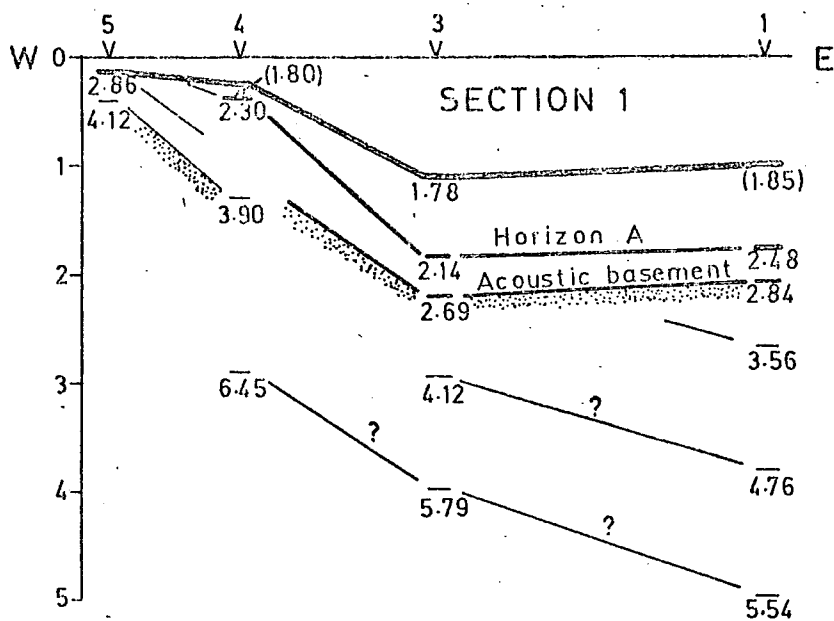


Figure 1.12

correlations in Figure 1.12 were derived as follows. Two way times computed from refractor depths and velocities for each profile were compared with the nearest seismic reflection profiles from the Capan 1, University of Durham, 1977 and University of Bergen, 1978 surveys. In most cases there was quite a good correspondence between refractors and interfaces visible on the reflection seismic sections. This is illustrated in Figure 1.9 where the two way times corresponding to profiles near the Capan 1 lines are plotted on the reflection seismic sections. Important interfaces which correspond to refractors are Horizon A, two prominent reflectors within sequence 2, Horizon 0 and oceanic basement. Note that the velocities and depths from profile 105 are consistently small but there is a good correlation between two way times and reflectors (possibly this profile was shot down-dip).

Seismic structure section 1 crosses the Jan Mayen bank and will be discussed in Section 1.7. The most important observation to be made concerning sections 2 and 3 is that a refractor underlies Horizon 0 in only two instances (106 and 9) so that no pre-Tertiary surface is continuously mapped. This is contrary to the inference of Talwani and Eldholm (1977) who stated that "the R.V. Vema sonobuoy profiles showed great thicknesses of sediments and layer velocities similar to those obtained on the Norwegian continental margin". Amplifying this statement Talwani (1978) suggested that velocities of 3.1 km s^{-1} and over might represent Mesozoic sediments.

The velocities immediately below the sea bed (circa 1.8 km s^{-1}) are consistent with the interval velocities inferred for sequence 1 ($1.7\text{--}2.0 \text{ km s}^{-1}$) by Gairaud et al. (1978), although not with the DSDP Leg 38 results, which indicate velocities of 1.5 to 1.6 km s^{-1} in the surficial layers throughout the Norwegian-Greenland Sea. Sequence 2 shows refraction velocities of 2.2 to 4.1 km s^{-1} , compared with average interval velocities ranging from 2.2 to 3.3 km s^{-1} with extremes of 3.6 km s^{-1} near the base of the sequence, and with sonic velocities of 2.2 km s^{-1} measured at the top of the sequence in DSDP holes 346 to 349. The mean

velocity of the refractor immediately above Horizon 0 is 3.78 km s^{-1} with a standard deviation of 0.30 km s^{-1} (sonobuoys 102, 106, 107, 108, 6, 8, 9, 10). This value is consistent with the higher interval velocity determinations, bearing in mind that refraction velocities are often 10% to 15% higher than interval velocities (Sherrif, 1973).

The velocity values derived for sequence 2 are not abnormally high for Tertiary sediments, in fact they are consistent with an Eocene age, as the empirical relationships of Faust (1951) indicate that Eocene sandstones and shales buried to depths of 2 km and 4 km have interval velocities of about 3.05 km s^{-1} and 3.87 km s^{-1} respectively. A fuller discussion of velocity determinations for the sediments of the Jan Mayen Ridge is given in Section 3.2.

The mean refraction velocity of the crust below Horizon 0 is 5.22 km s^{-1} with a standard deviation of 0.42 km s^{-1} (for the same sonobuoys as above). As will be shown in Section 3.2 the basement velocities are high, relative to depth, compared with the velocities in sequence 2. This supports the general correlation scheme which is shown.

Gairaud et al. (1978) obtained interval velocities of about 4.5 km s^{-1} in regions where reflectors are present below Horizon 0. A variety of stratified rock types could exhibit velocities of this order, including old or well lithified sandstones or shales (for example Carboniferous sandstones and shales at 4 km depth have interval velocities of 4.5 km s^{-1} ; Faust (1951)), limestones, plateau basalts or perhaps sediments intruded by igneous sheets. The deep refractors, which lie about 0.6 s below Horizon 0, cannot be correlated with any specific reflectors. In some places reflectors are visible to 0.9 s (2 km) below Horizon 0.

Summary

In this Section a large amount of detailed information concerning the Jan Mayen Ridge has been discussed. The main points are as follows. Firstly the known constraints on the age of Horizon A were discussed and it was shown that the

unconformity is bracketted by anomalies 7 and 17. Following Johnson and Heezen (1967), Talwani and Eldholm (1977) consider that the formation of the unconformity was associated with the shift of axis from the Norway Basin. In Section 7.2 an alternative possibility will be suggested, which is also consistent with the knowledge of the age of the unconformity. It was also noted that the drilling results do not conclusively establish the continental nature of the Jan Mayen Ridge. However multichannel seismic data suggest a continental origin for the crust underlying the Jan Mayen Ridge in two ways. Firstly the reflection characteristics of the basement beneath the ridge are not typically "oceanic". Secondly, the data suggest that the crust beneath the ridge is older than anomaly 24B, from which it can be inferred that it is probably continental. The seismic refraction data, which have hitherto been cited in favour of a continental origin, were shown not to provide very convincing evidence for or against the continental crust hypothesis.

In spite of the fact that the evidence so far presented is not as definite as one might desire, a continental origin for the Jan Mayen Ridge will be assumed as a working hypothesis for the remainder of this study and the prime indicator of continental crust will be taken to be Horizon 0 (following Gairaud et al., (1978)), even though the nature of the rocks underlying this horizon is not known for certain.

1.7 The Jan Mayen bank

The nature of the crust underlying the Jan Mayen bank (north of 70.5°N) is uncertain. The seismic reflection profile 3/78 of Sundvor et al. (1979) shows an opaque reflector at about 1 s depth below the eastern flank. Beneath the bank itself water bottom multiples obscure deeper reflectors, but it is likely that the opaque layer is fairly shallow. Correlation with seismic refraction data (Figure 1.12) shows deeper layers at depth below acoustic basement. It is possible that Horizon 0, if it exists in this region, is one of these deeper layers, and that acoustic basement represents a later extrusive basalt layer. If the identification of Horizon A (after Sundvor et al., 1978) is

correct, then acoustic basement must predate this horizon. Gronlie et al. (1979) have suggested that the bank is underlain by oceanic crust, on the basis of the high amplitude magnetic anomalies which are observed over it. The aeromagnetic data of Navrestad and Jorgensen (1979) show short wavelength, disturbed magnetic anomalies around Jan Mayen Island, roughly within the area bounded by the 200 fathom contour. These anomalies may be primarily due to shallow intrasedimentary basalts derived from fairly recent volcanism. Over the eastern flank of the ridge, high amplitude positive anomalies persist southward to 70°N , where there is a sharp ESE trending gradient bounding the smooth zone further to the south. This gradient coincides with the westward extension of the central Jan Mayen fracture zone, and it is suggested here that if there is any major change in the structure of the Jan Mayen Ridge, it probably occurs at this latitude.

1.8 The southern ridge complex

Six main basement ridges constitute the Jan Mayen southern ridge complex. These ridges are mantled by quite thick sediment but are still visible on bathymetric charts. They are more clearly delineated by free air gravity highs. In Figure 1.13, a reproduction of the detailed anomaly chart of Gronlie et al. (1979), the positions of ridges 1 to 6 are shown, slightly modified after Talwani and Eldholm (1977).

All six basement ridges are visible on single channel seismic reflection records. Characteristic of the region is an opaque acoustic layer similar to that observed in the central Iceland Plateau (Section 1.4). Talwani and Eldholm consider that ridges 4 to 6 interrupt the opaque layer to form "holes" or "windows", whereas ridges 1 to 3 are formed of acoustically opaque material. Accordingly, they have proposed that ridges 4 to 6 form a southward continuation of the Jan Mayen Ridge, sensu stricto, and are similarly composed of continental crust and that the region occupied by ridges 1 to 3 is underlain by oceanic crust, possibly of an anomalous nature, which may have formed during the period of fan-shaped spreading in the Norway Basin. Note that

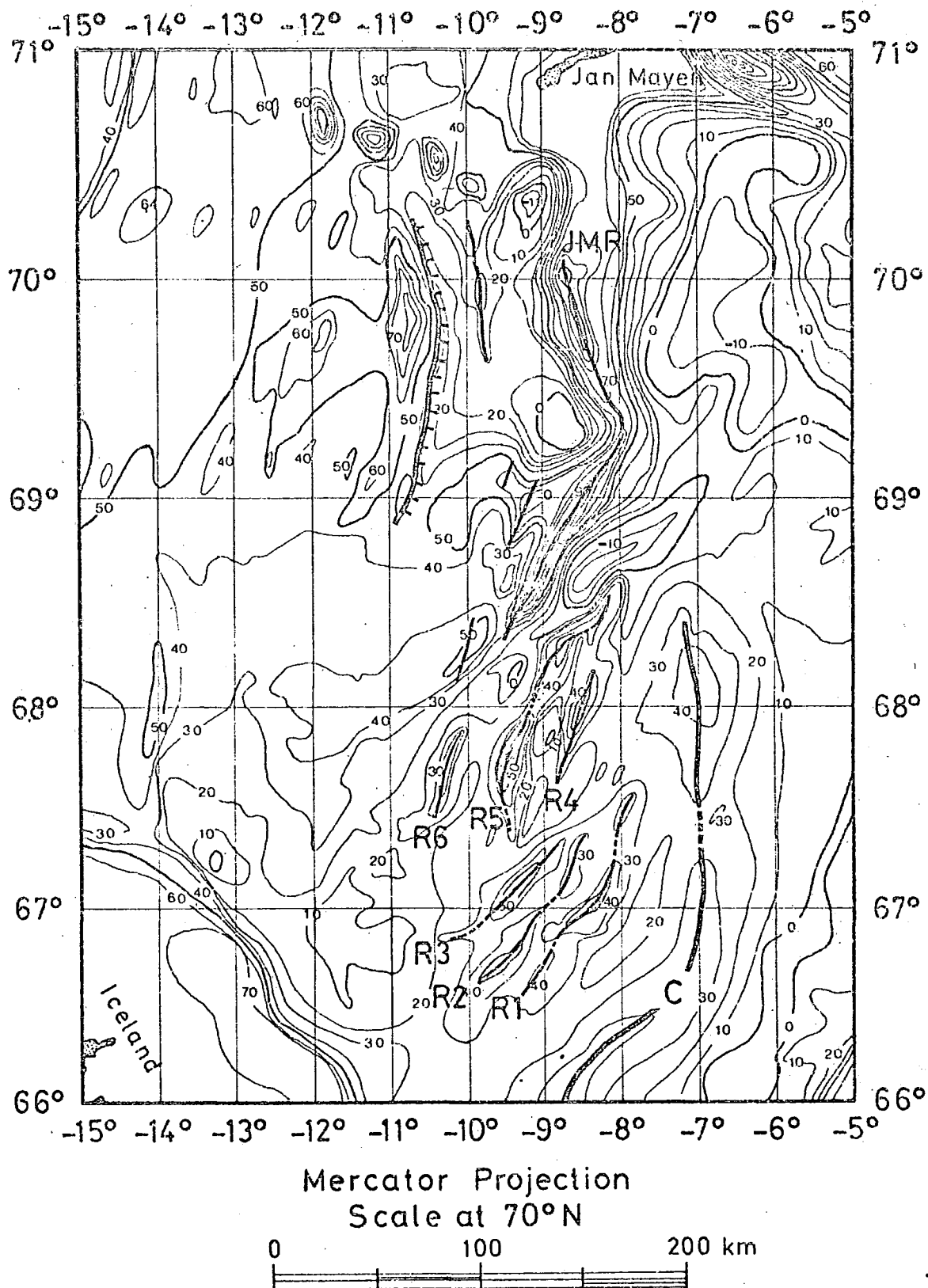


Figure 1.13 Free air gravity anomalies over the Jan Mayen block, redrawn from Gronlie et al. (1979). International Reference Ellipsoid (Flattening 1/297). Contour interval 10 mgal. Solid lines indicate linear positive anomalies which, with the exception of "C", mark buried basement ridges (Talwani and Eldholm 1977). The main ridges comprising the southern ridge complex are labelled R1 to R6. Also marked is the position of the scarp bounding the Jan Mayen basin.

Figure 1.14 Location of seismic reflection profiles over the Jan Mayen Block as summarised by Gairaud et al. (1978, Figure 11). Dotted, dashed and solid lines are used to distinguish each composite profile. Also shown (bold, dashed) are SHACK977 profiles, with each interval representing 10 km. Basement ridges (Figure 1.13) are shown dotted.

Figure 1.15 Single channel seismic sections over the Jan Mayen block, corresponding to the profiles in Figure 1.14. Redrawn with modifications from Gairaud et al. (1978, Figure 11). The sections are composed of R.V. Vema (Talwani and Eldholm 1977) and Capan 1 single channel records as indicated. The sections are aligned along 8°W . Note that the horizontal scale is variable. Reference should be made to Figure 1.14 to establish the scale of each segment. Crossings with SHACK977 profiles are indicated in the manner S1, with line 24 in the manner 1.

In the style of Talwani and Eldholm (1977) opaque acoustic basement is indicated in black, with "holes" or "windows" in the region of the Jan Mayen block. The various basement ridges are identified where possible, (R1 to R6) and the median axis of crustal extension (see text) is also marked (M).

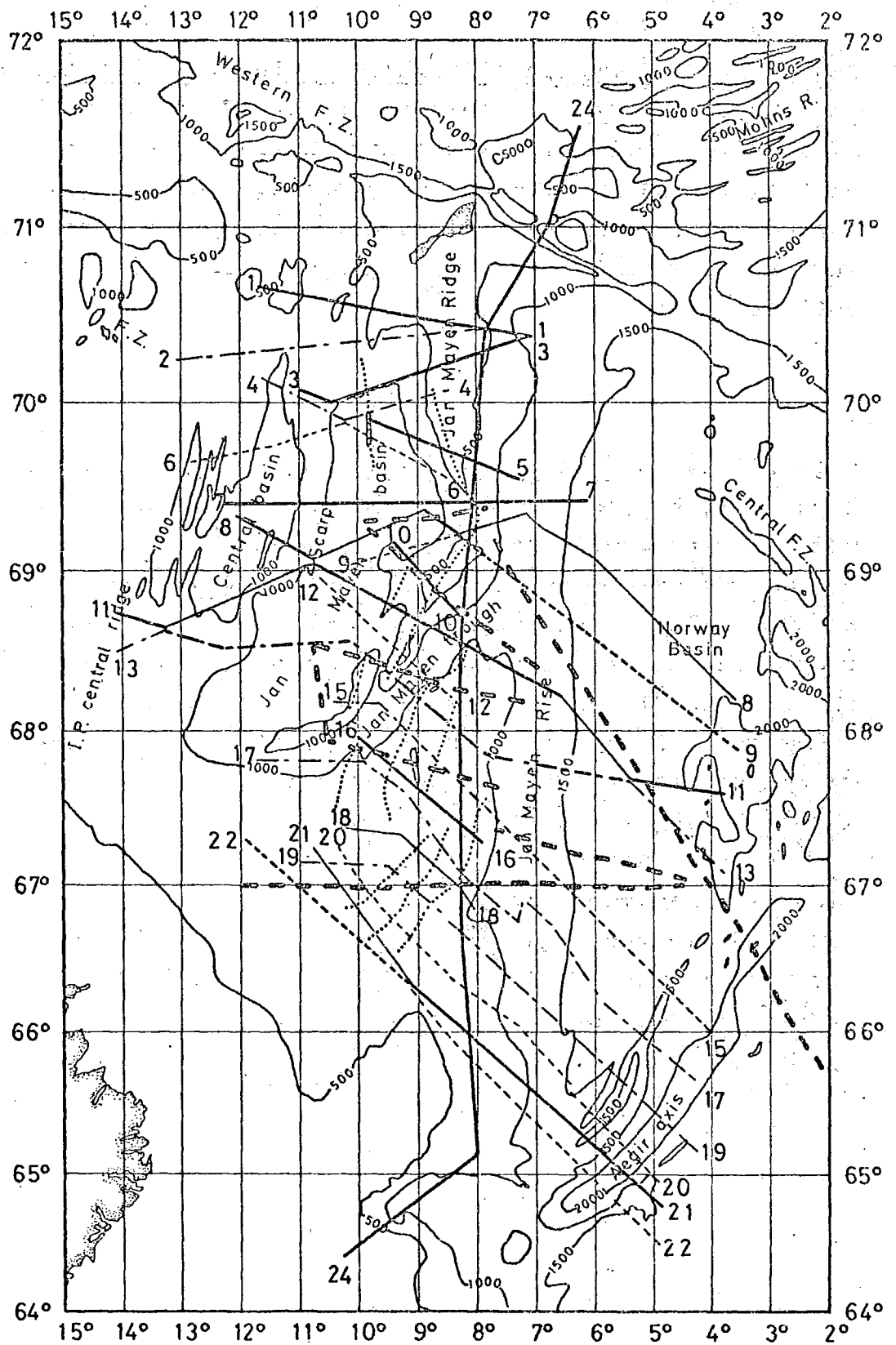


Figure 1.14

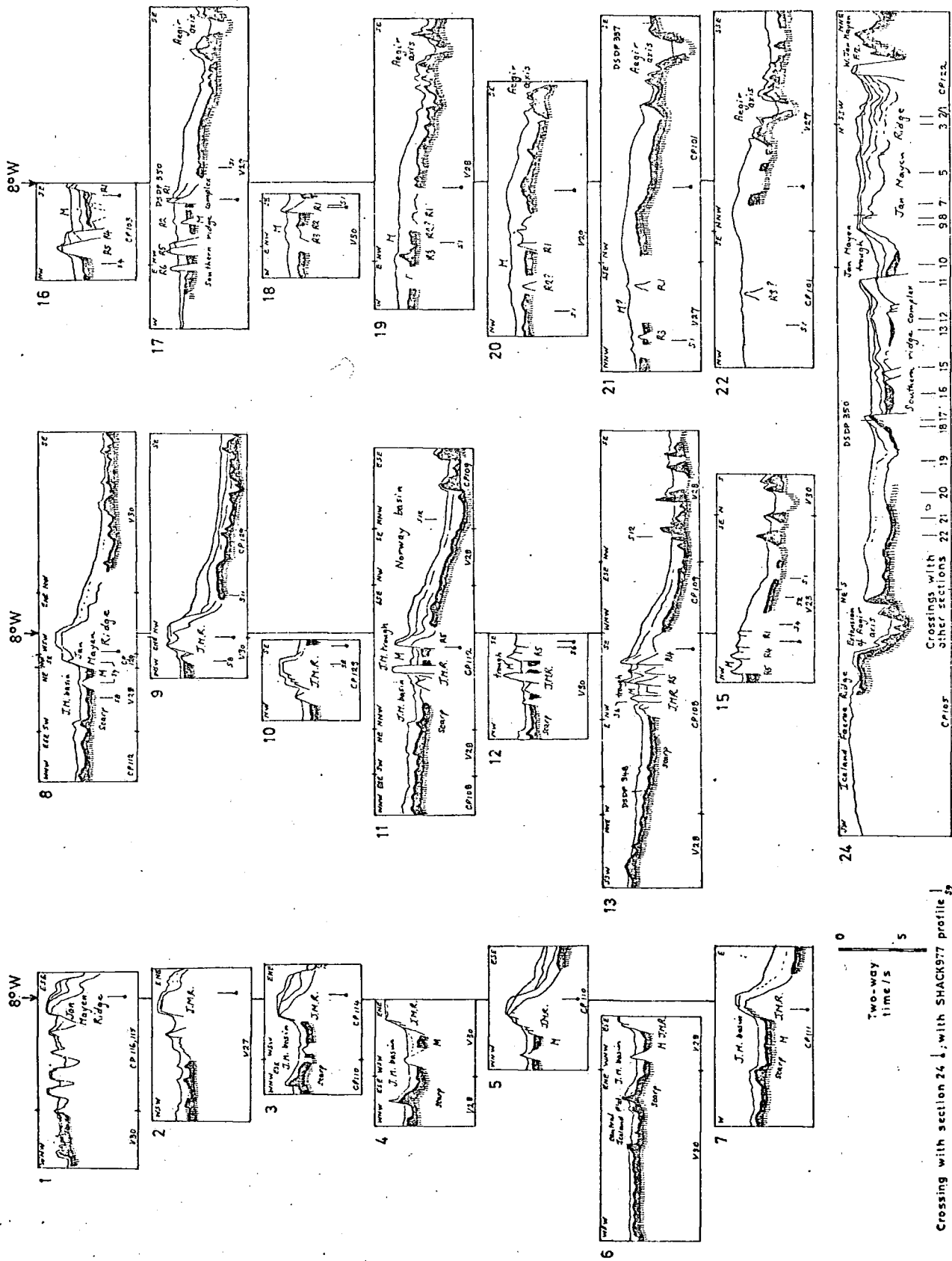


Figure 1.15

"windows" in the single channel seismic records generally occur where Horizon 0 is visible in the multichannel records.

A more detailed picture, differing in some important respects, has been obtained by Gairaud et al. (1978) using the multichannel profiles of the Capan 1 survey. In Figures 1.9 and 1.10 lines CP112, 108, 103 and 105 cross parts of the southern ridge complex. In section CP112 the essential features of ridge 5 are seen to be similar to those of the Jan Mayen Ridge further north, with the same important horizons being visible and it can be inferred that continental crust underlies Horizon 0 here also. The Jan Mayen trough, which separates the ridges, is believed by Gairaud et al. to be a graben floored by extrusive basalt. The sediment thickness within the trough is only slightly greater than that in the central Iceland Plateau and it has been suggested that rifting and volcanism was associated in time with the separation of the Jan Mayen continental fragment from Greenland.

Further south in section CP108 a broader expanse of continental crust is evident east of the Jan Mayen trough, and it is dissected into a number of faulted and antithetically tilted blocks, the most prominent of which form ridges 4 and 5. The faulting appears to have taken place after the deposition of sequence 2 in this region.

The structure revealed in line CP103, about 60 km to the southwest, is quite different. The blocks associated with ridge 5 are here downfaulted to the east and tilted to the west, and lie at the boundary of what appears to be another graben, floored by basalt. Ridge 1 forms a prominent horst block to the east of this graben, and is tilted away from the graben. Gairaud et al. believe that the ridges 1 to 3, although covered with a basaltic layer, which renders them "opaque" on single channel profiles, have a similar deep structure to ridges 4 to 6 and the Jan Mayen Ridge. This will be elaborated below.

Gairaud et al. (1978) noted that sections CP108 and 103 differ if they are aligned as in Figure 1.9 (that is in

their longitudinal relationship) but that if CP103 is shifted about 50 km northwest there is a good correspondence between the Jan Mayen trough and the rift in CP103 and between the faulted blocks on either side of the respective rifts. They therefore proposed the existence of a southeasterly trending sinistral transcurrent fault (the JAG fault) between the two sections. On the north-south line CP105 (Figure 1.10) the crossing of the proposed fault corresponds to a zone of faulting. Gairaud et al. furthermore proposed that the fault continues into the Norway Basin, offsetting a basement ridge and the Aegir axis at about 66.5°N . This part of the hypothesis can be discounted for the following reasons. Firstly, as described in Section 1.3, the postulated offset of the Aegir axis is based on a mis-identification of the axis north of 66.8°N . Secondly the pattern of magnetic anomalies (see Chapter 5) precludes the possibility of the required offset in the western Norway Basin. Thus, if the fault is present, it is restricted to the continental crust of the Jan Mayen block.

There is, however, no strong evidence for the existence of the fault, other than that based purely on a comparison of profiles CP108 and 103. Both gravity anomalies and weak magnetic lineations (Vogt et al., 1980) appear to cross the proposed fault without interruption. A detailed study of all available single channel data also failed to provide support for the hypothesis. It is, therefore suggested that the JAG fault does not exist. With this interpretation the ridge marked 5 in CP103 is thought to be a continuation of ridge 5 in CP108, as it would appear from gravity and bathymetry, rather than an offset equivalent of the Jan Mayen Ridge as proposed by Gairaud et al. The change from westward downfaulting and eastward tilting to eastward downfaulting and westward tilting is no proof of discontinuity. It is observed that the same progression takes place along the Jan Mayen Ridge sensu stricto from CP129 to CP108. This rotation of horst blocks may be a consequence of an eastward stepping of the median axis of crustal downfaulting, as one moves southward down the Jan Mayen block. Thus the rift in CP103 is an en echelon equivalent of the Jan Mayen trough which develops as the trough dies out to the southwest and it is a

tectonic axis, rather than crust itself, which is sinistrally offset. Between these profiles the median axis occupies an intermediate position, between ridges 5 and 4 as can be seen in SHACK977 profile 6 (Section 3.1).

Figure 1.15 shows a compilation of single channel seismic reflection profiles, redrawn and partially re-interpreted from Gairaud et al. (1978). Note that the horizontal scale is variable - reference should be made to Figure 1.13 to establish the approximate scale of any particular segment of a profile. The median axis of crustal downfaulting has been identified where possible, on the basis of maximum depression and also observations of the tilts of crustal blocks, which are expected to be away from the axis. The easterly stepwise transgression described above is evident, going south from profile 10 to profile 17. Further south the exact nature of the median axis is uncertain. It would appear (from topography, gravity and reflection profiles) that ridge 1 remains the eastern boundary of the rift and is consistently tilted to the east. Similarly ridge 3 appears to form the western boundary and is tilted to the west, while ridge 2 has a central position.

Gairaud et al. position the eastern boundary of the Jan Mayen microcontinental fragment about 30 km east of the crest of ridge 1, mainly on the basis of a change in the reflector characteristics of the acoustic basement. They consider that the opaque basement overlying ridge 1 is in fact "continental" basalt rather than oceanic basement. On line CP105 (Figure 1.10) what is interpreted as Horizon 0 is seen underlying about 0.8 km of basalt (assuming a velocity of 4 km s^{-1}). DSDP hole 350 was drilled through the sedimentary sequence in this vicinity, reaching basaltic tuff then fresh basalt. The presence of both Isthmolithus recurvus and Reticulofenestrata umbilica in the oldest sediments suggest an age of about 41 Ma, slightly older than the boundary between the zones defined by these nanno-plankton and corresponding to anomaly 18 time (Muller, 1976; Hailwood et al., 1979). This age is consistent with the radiometric determination. A range from 33.5 to 50 Ma was measured, with the three most concordant determinations giving a

preferred age of 41.2 ± 10 Ma (Kharin et al., 1976). These ages together seem to rule out the inference of Gairaud et al. (1978) that the basalt overlying ridge 1 was extruded coevally with the plateau basalts of the Blosseville Coast, East Greenland, which were erupted prior to anomaly 24 time at the initiation of sea floor spreading in the northeast Atlantic (Section 6.2). Horizon A, in hole 350, corresponds to a change, at about 225 m depth, from unconsolidated low density clays and muds to lithified mudstones, claystones and limestones, with proximal turbidites well developed towards the base. The rocks in the vicinity of Horizon A are of Oligocene age but data are too sparse to argue for or against the presence of a stratigraphic break.

1.9 The Jan Mayen basin

The Jan Mayen basin, and the corresponding region further south, is characterised by relatively thin sediment (0.5 to 1 s) underlain by a flat acoustically opaque layer which is thought to represent basalt and which lies at about the same level as the basalt horizon which underlies the Jan Mayen trough and the rifts further south. A reasonable model is that this basalt flooded in from the west during, and perhaps after, the separation of the Jan Mayen microcontinent from the Greenland margin. Note that there must have been at least two periods of basaltic effusion in the southern area, assuming that the basaltic layer which caps ridge 1 was emplaced horizontally. The first of these (at 41 Ma) preceded the tectonic disruption of the southern ridge complex, the second postdated it.

On account of the fact that similar flat basaltic acoustic basement occurs over wide areas of the Iceland Plateau (Section 1.4) it is difficult to estimate where the likely ocean-continent boundary lies on the western margin of the Jan Mayen block. In the northern part of the Jan Mayen basin, the opaque layer is penetrated by a narrow ridge which follows the centre of the basin and is associated with a free air gravity high (Figure 1.13). On the single channel profiles (Figure 1.15, lines 3 to 6) and on CP110 this ridge appears to be composed of "continental material". Further

south, there is a line of similar peaks immediately west of the southern part of the Jan Mayen Ridge (Figure 1.15, lines 8 to 13 also CP129, CP112 and CP108). Adjacent to the southern ridge complex, there is no indication of continental crust west of ridge 6. It seems a fair assumption that these most westerly ridges represent the limit of continental crust in the Jan Mayen block and that despite the lack of recognisable anomalies, the western part of Jan Mayen basin is underlain by oceanic crust which is older than anomaly 6B. This location of the ocean-continent boundary, which is different from that proposed previously (Talwani and Eldholm (1977), Gairaud et al. (1978), see Section 6.5, Figure 6.14) is consistent with the plate tectonic synthesis presented in Chapter 7.

1.10 The western, central and eastern Jan Mayen fracture zones

In this study a three fold division of the Jan Mayen fracture zone is proposed.

The western Jan Mayen fracture zone is primarily defined by a prominent north facing scarp (of about 1 km relief) which divides the Iceland Plateau from the Greenland Basin. This scarp extends from near the Greenland shelf at about 72.4°N , 16.2°W to near Jan Mayen Island at 71.2°N , 8.4°W , with an azimuth of about $\text{N}113^{\circ}\text{E}$. An elongate ridge which extends from the western end of the scarp eastwards into the Greenland Basin is of interest as it does not parallel the magnetic anomalies identified in this vicinity by Talwani and Eldholm (1977). A conspicuous gap, with a central hole of 3600 m depth occurs in the scarp at 12°W ; according to Gronlie and Talwani (1979) this marks the terminus of the Kolbeinsey Ridge. Features corresponding to the western Jan Mayen fracture zone can not be identified with certainty in the northern Iceland Plateau zone south of the scarp although a number of bathymetric trends do parallel the scarp. East of Jan Mayen Island the western Jan Mayen fracture zone continues with a similar azimuth to about 70.4°N , 2.7°W , as a narrow ridge whose north face corresponds to the scarp of the more westerly segment. Further east this ridge diminishes in

size and curves around to 69.6°N , 1°W . This ridge divides the most northwesterly part of the Norway Basin from the flank of the Mohns Ridge.

The central Jan Mayen fracture zone is primarily defined by a nearly continuous ridge (of about 1 km relief) which runs from 69.7°N , 5°W to 67.8°N , 0.6°W . Indications of a buried ridge are present beyond both limits given. The ridge is paralleled to the southwest by a topographic depression and a more pronounced linear gravity low. The ridge is markedly arcuate, concave to the southwest.

The eastern Jan Mayen fracture zone lies northeast of the central fracture zone and is delineated by a southwesterly facing scarp extending from 69.2°N , 2°W to 67.6°N , 0.6°E . Between this scarp and the ridge of the central Jan Mayen fracture zone a prominent negative gravity anomaly indicates a buried basement trough. The continuation of the eastern Jan Mayen fracture zone beneath the southwest margin of the Voring Plateau is not known, on account of sediment cover. The eastern Jan Mayen fracture zone is not markedly arcuate.

In general terms the azimuth of the western Jan Mayen fracture zone represents the current direction of spreading in the northeast Atlantic while the azimuth of the central and eastern Jan Mayen fracture zones represents the spreading direction before the cessation of spreading in the Labrador Sea, when the motion of Greenland relative to North America allowed a more NW-SE separation between Greenland and Europe. In Chapter 7 more detailed inferences will be drawn from the fracture zone geometries.

1.11 Outline of the study

In this Chapter, published bathymetric and geophysical data over the southern Norwegian-Greenland Sea has been discussed and, in part, reinterpreted. There are major uncertainties concerning

- (i) the structure and evolution of the Jan Mayen block

(ii) the identification of magnetic anomalies in the Norway Basin

(iii) the location of the complementary spreading axis and the overall kinematic development of the area.

The remainder of this study concentrates on these problems.

The SHACK977 cruise, which is described in Chapter 2, included a survey of the southern part of the Jan Mayen block and a traverse across the Norway Basin, which at its eastern end, fills a gap in the Durham University 1976 Norwegian margin survey (SHACK476, Robinson 1980). The Jan Mayen survey was intended to elucidate the deep structure of the region by multichannel seismic profiling. At the time that the survey was planned, no multichannel data had been published and it was anticipated that the Cepen 1 results would remain confidential to CNEXO. The emphasis of the present study changed when it became clear that there would be considerable delays in the processing of the multichannel data. The necessary processing hardware did not arrive until October 1978, following which an entire set of processing programs had to be written and implemented by members of the department. A short trial section was processed in December 1979 but this was followed by severe hardware failure which put the system out of action for nearly six months so that no further processing was possible. Fortunately, the publication of the Cepen 1 results allowed the SHACK977 single channel monitor records to be interpreted with much more confidence than would otherwise have been possible.

In Chapter 3, interpreted line drawings of the Jan Mayen and Norway Basin monitor records are presented and discussed. No multichannel control is available for most of the Norway Basin profile, but the single channel record for this profile is clearer than the Jan Mayen records because the water depths are greater. Apart from revealing a major slump structure which has hitherto been undetected, the SHACK977 seismic data has, to date, confirmed rather than extended the knowledge of the Jan Mayen block which is afforded by the Cepen 1 results. Its main use, in the present study, is to provide depth to basement values which can be used for

purposes of gravity and magnetic interpretation. Towards this end, in the second part of Chapter 3, velocity data for the sediments of the southern Norwegian-Greenland Sea are collated and velocity functions are derived to allow conversion of reflection times to depth. Particularly in the vicinity of the Jan Mayen Ridge, an average sediment velocity of 2.0 km s^{-1} , which is commonly assumed, is not appropriate. In the final part of the chapter, for the purposes of documentation, a description is given of the Durham University Reflection Seismic Processing System and the author's contribution towards software development.

Chapter 4 is concerned with an interpretation of the gravity data acquired during the SHACK977 cruise. The method used is the frequency domain technique of Oldenburg (1974) which is described at the beginning of the Chapter, followed by a discussion of density data for the southern Norwegian-Greenland Sea. In the interpretations, base level control (subject to uncertainty concerning mantle density variations) is provided by the Norway Basin profile, which is assumed to cross "standard" ocean crust. The main aim of the Jan Mayen interpretations is to establish whether or not low density pre-Tertiary sedimentary basins exist beneath the Jan Mayen block. The evidence from the Durham data, which is unfortunately accrued in a fairly indirect manner (because of the imprecision of the monitor records), indicates that the observed anomalies can be accounted for by variations in water depth, Tertiary sediment thickness and long wavelength variations on the Moho. This is in contradistinction with the results of Gronlie et al. (1979), whose interpretations (Figure 4.2) indicate substantial thicknesses of pre-Tertiary sediments. According to the present interpretation, the gravity data do not prove the continental origin of the Jan Mayen block. This result is consistent with the reinterpretation of the refraction sections (Section 1.6) which indicate high velocities (and hence densities) beneath Horizon 0.

In Chapter 5 an interpretation of the magnetic anomalies along the Norway Basin profile is given using the direct inversion method of Parker and Huestis (1974). This

interpretation was originally conceived as an aid to anomaly identification, but in retrospect, the computed magnetisation profile is probably no more useful in this respect than the original anomaly profile, although it is used in Chapter 6 to give a spreading curve for the eastern Norway Basin. However some interesting insights into the nature of the magnetised layer in the Norway Basin are obtained. The magnetic anomalies along the Jan Mayen profiles were interpreted using the same technique, but because the technique is not really appropriate to areas which are not underlain by oceanic crust these interpretations are not presented. They do show, however, that the basaltic layers of the southern ridge complex are quite highly magnetised, and exhibit reversals indicating that they were extruded over a number of epochs.

Chapter 6 is concerned primarily with the correlation and identification of magnetic anomalies in the southern Norwegian-Greenland Sea, with emphasis on the Norway Basin, but the related topic of ocean-continent boundaries is also dealt with. A revision of Talwani and Eldholm's identifications for the Norway Basin is proposed, but their fan shaped spreading hypothesis is confirmed. These results are based on a new compilation of marine magnetic anomaly data which included SHACK476 and SHACK977 profiles as well as a variety of published data. As a preface to the Norway Basin studies, the introductory sections to Chapter 6 deal with the overall magnetic anomaly pattern in the northeast Atlantic and the evidence for an abrupt onset of spreading shortly before anomaly 24 time.

In Chapter 7 the fan shaped spreading in the Norway Basin is described in terms of rotation about a near pole. It is proposed that the complementary spreading took place about the newly formed Kolbeinsey axis and plate tectonic reconstructions are presented which are consistent with this hypothesis. The implications of the scheme with respect to the Jan Mayen block are described and finally, a hypothesis is presented which purports to explain why the Jan Mayen Ridge was detached.

In the present Chapter, very little mention has been

made of the areas which are marginal to the southern Norwegian-Greenland Sea, such as the Icelandic transverse ridge and onshore Greenland. These areas are discussed briefly where appropriate in the remainder of the text. This study is not explicitly concerned with the "Iceland phenomenon" (Vogt 1974) even though it has been suggested (for example Vogt et al., 1980) that the complexities of the southern Norwegian-Greenland Sea, such as the shift in the spreading axis, are due to the effects of this hot spot phenomenon. An attempt is made to deal with the area within the framework of plate tectonic principles which have been applied to other areas of the globe and it is believed that the final results of Chapter 7 justify this approach.

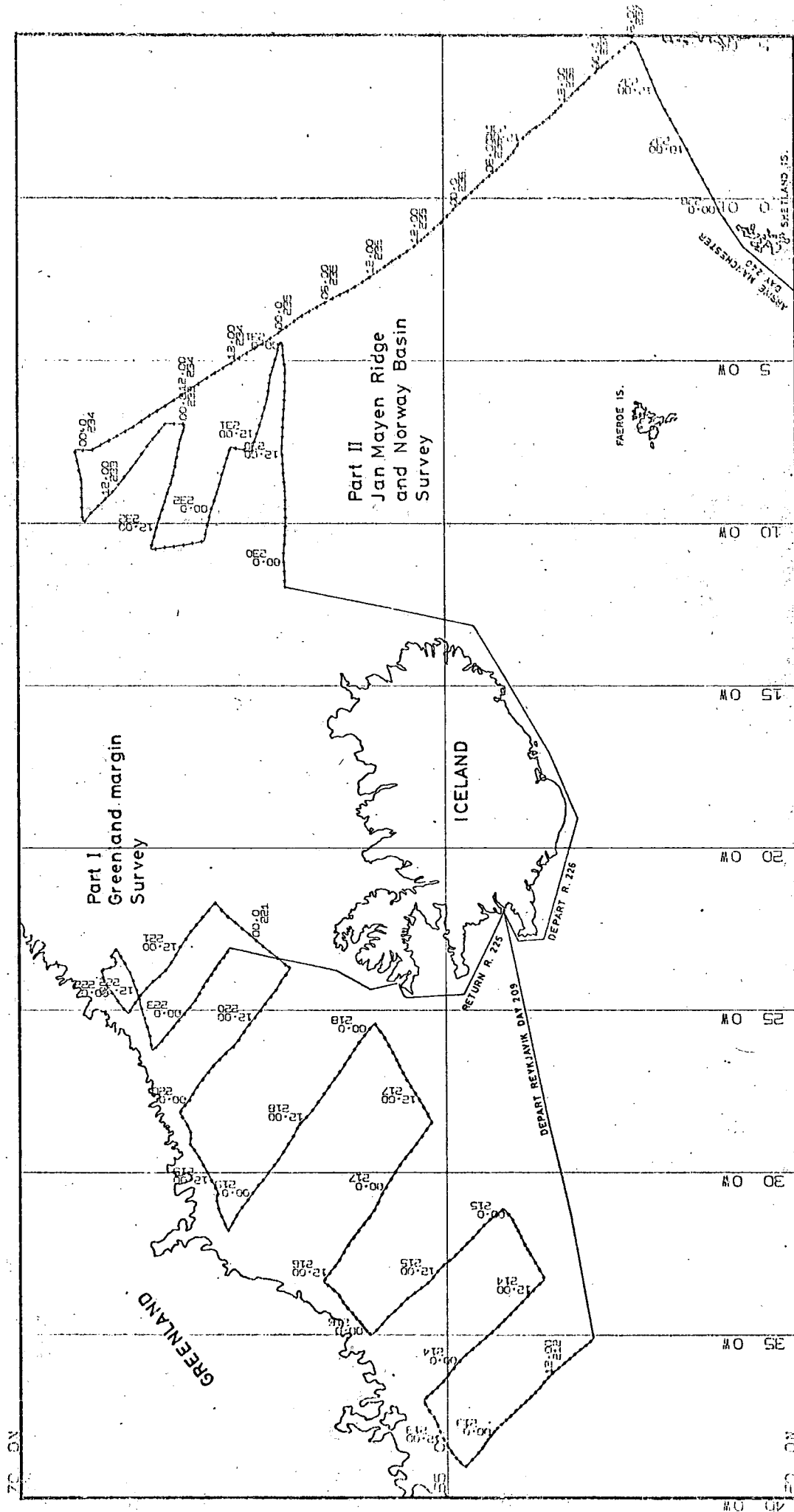
Chapter 2

THE DURHAM UNIVERSITY 1977 CRUISE

During the Durham University 1977 marine geophysical cruise in the northeast Atlantic (R.R.S. Shackleton, 9/77; senior scientist J.H. Peacock) seismic reflection and refraction, bathymetric, gravity and magnetic data were collected in two survey areas adjacent to Iceland (Figure 2.1). Part I of the cruise covered a previously unsurveyed area of the Greenland margin and has been described in part by Roberts (1978) and in full by Armstrong (thesis to be submitted).

Part II, which is a concern of the present study, comprised a survey of the Jan Mayen block between the latitudes 67°N and 69.5°N , a traverse from the Jan Mayen rise at 69.3°N across the Norway Basin to the Norwegian coast at 62.4°N and a traverse across the Norwegian channel and northern Viking graben to the Shetland Islands. In Figure 2.2 the ship's track is shown, together with hour marks and a synopsis of navigational and geophysical techniques used. In Figure 2.3 the ship's track is shown, divided into profiles for the purposes of interpretation, together with distance marks and bathymetry (after Gronlie and Talwani, 1978). The major profiles in the Jan Mayen survey were planned to cross morphological elements at right angles with sonobuoy runs along short lines parallel to the structure. The Norway Basin traverse is approximately perpendicular to the strike of magnetic lineations and at its eastern end fills a gap in the Durham University 1976 Norwegian margin survey (Robinson, 1980). The weather was never severe enough to hinder geophysical data logging, which was accomplished as planned, although the total length of the survey was shorter than anticipated.

Gerdes (1978) has described and interpreted sonobuoy runs 1, 2 and 3. Hamilton (1978) presented a section based on the seismic reflection data along the Norway Basin traverse, together with a qualitative discussion of the



MERCATOR PROJECTION

(NATURAL SCALE AT LAT. 55)

0 100 200
 KM
 INTERNATIONAL SPEED

Figure 2.1

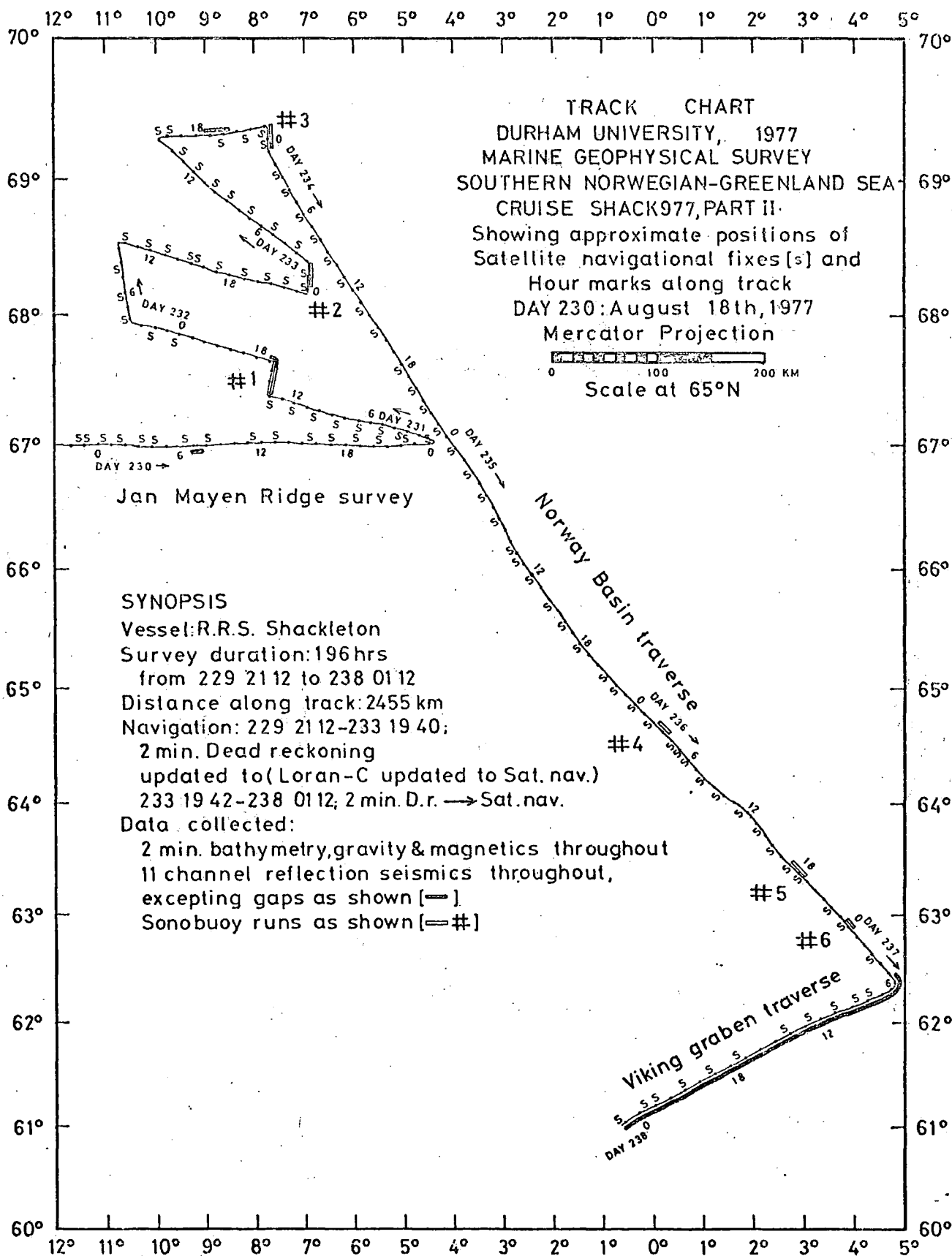


Figure 2.2

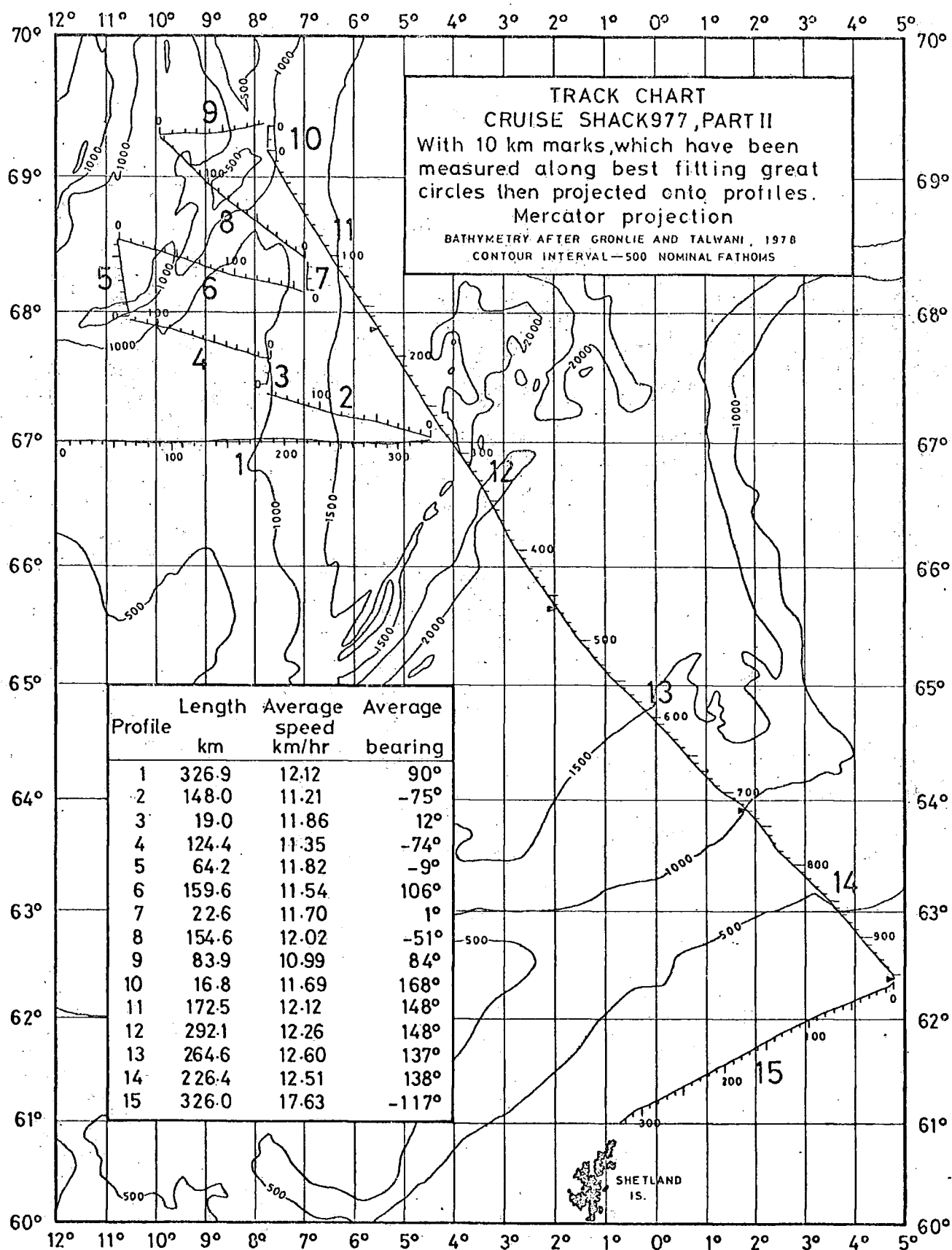


Figure 2.3

gravity data and an interpretation of sonobuoy runs 3 to 6. Arnold (1978) has interpreted the gravity data along the Viking graben traverse and Robinson (1980) has treated all of the data along the Norway Basin traverse, in conjunction with the data from the cruise Shackleton 4/76. In the present study data from lines 1 to 13 are used.

2.1 Multichannel seismic data acquisition

Figure 2.4a shows the physical set-up used in acquiring multichannel seismic data. The streamer, airguns and compressors were supplied by the Natural Environmental Research Council, through the Research Vessel Base at Barry, Glamorgan. A ship's speed of 8.6 km/hr would have provided for optimum common depth point stacking with the shot period of 21 s, which was governed by the limitations of the compressor system. A higher survey speed of 11-12 km/hr was maintained in order to increase the length of the survey, thereby diminishing the effectiveness of stacking (Section 3.4).

The hydrophone array response was recorded digitally using Durham University's Western SDS 10/10 seismic acquisition system. A block diagram of this system is given in Figure 2.4b. The general principles of operation are given in Dobrin (1976, pp 68-75). The equipment functioned well throughout the cruise. Three small unscheduled breaks in seismic acquisition (days 230, 231 and 233) were caused by airgun and tape faults.

Refraction data was collected using disposable sonobuoys (UEL seismic buoy type SB6 E4 with a hydrophone depth of 18 m). The sonobuoy response was transmitted back to the ship, recorded on channel 12 and displayed on the monitors. Normal reflection seismic acquisition continued during sonobuoy runs.

2.2 Collection and reduction of navigation, depth, gravity and magnetic data using the ship-borne computer system

An integrated system for collecting and processing

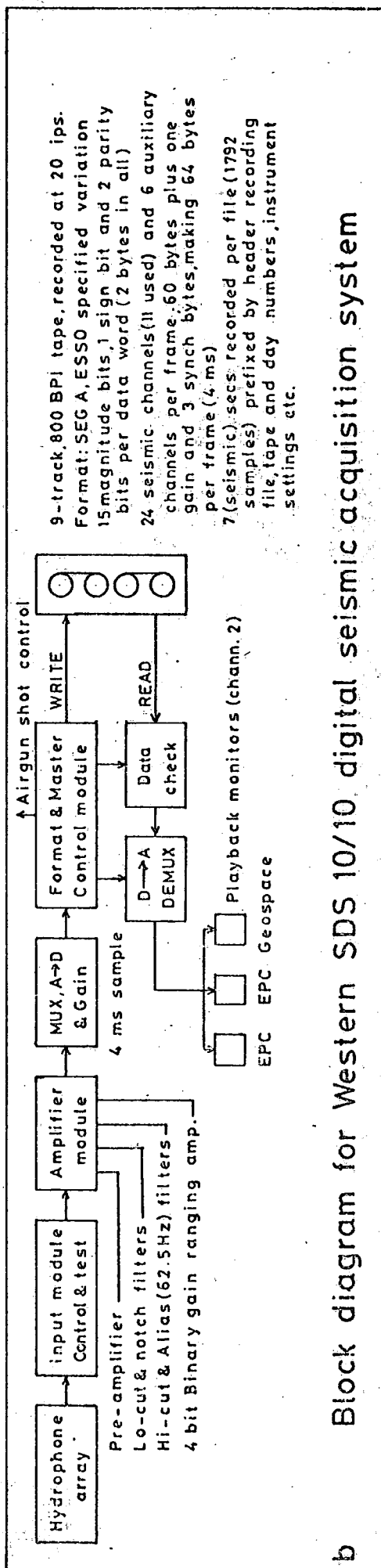
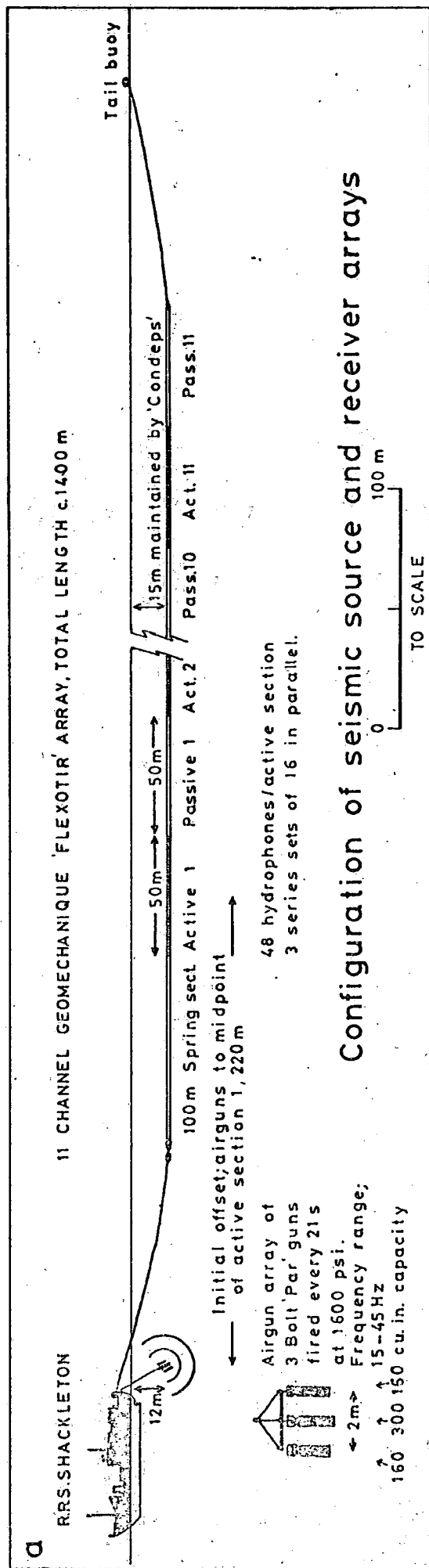


Figure 2.4

navigation, depth, gravity and magnetic data was operated independently of the seismic acquisition system. This system provided automatic logging of most parameters, followed by off-line processing on an IBM 1130 computer, which was on board during the cruise. Table 2.1 lists the equipment, which was mainly supplied by the Natural Environmental Research Council and was similar to that used on previous Durham University North Atlantic cruises (Featherstone, 1976; Robinson, 1980). The Loran-C navigational aid was unusable during the second half of the survey due to sky-wave reception, and radio interference on day 231 caused a gap in satellite navigational fixes. Minor breaks and inaccuracies in logged data (mainly in bathymetry due to the Digitrak losing track) were corrected during processing.

Figure 2.5 illustrates the data flow from acquisition through to final display and storage. Processing was carried out by Miss Doriel Jones of the Institute of Oceanographic Sciences (IOS), Mr T.L. Armstrong and the author. A malfunction of the console teletype put the computer out of action for the last part of the survey so processing was completed at R.V.B. Barry. Details of the computer software are given in the Institute of Oceanographic Sciences manuals (1974). All of the important options chosen during processing are indicated in Figure 2.5. The method of course adjustment used was as recommended in the IOS software manuals. Judging from the specifications therein the absolute positioning errors may be of the order of several hundreds of metres but the relative errors are expected to be considerably smaller. No additional smoothing of gravity and magnetic data was undertaken after resampling at two minute intervals and high frequency noise present in some of the data was removed during subsequent computer processing at Durham.

A list of raw and reduced data available at Durham University is given in Appendix A, together with copies of data profiles for the whole survey.

In the next two Sections the subjects of gravity tie-ins and temporal magnetic variations will be treated in detail,

Table 2.1 SCIENTIFIC INSTRUMENTATION

NAVIGATION	Two component electromagnetic log MICROTECHNICA gyrocompass LORAN-C hyperbolic navigational system Satellite navigational system: MAGNAVOX 702 receiver with HEWLETT-PACKARD 2100 computer Output to ASR 33 teletype
BATHYMETRY	IOS MARK 3 PRECISION ECHO SOUNDER with towed P.E.S. transducer. Output via MUFAX MS38 chart recorder and EDO-WESTERN 26/C DIGITRAK
MAGNETOMETER	VARIAN 4937 proton precession magnetometer Twin coil sensor towed 185 m astern at 10 m depth Cycle time 6 s. Digital and chart output
GRAVIMETER	LA COSTE AND ROMBERG AIR-SEA gravity meter, Model 5, No. S40 4min filter. Chart record, drum printer and digital output
DATA-LOGGER	Modified DECCA MAG-LOG system 2 RDL magnetic tape decks (Series 10500) and controller 1/2 inch, 9-track, 800 BPI, NRZI, IBM compatible tape Read-after-write output to DATA DYNAMICS 390 teletype
COMPUTER	IBM 1130 system, installed in portable container IBM 1131 CPU (8K 16 bit words) with disk drive and teletype IBM 2310 double disk drive (512K words/disk) DATA DYNAMICS fixed head disk (655K words) TEKTRONIX 4012 VDU with 4610-1 Hard copy unit 76 cm CALCOMP 1627 drum plotter (0.01 cm increment) FACIT paper tape I/O unit RDL Series 10500 magnetic tape deck

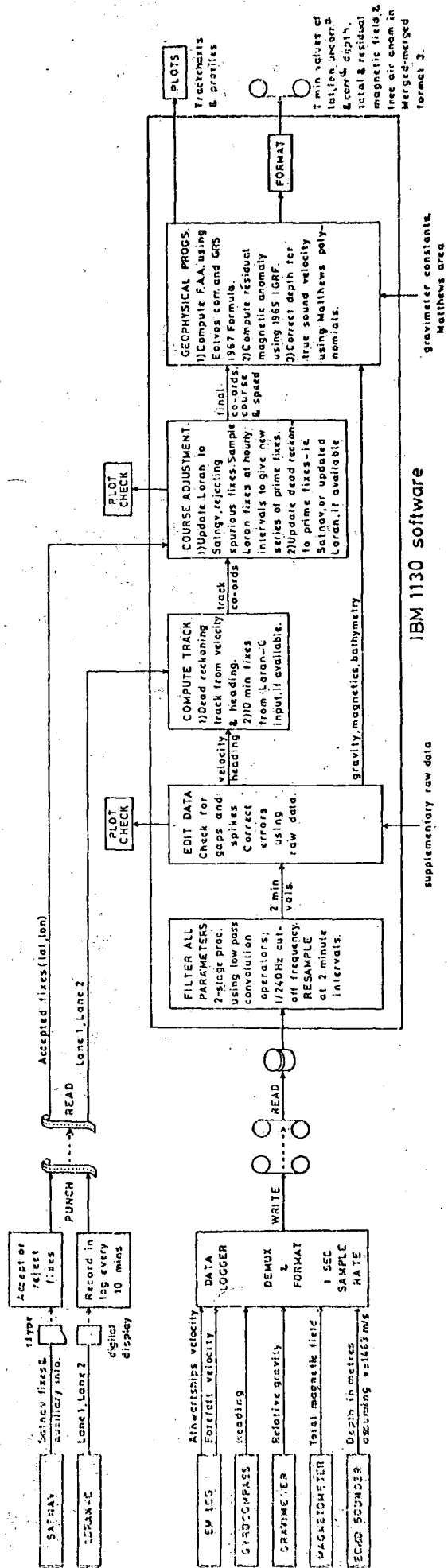


Figure 2.5

as they are important in establishing the reliability of the reduced data.

2.3 Gravity tie-ins

The La Coste-Romberg sea gravimeter is a relative instrument. The relationship between the gravity, g and observed meter reading, o is

$$g = \text{calib} * o + d$$

where the calibration constant, calib , is effectively invariant with respect to time and reading. For the meter used (S40), calib is 0.9915 mgal/scale unit. The difference d is subject to drift and must be determined by tying in to land base stations.

Prior to the start of the 1977 cruise the meter was tied in to the R.V.B. Barry base station by IOS personnel in March 1977. Tie-ins were made at Reykjavik before and after the first part of the cruise, to the station 0081 Reykjavik CH of the Icelandic network (Palmason et al. 1973). At the end of the cruise a tie was made to a temporary station at Manchester Dry Docks and the tie was later completed to the Daresbury F.B.M. of the National Gravity Reference Net (Masson-Smith et al., 1974).

Table 2.2 gives details of these tie-ins with values of d in mgal. A complication arises because the Icelandic network and the British network are not based on the same datum. The British stations are tied, through NGRN73 to the International Gravity Standardisation Network, 1971 (IGSN71) which is based, in large measure, on ten modern absolute gravity determinations (Woollard, 1979). The Icelandic network is indirectly linked into the old Potsdam system via a number of ties to European networks (for example Saxov and Spellauge, 1967). The Potsdam System was based on a single absolute gravity determination made at Potsdam, which has since been shown to be 14.0 mgal too high, relative to IGSN71. If the 1930 International Gravity Formula is used to calculate free air anomalies this error is compensated for at medium latitudes (Woollard, 1979). The 1967 Geodetic Reference System (GRS67) Formula which was used for the

Table 2.2 Details of gravity tie-ins, SHACK977

Base station	g_{base} (mgal) and datum	$g_{dock} - g_{base}$ (mgal)	Date	$g - g_{dock}$ (mgal) Ht. corr. of 0.309 mgal/m	g (mgal) abs. grav, meter room	o (meter reading in nomin. mgal)	$d = g - calib * o$ (mgal) calib=0.9915
R.V.B. Barry (Gravity laboratory internal memorandum)	981190.27 (IGSN71)		16 Mar 77 Day 74		981191.20 (IGSN71)	979312.75 (mean of 2)	10202.61 (IGSN71)
0081 Reykjavik CH (Palmason et al. 1973)	982277.36 (Potsdam) 982262.80 (IGSN71) Datum corr. -14.56 mgal	3.49 Tie established 27-28 Jul 1977 6 consecutive ties over 4 hrs using Worden 748 Less than 0.23 mgal drift	28 Jul 77 Day 209	1.13 (mean of 2)	982281.98 (Potsdam) 982267.42 (IGSN71)	980401.50 (mean of 2)	10213.89 (Potsdam) 10199.33 (IGSN71)
			13 Aug 77 Day 225	0.70 (mean of 2)	982281.55 (Potsdam) 982266.99 (IGSN71)	980401.45 (mean of 2)	10213.51 (Potsdam) 10198.95 (IGSN71)
Daresbury F.B.M. (Masson-Smith et al., 1974)	981351.80 (IGSN71)	21.13 Tie established 20 Jun 1979 3 consecutive ties over 5 hrs using L&R 453 0.11 mgal drift	30 Aug 77 Day 232	0.34	981373.27 (IGSN71)	979500.50	10198.52 (IGSN71)

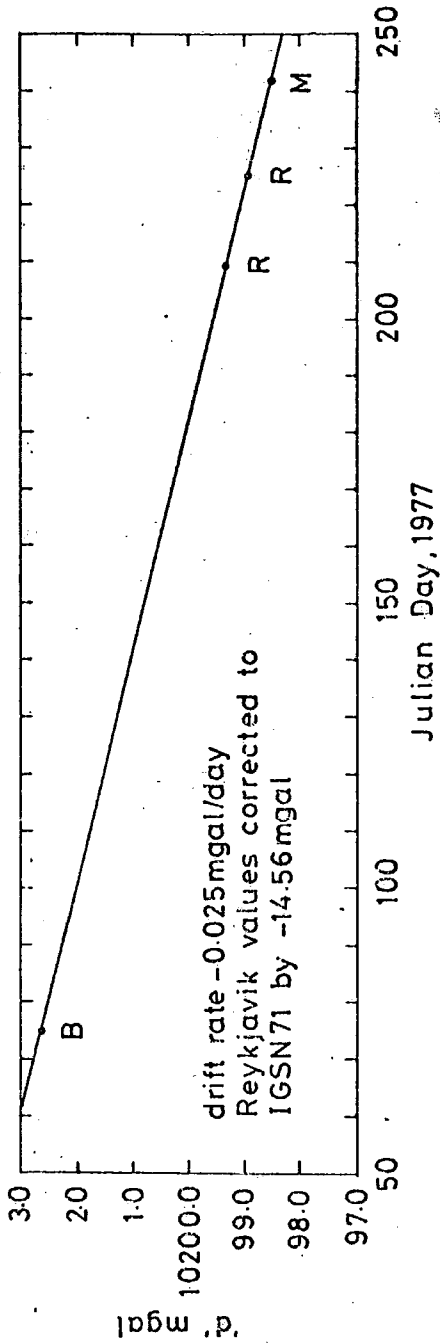


Figure 2.6 Drift curve for La Coste and Romberg Air-Sea gravity meter, No. S40.

reduction of the SHACK977 data, is on the other hand consistent with IGSN71 but not with the Potsdam System (Association Internationale Geodesique, Special Report 3, 1971). The Reykjavik values therefore need to be referred to IGSN71. Because of errors in tie-ins, older networks linked indirectly to Potsdam have datum differences with IGSN71 which may differ by up to 1 mgal from the -14.0 mgal Potsdam correction. The Icelandic network is based on a value of 982 279.86 mgal for Reykjavik A (Palmason et al., 1973). Woollard and Rose (1963) obtained a value of 982 280.00 mgal for this station and as the Woollard and Rose datum is 14.70 mgal higher than IGSN71 (Woollard, 1979) one can infer a datum difference of 14.56 mgal between the Icelandic network and IGSN71. Figure 2.6 shows that a smooth drift curve is obtained if this datum correction is applied. The total drift during the survey period was only -0.20 mgal. It is therefore satisfactory to use a constant value of d in reducing the data.

Unfortunately, when data reduction was begun on board ship these complications were not appreciated. Effectively, a value of $d = 10213.5$ mgal corresponding to the uncorrected second Reykjavik tie was input into the gravity reduction program, instead of the appropriate corrected value for the second part of the cruise, which is 10198.7 mgal. The reduced free-air anomaly values obtained of the IBM 1130 are therefore about 14.8 mgal too high. This error has been corrected during processing at Durham, but is still present in the original reduced data tape and master plots produced on the IBM 1130 computer (Appendix A).

2.4 Magnetic temporal disturbances

Magnetograms covering the period of the survey were obtained from the nearest land observatories: Tromsø, Leirvogur, Dombas and Lerwick (Figure 2.7). Six periods of quite severe magnetic disturbances were recorded simultaneously at all four stations. In Figures 2.8 and 2.9 the vertical and horizontal fields, ΔZ and ΔH , are displayed for these events together with the measured total field anomaly and the Lerwick K indices (Institute of Geological

Figure 2.7. Chart showing the locations of magnetic observatories and the positions of the R.R.S. Shackleton during periods of magnetic temporal disturbances. The lines of equal magnetic inclination (I) and declination (D) are for Epoch 1965.0 (U.S. Naval Oceanographic Office, 1966, 1965) and are therefore slightly in error (the annual change in declination is about $-5'$, in inclination less than $-1'$). The barred solid lines along ship's track indicate the periods for which magnetograms are shown in Figures 2.8 and 2.9 (Events 1 to 6) and the heavy lines indicate periods when the disturbance was quite severe.

Figures 2.8 and 2.9. Magnetograms from Leirvogur, Lerwick, Dombas and Tromso, covering periods of temporal magnetic disturbance during the survey. The original magnetograms were kindly supplied by

Dr Th. Saemundsson of the Science Institute of the University of Iceland (Leirvogur).

Dr A. Forbes of the Geomagnetism Unit of the Institute of Geological Sciences (Lerwick).

The Director of the Geophysical Institute of the University of Bergen (Dombas).

Dr S. Berger of the Institute of Mathematical and Physical Sciences at the University of Tromso (Tromso).

The horizontal field variations (H) are shown as light lines and the vertical field variations (Z) as heavy lines (note the differing scales for different observatories). Also shown are the measured total field anomaly and the Lerwick K indices (Institute of Geological Sciences 1980), each one indicated at the beginning of the three hour period to which it applies.

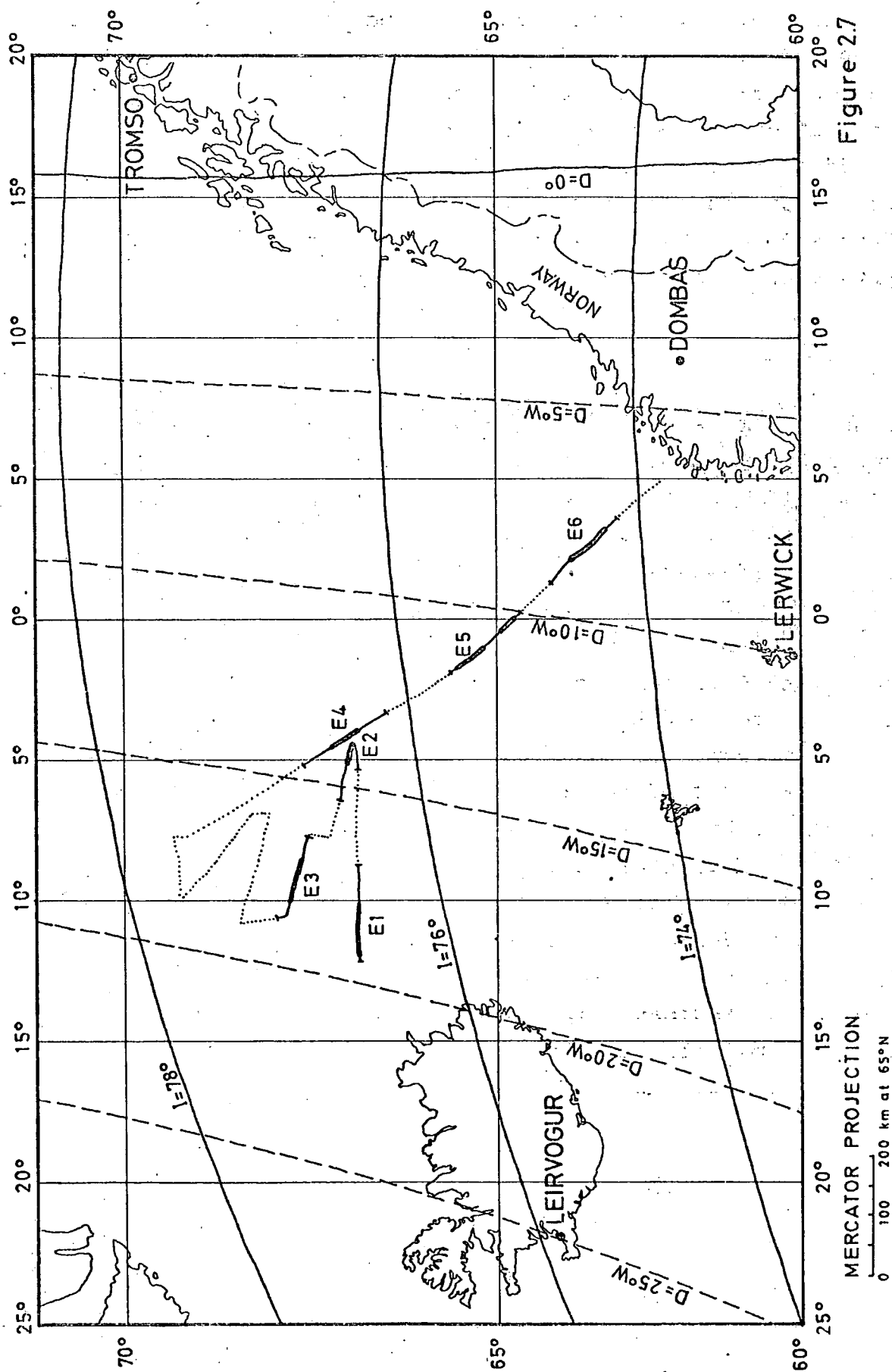
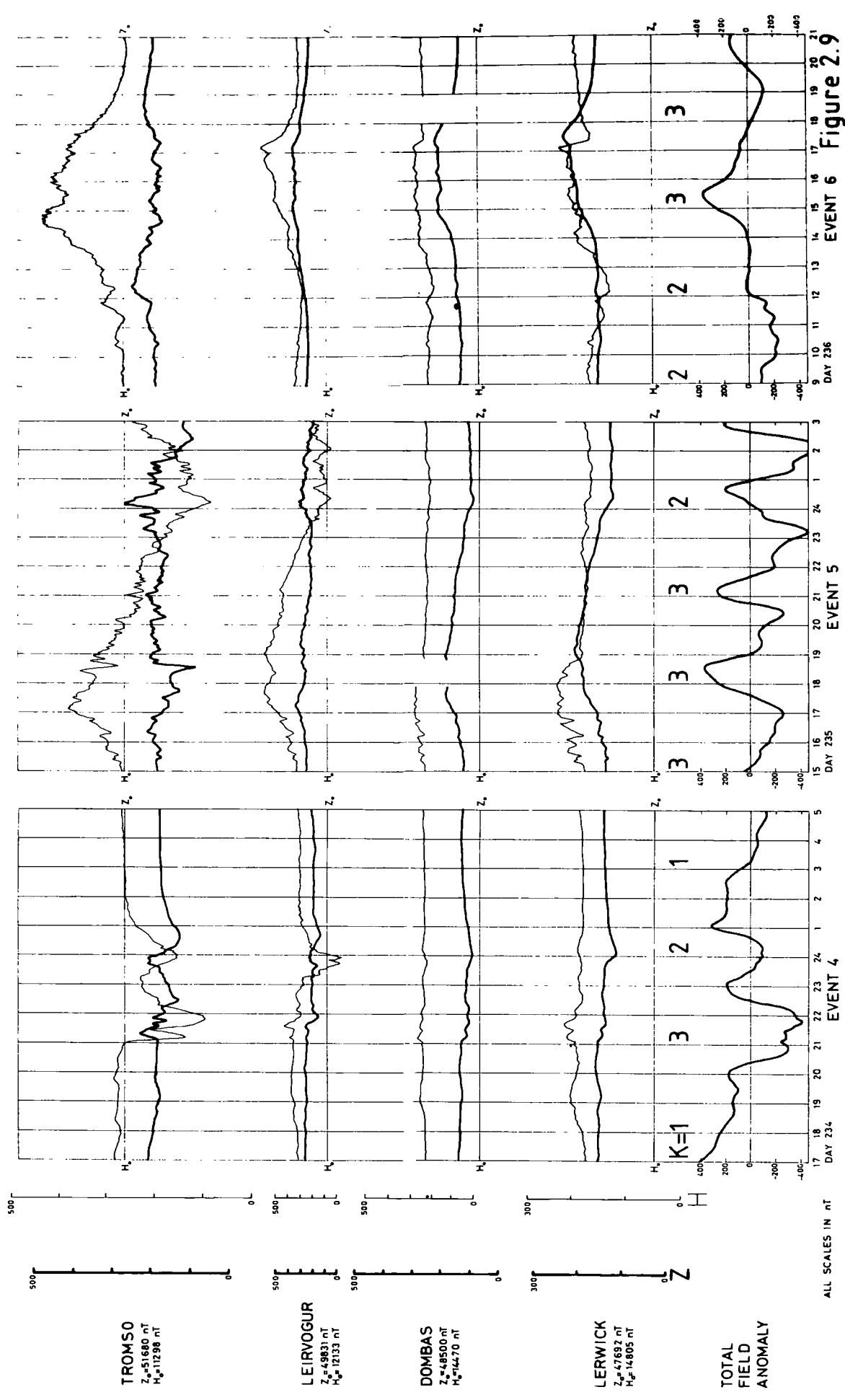


Figure 2.7



EVENT 6 Figure 2.9

EVENT 5

EVENT 4

ALL SCALES IN nT

Sciences, 1980). The location of R.R.S. Shackleton during these events is shown in Figure 2.7. Apart from these disturbances the magnetograms are quiet although there was some diurnal variation. The mean K index at Lerwick for the whole survey was 1.7 with a standard deviation of 1.2 and a maximum value of 5 (on a scale of 9; Lincoln (1967)).

In general magnetic disturbances are more intense in high latitudes than they are elsewhere. The dominant form of disturbance is the polar magnetic substorm (Rostoker, 1972). Substorms may occur as isolated events during quiet periods or in sequence during the main phase of a magnetic storm (Chapman, 1965). The duration of an isolated substorm is generally about 1 hour although a series of substorms may blend together. The most characteristic signature is a magnetic "bay disturbance" of the horizontal component occurring around midnight. The main features of this type of disturbance can be accounted for by an intense electric current flowing westward in the nightside ionosphere. Except during periods of very strong storm activity this "intensified auroral electrojet" flows approximately along the line of 70° magnetic latitude (Akasofu and Chapman, 1967) and is of limited length (10° - 90° of longitude; Rostoker, 1972) centred on the midnight meridian.

Events 1 to 4 appear to have prominent substorm components. The line of 70° magnetic latitude (coinciding with the auroral zone) runs approximately from Leirvogur to slightly north of Tromso (Matsushita and Campbell, 1967). Hence in normal circumstances, these stations lie nearly under the electrojet, which accounts for the large erratic disturbances relative to Dombas and Lerwick. In the case of Event 2, which is a simple example, Z is initially positive for Tromso and Leirvogur but negative for Dombas and Lerwick, implying that the westward electrojet was situated between these pairs of stations. During the course of the event the electrojet must have passed over Tromso, then over Leirvogur somewhat later, causing Z to become negative. This progression is explicable, in that the electrojet follows an oval which is furthest from the geomagnetic pole in the midnight sector (Chapman 1965). Events 5 and 6 do not appear

to be composed of typical substorm components, but nevertheless it is likely that at least some of the causative current system was located in the auroral zone ionosphere.

For the range observed, the total field variation is given as

$$\Delta T = \Delta Z \sin I + \Delta H \cos I$$

where I is the inclination of the total magnetic vector, which can be taken as 76° for the survey area (Figure 2.7), giving

$$\Delta T = 0.97\Delta Z + 0.24\Delta H$$

This formula allows the computation of the ΔT variations at the land stations. However it was found to be impossible to subtract the ΔT variations measured on land in a systematic way to yield smoother anomalies. Part of the problem is the proximity of the electrojet, which gives rise to large changes over small distances, as is clear from the example discussed above. In order to overcome this problem one would need to model accurately the causative current source, which is not possible using only four stations.

Another complication is that temporal variations at sea are expected to be different to those measured on land, because of direct and indirect effects of the high conductivity of sea water, which in general acts to attenuate the total field variations at sea.

The direct effect is the geomagnetic "ocean effect" (Price 1967). In an ocean of large horizontal extent, externally varying magnetic fields can induce electrical currents which produce a secondary magnetic field which opposes the vertical component and reinforces the horizontal component. Price gives the plane layer theory for this effect, which is applicable for periods of up to 1000 s, in which case the average local water depth is the principal factor determining the extent of the attenuation. Computations for each of the six events predict a typical total field attenuation of about 2 for a period of 1000 s (roughly corresponding to the highest frequency spikes on the magnetograms), assuming that the vertical and horizontal

fields are either completely in or out of phase and are of roughly equal magnitude, which is a justifiable simplification. The ocean effect is therefore expected to cause a substantial attenuation of high frequency ΔT variations. These are also filtered out directly during processing (Figure 2.5). However at the dominant frequencies of the bay disturbances (7000 to 11000 s) the ocean effect is unlikely to be significant.

The indirect effect is the "coast effect" (Price, 1967; Parkinson and Jones, 1979; Auld et al., 1979) which is likely to be important at larger periods. This effect is due to coast parallel eddy currents which flow within the ocean (and also perhaps at greater depths) in response to externally varying magnetic fields. These currents cause a considerable enhancement of ΔZ relative to ΔH near the coast. The extent of this enhancement is very difficult to estimate in the absence of detailed information but an increase of ΔZ to at least 5 times its normal value is not exceptional near the coast in other areas. As all the base stations used are coastal this effect could therefore be very important.

Examination of Figure 6.5 (magnetic chart) supports the idea that temporal variations are attenuated at sea, as very little of the available marine magnetic data from the southern Norwegian- Greenland Sea is obviously affected by temporal disturbances (although quite severe temporal disturbances are known to occur regularly in this region). In some instances SHACK977 data which was recorded during disturbed periods can be compared directly with other data recorded along closely approaching or cross-cutting older tracks. In general there is good agreement indicating that the Shackleton 9/77 data is not severely affected. As an example, it would be reasonable to suppose, from Figure 2.8, that the roughness of the anomaly measured during Event 1 was due to high frequency temporal disturbance. However the magnetic anomaly map shows that this roughness is a characteristic feature of the southern ridge complex and Line 1 exhibits good ties with other profiles which it crosses.

In view of the various uncertainties involved it was

decided not to make any corrections for short and intermediate period temporal variations. In the case of events 1 to 3 the only obvious temporal effects correspond to H bays at 0500 and 2300 hours on Day 231. However because of the unpredictability of the anomalies over the Jan Mayen block other disturbances may be present but hidden. In this thesis a quantitative interpretation is presented for the Norway Basin profile only, over which the anomalies are relatively undisturbed. The sole effect of Event 4 appears to be a small irregularity of about 100 nT at 2100 to 2200 hours, which correlates quite well with the ΔZ variation at Leirvogur, perhaps indicating a small coast effect in this case. Events 5 and 6 seem to have produced long wavelength undulations only. These can be removed, along with diurnal variation, by use of a suitable low-cut frequency domain filter during interpretation (Section 5.2).

2.5 Preliminary processing of the 1977 cruise data

After processing on the IOS IBM 1130 computer was completed (Figure 2.5) a tape was brought back to Durham containing values of time, latitude, longitude, corrected and uncorrected depth, free air gravity anomaly and total and residual magnetic field, tabulated at 2 minute intervals in Merged-Merged Format 3 (Talwani et al., 1972).

A working data file was produced at Durham in which the data from the second part of the cruise were stored in a modified Merged-Merged Format (described in Appendix A). For the purposes of interpretation the cruise was divided into a number of nearly straight profiles (Figure 2.3). The start and end times of these profiles are shown in Figures 3.1 and 3.2. In the working data file the redundant ship identification occupying the first 8 columns was deleted and profile and seismic data tape identification were edited in. Distances along each profile were then calculated in a manner which reduced the effects of irregularities in heading and were stored in columns 43 to 52, which would otherwise be empty.

The distance computations were effected using the

FORTTRAN program DIST which acts on a single profile in the following way. The profile is translated over the sphere to place the first and last points on the equator, with the first point lying on the zero meridian. The profile is then mapped onto a plane with a minimum of distortion using a cylindrical projection, such that the abscissa and ordinate of a point are proportional to its new longitude and latitude respectively. On the plane, each point is projected normally onto a best fit straight line, and distances are measured along the line. The mathematical development of the method, taking into account the Earth's ellipticity, is as follows.

Consider a profile consisting of N points

$$(\theta_i, \phi'_i) \quad i=1, 2, \dots, N$$

where θ_i is longitude and ϕ'_i is geographic latitude. Conversion to geocentric latitude ϕ_i (Jeffreys 1970, pp 67-69) is made using an average value of ϕ_i

$$\phi_i = \phi'_i - \epsilon \sin(\phi'_1 + \phi'_N)$$

where ϵ is the earth's ellipticity (ignoring terms of order $\epsilon(\phi'_N - \phi'_1)$ along with terms of order ϵ^2). The profile can now be translated as described above, with the transformed longitude u_i and latitude v_i calculated as

$$(u_i, v_i) = (\arccos(A/(1-B^2)^{0.5}), \arcsin(B))$$

where

$$A = \cos(\phi_1) \cos(\phi_N) \cos(\theta_1 - \theta_N) + \sin(\phi_1) \sin(\phi_N)$$

$$B = C_1 \cos(\phi_1) \sin(\theta_1 - \theta_N) + C_2 \sin(\phi_1) + C_3 \cos(\phi_1) \sin(\theta_1 - \theta_N)$$

$$C_1 = (\sin(\phi_1) \cos(\phi_N) / \sin(\alpha))$$

$$C_2 = (\cos(\phi_1) \cos(\phi_N) \sin(\theta_N - \theta_1)) / \sin(\alpha)$$

$$C_3 = -(\cos(\phi_1) \sin(\phi_N)) / \sin(\alpha)$$

$$\alpha = \arcsin(1 -$$

$$(\cos(\phi_1) \cos(\phi_N) \cos(\theta_N - \theta_1) + \sin(\phi_1) \sin(\phi_N))^2)^{0.5}$$

These equations are readily derived using either vector algebra or spherical geometry. The corresponding planar coordinates, x_i and y_i , are given by

$$(x_i, y_i) = r(u_i, v_i) \quad i=1, 2, \dots, N$$

where

$$r = a(1 - \epsilon \sin^2((\phi_1 + \phi_N)/2))$$

is the average radial distance from the centre of the earth to the profile, given in terms of the equatorial radius, a . The computations so far described are performed by the subroutine PROJ.

A "maximum likelihood" straight line is then fitted through the planar coordinates, such that the sum of squared perpendicular distances from the line is minimised, then the profile points are projected onto the line. The general purpose subroutine FIT, based on the theory outlined by Davis and Goldsmith (1972, pp 185-187 & 208-209), was written to compute the set of projected distances d_i .

In order to minimise rounding errors new coordinates (X_i, Y_i) are defined as

$$(X_i, Y_i) = (X_i - x_1, Y_i - y_1) \quad i=1, 2, \dots, N.$$

In the specific application of FIT described here $(x_1, y_1) = (0, 0)$ so this computation is redundant. The gradient, a is then efficiently computed as

$$a = F/2E + (E/|E|) ((F/2E)^2 + 1)^{0.5}$$

where

$$E = S_{xy} - S_x S_y / N$$

$$F = S_{yy} - S_y^2 / N - S_{xx} + S_x^2 / N$$

where, for example,

$$S_{xy} = \sum_{i=1, N} X_i Y_i$$

The distances d_i are then calculated as

$$d_i = (X_i + aY_i) / (1 + a^2)^{0.5}$$

Also computed by FIT are the intercept b and the set of projected points, but these are not used in the application to distance computation.

A listing of DIST, and instructions for use are to be found in Appendix B. The program takes as its input a section of a working data file corresponding to a single profile and generates another file with distances inserted, together with some ancillary information, which is shown in Figure 2.3 for the 1977 cruise. In the case of the Norway Basin traverse, DIST was run over the separate profiles 11, 12 and 13, but the distances were concatenated so as to be continuous over the whole traverse.

Other preliminary processing steps were as follows. Spot checks were made of the free air anomaly using the original chart records to determine the raw value of the gravitational field, the charts of Appendix A to recompute the Eotvos correction and the GRS67 gravity formula.

Agreement was to within 1 mgal. All of the values on the working data file were adjusted by -14.8 mgal as described in Section 2.3. The computation of the residual magnetic anomaly was systematically checked using the departmental program IGRF, based on the International Geomagnetic Reference Field 1975 (IAGA Division I Study Group, 1976). The difference between original and recomputed anomalies was as great as -6 nT between 1010 and 1224 hours on Day 234, but elsewhere was less than 2 nT in magnitude, so no alterations were made.

Chapter 3

SEISMIC STUDIES3.1 Seismic reflection and refraction data from the 1977 cruise

In Figures 3.1 and 3.2 line drawings of the single channel EPC monitor records are presented, together with interpretations of sonobuoy profiles 1 to 4. The identification of horizons was aided by detailed comparison with the sections presented by Gairaud et al. (1978), which are shown in Figures 1.9, 1.10 and 1.15. Important features of the profiles will now be described. The Jan Mayen profiles are treated from south to north, followed by the Norway Basin traverse.

Profile 1 (Figure 3.1)

Profile 1 crosses the southern ridge complex at 67°N; further south than any published multichannel section and would be an interesting candidate for full processing.

The acoustic basement is well defined and typically "oceanic" at distances less than 40 km and greater than 205 km along the profile. The western and eastern boundaries of continental crust may lie near these limits.

Between 40 km and 140 km the basement is indistinct and was defined as the downward limit of sedimentary reflectors rather than as a clear reflector in itself. Further east the basement becomes more distinct. The whole central zone, between the limits of oceanic basement, appears to be composed of faulted blocks which are antithetically tilted away from an axis at about 130 km along the profile. Ridge 2, as defined by the free air gravity anomaly chart (Figure 1.13) is not well developed along the line of profile 1, and lies within the median zone. Ridges 1 and 3 on either side are more prominent. The median zone corresponds to the rift

Figures 3.1 and 3.2. Line drawings of the single channel seismic reflection monitor records from the SHACK977 cruise (Part II). See Figure 2.3 for profile locations. Along the top of each section distances in kilometres (Figure 2.3) are indicated, along the base, times in hours with start and end times indicated in Julian days, hours and minutes (Figure 2.2). Crossings with multichannel Capan 1 profiles (Figures 1.8 to 1.10) are indicated in the manner CP108, with the single channel sections of Figures 1.14 and 1.15, in the manner J12. Also shown (for example #1) are interpretations of the SHACK977 sonobuoy refraction profiles (Figure 2.2). Vertical axes: two way time in seconds.

Figure 3.1 illustrates profiles which cross the Jan Mayen block. Identification of horizons follows that in Figures 1.9 and 1.10 and basement ridges (for example R1) are labelled as in Figure 1.13. The median axis of crustal extension is indicated as M.

Figure 3.2 shows the most northerly Jan Mayen profile (9) and the Norway Basin traverse. Inset in section 11 is a diagram illustrating the major morphological features of submarine slumps (redrawn from Dingle 1977), for comparison with the structures at the western end of the line.

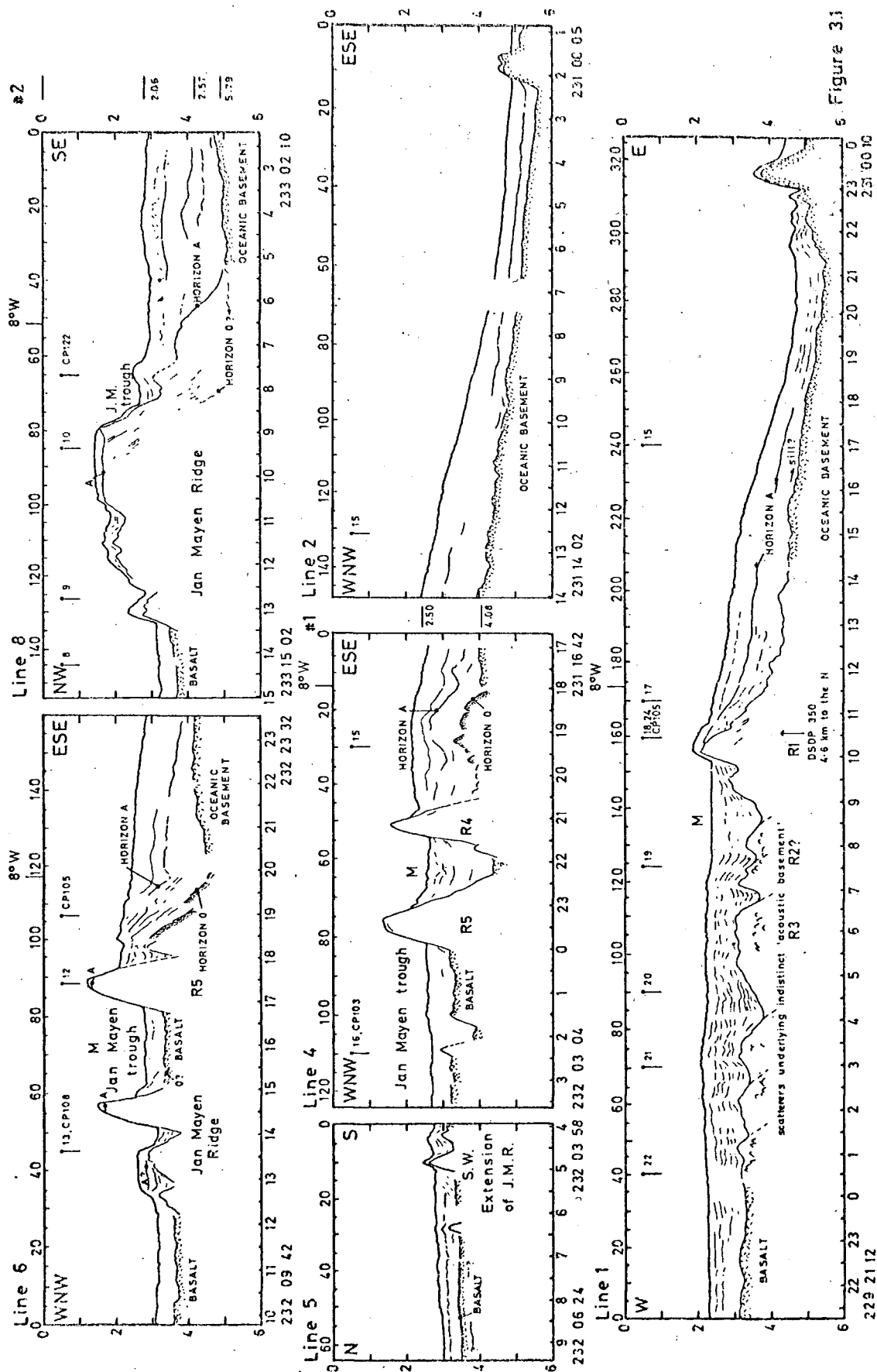


Figure 3.1

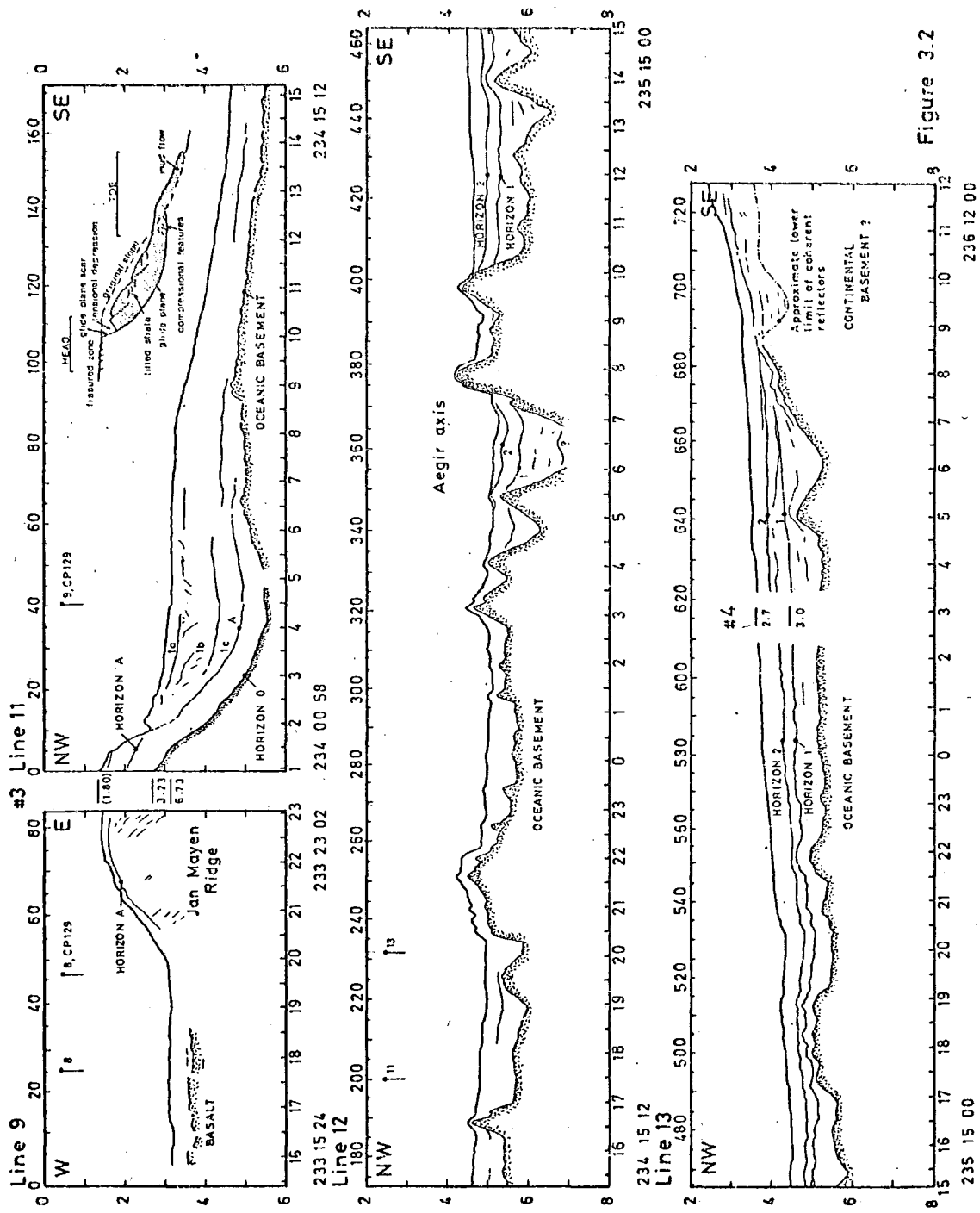


Figure 3.2

described in section 1.8, but there is no broad floor of basalts as there is further north. DSDP site 350 lies only 5 km north of the line near the crest of ridge 1, hence it is likely that basalt of late Eocene age forms the basement in this region. As on CP105, which intersects line 1 there is an indication of another reflector about 0.3 s below, which may correspond to Horizon 0. Further west the nature of the basement is uncertain but it may well be basaltic also. There is some indication of reflectors below the basement.

As a general point it is worth commenting on the variation of the number of sedimentary reflectors which are depicted on this and other sections. If there is a strong specular sea-bottom reflection (for example between 130 and 145 km) the automatic gain control on playback diminishes the amplitude of subsequent reflections, so fewer are picked. If, in contrast the sea bottom is rugged (for example between 45 and 130 km along profile 1) and gives rise to a diffuse, weak reflection, a large amount of detail is visible below, although reflector coherence may be destroyed.

Profiles 5, 4 and 2 (Figure 3.1)

Profiles 5, 4 and 2 make up a composite line crossing the Jan Mayen block just south of the Jan Mayen Ridge. Conservatively estimated, continental crust may extend only as far westward as 30 km on line 5, just beyond the last peak to protrude through the opaque layer. Alternatively the opaque layer further west may represent flood basalt overlying deeper continental crust.

The Jan Mayen trough, at this latitude, does not appear as a pronounced graben, and ridge 5 tilts away from the deeper rift which is developed to the east, between ridges 5 and 4. The internal structure of these ridges is not distinguishable on the monitor records, mainly because of the large vertical exaggeration and steep dips. Considering profiles CP108 and 103, which lie to the north and south of line 6, it is possible that sediment at least 1 s thick overlies Horizon 0 underneath the ridges. Comparison with CP105 suggests that basalt may overlie Horizon 0 east of

ridge 4; alternatively the hummocky appearance may be due to closely spaced faulting.

An abrupt change in basement reflector characteristic at 15 km on line 4 suggests the boundary of oceanic crust might lie as far west as this. Sonobuoy 1 gives a refraction velocity of 4.08 km s^{-1} for the basement in this region, a value typical of oceanic layer 2 in the southern Norwegian-Greenland Sea, but equally characteristic of Horizon 0 beneath the Jan Mayen Ridge

Profile 6 (Figure 3.1)

Profile 6 crosses the southern end of the Jan Mayen Ridge and the northern end of ridge 5. The Jan Mayen trough appears to be a well developed graben at this latitude. As a conservative estimate, continental crust extends from 30 to 120 km. The western end of the profile resembles CP108, which it crosses. Horizon 0 is not clearly identifiable under the Jan Mayen Ridge on either the monitor or CEPAN record. In the case of the smaller ridges to the west the deepest reflector shown possibly corresponds to basement. East of the Jan Mayen Ridge, Horizon 0 may disappear beneath the basalt flooring the trough, as interpreted on CP108. A sediment thickness of about 1.5 s is expected beneath ridge 5.

Profile 8 (Figure 3.1)

Profile 8 crosses the Jan Mayen Ridge at about 69°N , near the latitude of the original profile of Johnson and Heezen (1967). Although profile 8 does not extend very far into the Jan Mayen Basin it is likely that no further "windows" in the opaque layer occur to the west of the line. This is because the bathymetry and gravity anomaly trends corresponding to the most westerly occurrences of continental crust in lines CP110 (to the north) and CP112 (to the south) coincide with the most westerly knoll on profile 8 if they are projected in southerly and north-westerly directions respectively.

The deep structure of the Jan Mayen Ridge is not clear on the monitor record. The deepest reflectors underlying the eastern flank probably represent Horizon 0 as they are of comparable depth to this horizon on CPL29. Further west, Horizon 0 is probably quite shallow, but bubble pulse reverberations make the monitor records difficult to pick. To the east of the ridge there is a weak bathymetric extension of the Jan Mayen trough, but how this is reflected in the underlying structure is unclear. Continental crust exists at least as far east as 35 km.

Sonobuoy profile 2 was interpreted using both refraction and wide-angle reflection techniques (Gerdes 1978). The sediment structure and basement depth must change along the track, as can be seen by comparing the structure at the end of section 6 with that at the beginning of section 8. The sonobuoy interpretation seems to match the latter structure better which is surprising as only 3.4 km of data was used, from the beginning of the run (that is, near the end of line 6) and small dips were obtained from the wide-angle reflection analysis. If the interpretation is to be correlated with line 6, the 2.57 km s^{-1} layer must correspond to the top of oceanic layer 2, with the higher velocity representing an interface within the basement. A playout of the reflection data along the profile is required for an adequate interpretation of this sonobuoy.

Profile 9 and the Norway Basin traverse (Figure 3.2)

Profiles 9, 11, 12 and 13 together form a traverse of the Jan Mayen Ridge and rise, the Norway Basin and the Norwegian rise. Some of section 9 was lost (due to airgun failure) and the available records are difficult to pick, but they would be expected to show structures similar to those of CPL29. In the case of sonobuoy 3, there is again a problem. Correlation with the beginning of line 11 gives a velocity of 3.23 km s^{-1} for Horizon 0, which is rather low when compared to previous interpretations (Figure 1.12), which would perhaps suggest that this value should be correlated with a layer within sequence 2. In this case an anomalously high (6.73 km s^{-1}) velocity would result for Horizon 0. However

the interpretation of sonobuoy 3 was based on faded Geospace records, and is not expected to be as reliable as those of sonobuoys 1 and 2.

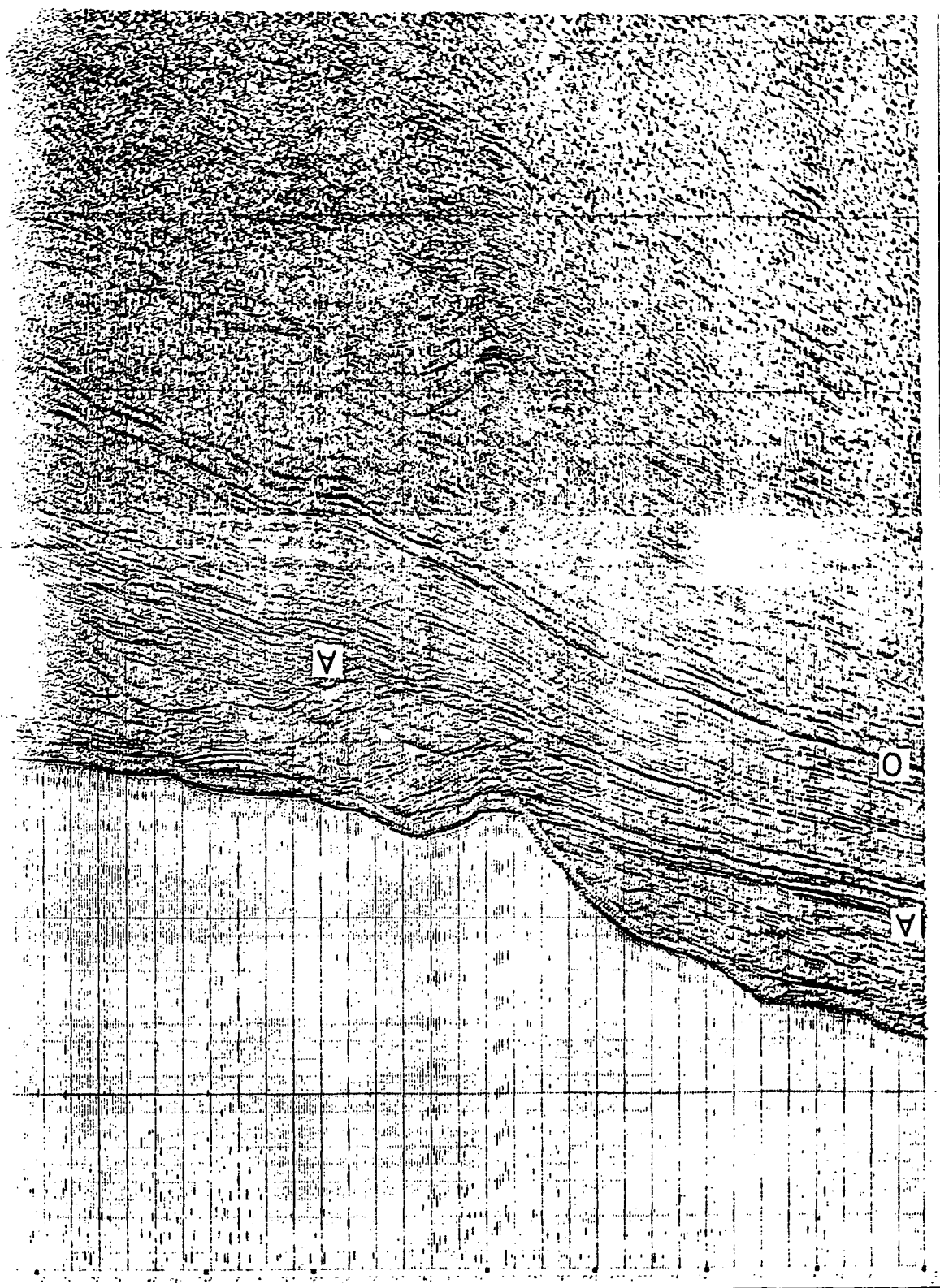
Part of the beginning of profile 11 has been processed using multichannel techniques (Section 3.3). The deeper structures shown for profile 11 (between 0 and 24 km) were derived from the processed trial section, which is shown in Figure 3.3. This section allowed an interpretation to be made of what was otherwise a confusing profile. At the beginning of the processed section, Horizon A is at about 2 s depth, overlying Horizon 0 at nearly 3 s. Horizon A can be traced to about shotpoint 140 where there is a bathymetric notch, but then loses continuity. It is considered likely that Horizon A is downfaulted to the east at this point. Other features revealed in the line drawing of Figure 3.2 also suggest that a major submarine slump is present on the Jan Mayen rise, as comparison with the inset diagram (Dingle, 1977) demonstrates. Inspection of the precision depth records shows a disrupted sediment surface with diffraction hyperbolae (similar to the echo type IIIc of Damuth (1978)) at the top of the rise just prior to the beginning of line 11, possibly corresponding to the characteristic fissured zone. There is bathymetric expression of subsidiary glide plane scars up to 10 - 15 km east of the main scar and further east there is a region of hummocky topography between 55 and 105 km along the line, possibly corresponding to the toe of the slump. As discussed in Section 1.6, with reference to CP129, the uppermost sedimentary unit, 1a, is parallel-bedded and probably post-dates slump activity. Therefore the bathymetric evidence for these fine scale features is muted.

The main glide plane appears to follow the base of unit 1, which on line 11 and in sections CP129 and 110, exhibits a chaotic reflector pattern. The original depth of the glide plane must have approached 1 km in places and a minimum length would appear to be about 50 km. Only two submarine slumps of greater cross-sectional area are reported by Dingle (1977); namely the Aghulas slump on the sheared continental margin of southeast Africa, and the Bassein slump associated with the Sunda Arc subduction zone.

Figure 3.3. Processed 11 fold seismic reflection section from the beginning of profile 11. Kindly supplied by J.H.Peacock and M.J.Poulter, who performed the processing. Each trace (340 in all) represents 21 seconds, or 70.7 m along track. The section starts at the begining of profile 11 and is about 24 km long. Processing parameters were as follows (J.H.Peacock personal communication):

Demultiplex	SEG A
Amplitude recovery	$t \exp(0.2t)$
Bandpass filter	4, 8, 50, 60 Hz
Deconvolution	160 msec +10% prewhitening
NMO correction	
CDP stack	
Velocity analysis	every 40 traces
Deconvolution	140 msec +3% prewhitening
Dynamic equalisation	gatewidth 0.256 sec.

Figure 3.3



The total length of the Jan Mayen rise affected by the slump is uncertain. A possible tensional depression is seen at the eastern end of CP114, (Figure 1.9) 150 km to the north, and there is evidence of chaotic facies on CP110 and 129. Conceivably the extension of the Jan Mayen trough at the end of line 8 also represents a large tensional depression caused by slumping. In all events the feature is clearly of considerable dimensions and merits further study.

Interpretation of the Norway Basin traverse is relatively straightforward. The western boundary of typically oceanic basement lies near 45 km on profile 11. Its position is consistent with the boundary as observed on CP129. The oceanic basement is remarkably smooth for a considerable distance along the traverse. Magnetic anomaly evidence (Section 6.4) suggests that this profile may cross two fracture zones at an acute angle; the first at about 70 km and the second at about 220 km along track. There is no real sign of a fracture zone at 70 km but the peaks and troughs at about 220 km along the profile may be related to the second proposed fracture zone, for which the magnetic anomaly evidence is more specific. Further along the profile the basement topography remains rugged, with the degree of relief culminating in the vicinity of the Aegir axis at 360 km along profile 12. The central zone of rough topography roughly corresponds to the area formed during the period of fan-shaped spreading in the Basin. The gravity interpretation presented in Section 4.5 suggests that the deepest reflector visible beneath the Aegir axis corresponds to oceanic basement, although this is not certain from the seismic records.

Northwest of the Aegir axis there are no prominent reflectors visible within the sedimentary sequence, although there is a fairly consistent weak parallel layering which generally drapes the topography. To the southeast of the axis, two prominent reflectors, designated Horizons 1 and 2 are apparent. Overall there is a greater sedimentary thickness on the eastern side of the basin, indicating a greater sediment supply from the Voring Plateau and Norwegian shelf region than from the Jan Mayen Ridge. The mode of

sediment occurrence, deep in troughs and shallow ridges suggests lateral transportation rather than pelagic sedimentation (particularly in the eastern part of the basin). Much of the sediment within the basin must postdate the cessation of spreading. Horizons 1 and 2, in particular, overlie the Aegir axis. Sonobuoy 4 indicates a change in velocity corresponding to Horizon 1. The interpretation is based on Geospace monitor records and should be treated with caution.

The "transparent layer" described by Jankowski (1977) from the Shackleton 4/76 profiles across the Norwegian margin would be expected to lie below Horizon 1, but is not an obvious feature. Although relatively few coherent reflections are observed below Horizon 1 this may be a consequence of the high reflectivities of Horizons 1 and 2.

The eastern limit of oceanic crust appears to be at about 685 km along line 13, as has been previously noted (J. Robinson, personal communication).

3.2 Seismic compressional wave velocities in the sediments of the southern Norwegian-Greenland Sea

Accurate conversion from vertical reflection time to depth is necessary for the purposes of the gravity anomaly interpretation (Chapter 4) and to a lesser degree magnetic anomaly interpretation (Chapter 5).

Ideally, multichannel reflection data carry sufficient information for the determination of the vertical velocity structure at any point along a profile. The velocity analysis procedure of Taner and Koehler (1969) yields stacking velocity as a function of reflection time. If the reflectors are horizontal, or nearly so, the stacking velocity closely approximates to the rms velocity of Dix (1955), which can be inverted to give the variation of velocity with depth. The depths calculated from high quality analyses are accurate to within 1 or 2 percent (Woeber and Penhollow, 1975).

Not very much velocity information from this source is available. As discussed in Section 3.4 the velocities determined from the velocity analyses which were performed for the start of profile 11 are probably consistently low. A certain amount of velocity information derived from the CEPAN-1 survey has been published (Gairaud et al. 1978) and is of use.

Leg 38 of the Deep Sea Drilling Project yielded velocity information in two ways. Firstly, sonic velocity measurements were made on the drillcores. The time average velocity from the sea surface down to a specific level can be calculated as $\sum d_i / \sum (d_i / v_i)$ where d_i is the thickness of the interval characterised by a velocity v_i . Secondly, where a reflector can be identified both in the drill cores and in the single channel seismic profile crossing the site, the average velocity from the sea surface down to the reflector can be calculated directly. Table 3.1 lists the average velocities down to basaltic basement or Horizon A, which were derived by these methods, for sites in and around the Norway Basin.

Seismic refraction data have been listed by Gronlie and Talwani (1978) and Sundvor et al. (1979). The fundamental quantities measured in a refraction survey are velocity and intercept time but the results are generally presented in terms of velocity and depth. The conversion from intercept time to depth involves the often unwarranted assumption that the material between two refractors is isotropic and homogeneous and has the same velocity as the upper refractor. For the purposes of application to reflection data it is more satisfactory to present refraction velocities as a function of two way time because the conversion from measured intercept time to two way time involves assumptions about velocities in a weak way only (via a relatively insensitive cosine term). Thus two way times can be obtained fairly accurately even if they are computed from published velocities and depths which may themselves be in error, if any of the assumptions used in deriving them are incorrect.

In addition to these sources of information comparison

Table 3.1

Site No	Location of site	Nature of reflector	Depth of reflector (km)	Two-way reflection time (s)	Average velocity (km/s)	Average velocity from sonic logs (km/s)
336	Iceland-Faeroe Ridge	Basalt	0.485	0.59	1.64	1.66
337	Aegir axis	Basalt	0.113	0.151?	1.50?	1.62
338	Outer Voring Plateau	Basalt	0.401	0.50	1.60	1.73
343	Outer Voring Plateau	Basalt	0.252	0.30	1.68	1.64
345	Lofoten Basin	Basalt	0.762	0.86?	1.79?	1.96
346	Jan Mayen Ridge	Horizon A	0.121	0.16	1.51	1.56
348	Iceland Plateau	Basalt	0.527	0.68	1.56	1.71
350	Southern Ridge Complex	Basalt	0.362	0.48?	1.51	1.73

can be made with general studies concerning the variation of sedimentary velocities.

Assumed velocity function

A velocity function which will prove useful in the description of the velocity data is

$$v = v_0 + kT \quad 3.1a$$

where v_0 is the surface velocity and T the two way time. A linear increase of velocity with time corresponds to a realistic decrease of velocity gradient with depth. The following relationships can be readily derived.

$$\bar{v} = v_0 + kT/2 \quad 3.1b$$

where $\bar{v}$ is the average velocity,

$$z = v_0 T/2 + kT^2/4 \quad 3.1c$$

$$v = (v_0^2 + 4kz)^{0.5} \quad 3.1d$$

$$dv/dz = 2k/v \quad 3.1e$$

where z is the depth below the sea bed.

The ocean basins

Unfortunately there is very little specific information in the seismic velocities of sediments in the Norway Basin. No velocity analyses based on multichannel seismic data appear to have been published and from this point of view analysis of the Durham data over a region of thick sediment, such as the Aegir axis would be interesting. Refraction data (Gronlie and Talwani, 1978) indicate velocities of about 2 km s^{-1} in the sediment layer but are too sparse to allow any inference about vertical velocity gradients. Measured sonic velocities of near surface sediments from nearly all DSDP Leg 38 holes are in the range $1.5 - 1.7 \text{ km s}^{-1}$ (Talwani, Udinstev et al., (1976), site reports). Velocities inferred from two way travel times and depths are generally somewhat smaller than the measured sonic velocities. This is possibly due to seismic anisotropy as the sonic velocities of unconsolidated sediments are generally measured through the core liner, parallel to the bedding. The only hole within the Norway Basin, DSDP 337, is too shallow to provide much useful information. In addition the hole is situated on a ridge and the underlying sediment

is probably more pelagic than in the deeper parts of the basin where one would expect to find higher velocity gradients. For the purposes of obtaining basement depths for gravity anomaly interpretations it is more important to obtain a realistic velocity function in areas of deep sediment than in areas of shallow sediment.

The two deepest holes which are situated in ocean basin environments similar to the Norway Basin are DSDP 345 and 348. Figure 3.4 illustrates the variation of sonic velocity with depth in these holes. In addition to a reasonably uniform increase of velocity with depth in both holes, there are sporadic high values due to lithified clay and mud layers in Hole 345 and limestone layers in hole 348. A similar origin is possible for Horizons 1 and 2 in the Norway Basin. The data are adequately represented by the following linear relationship between velocity (km s^{-1}) and depth (km)

$$v = 1.5 + 0.8z$$

which has been chosen to err on the low side, if at all, to account for the possible effects of seismic anisotropy, and also in recognition of the fact that isolated high values do not contribute greatly to the average velocity. The function of the form 3.1a, which has the same initial gradient is

$$v = 1.5 + 0.6T$$

where T is two way time in seconds. The depth time relationship derived from this equation is

$$z = 0.75T + 0.15T^2$$

The depths to basement in holes 345 and 348 are predicted to within 1 and 10% respectively using this relationship. An average velocity of 2 km s^{-1} , a value commonly used (Worzel 1974) is attained at a two way time of 1.25 s. An average velocity of 2 km s^{-1} was used for converting reflection time to depth prior to some of the magnetic anomaly interpretation of the Norway Basin traverse (Chapter 5). As the mean sediment thickness along the profile is of the order of 1 s this was sufficiently accurate.

The Jan Mayen Ridge

On the basis of the CEPAN-1 multichannel data collected over the Jan Mayen Ridge, Gairaud et al. (1978) stated that

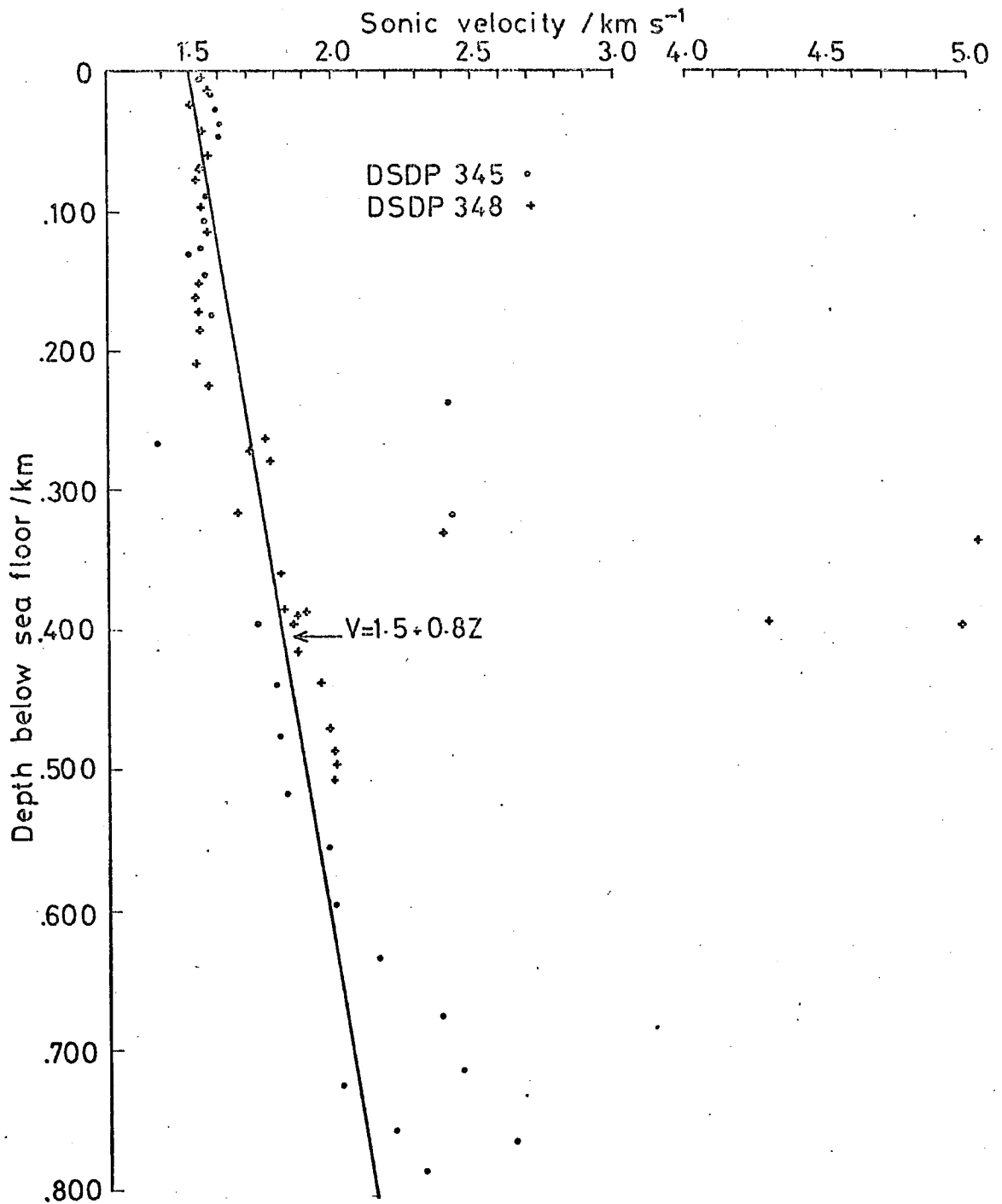


Figure 3.4 The variation of sonic velocity (measured through cores, parallel to bedding) with depth below the sea floor in DSDP holes 345 and 348. See Figure 1.2 for the location of the sites. Redrawn from Talwani, Udinstev et al. (1976, pages 459 and 602).

interval velocities range from 1.7 to 2.0 km s⁻¹ and from 2.2 to 3.3 km s⁻¹ for sequence 1 and sequence 2 respectively. However the velocities marked directly on the sections show a greater range. Where the sediment sequence is about 2 s thick, on the eastern flank of the ridge, the velocity at the base of sequence 2 appears to be as high as 4.0 km s⁻¹.

The DSDP results show that the average velocity through sequence 1 on the crest of the ridge is about 1.5 - 1.6 km s⁻¹ (Table 3.1). In DSDP holes 346 and 349 the average sonic velocity of the upper parts of sequence 2 was found to be 2.03 km s⁻¹ while in hole 347 a higher average velocity of 2.80 km s⁻¹ was measured.

In Figure 3.5 the refractor velocities for the profiles of Figures 1.11 and 1.12 are plotted against two-way times from the sea bed, which were computed from the tabulated velocities and depths given by Sundvor et al. (1979). The data from different sonobuoy stations have been distinguished by number and by joining the points by straight lines. Points corresponding to refractors within the acoustic basement (as indicated in Figure 1.12) are underlined. For nearly all the stations the variation of velocity with time for refractors within the sedimentary pile is approximately linear. The velocities of refractors within the acoustic basement generally lie well above those which would be predicted by extrapolation of the sedimentary velocities. Sonobuoys 1, 3 and 4 from the Jan Mayen bank are exceptional in this respect, and it is probable that the sedimentary sequence extends below the acoustic basement, which may represent a basalt flow as suggested in Chapter 1.

There is a wide range of velocity-time gradients within the sedimentary sequence. The highest gradients are understandably exhibited by those stations (shown dashed) overlying the western part of the ridge where Horizon A is a marked unconformity. Careful inspection of Figures 1.11 and 3.5 show that there is a progressive decrease of the velocity-time gradient as one moves eastward towards the Norway Basin. Taking advantage of the fact that the Jan

Figure 3.5 Variation of refractor velocity with two way time over the Jan Mayen Ridge. See Figure 1.11 for the locations of the sonobuoy profiles. Two way times were computed from the depths and velocities listed by Sundvor et al. (1979), which are displayed in Figure 1.12. Points corresponding to refractions within the acoustic basement (as shown in Figure 1.12) are underlined. The curves for sonobuoys overlying the western part of the ridge are dashed.

Also shown are two velocity time curves determined along a profile in the Hammerfest Basin (Tryti 1979), illustrating how the velocity determined by interval velocity analysis v_i , is consistently smaller than v_r , the velocity determined by travel time inversion of refraction data.

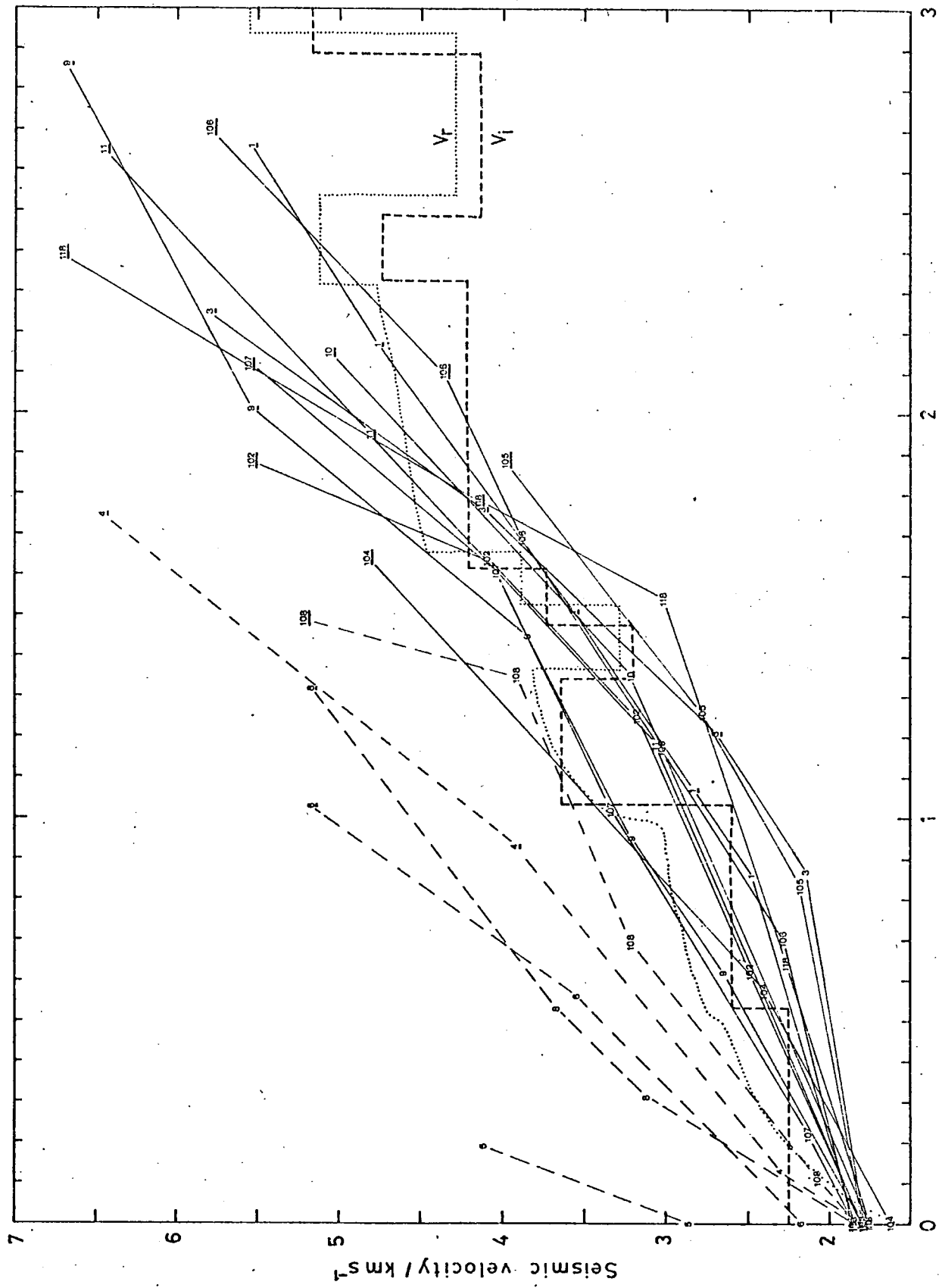


Figure 3.5

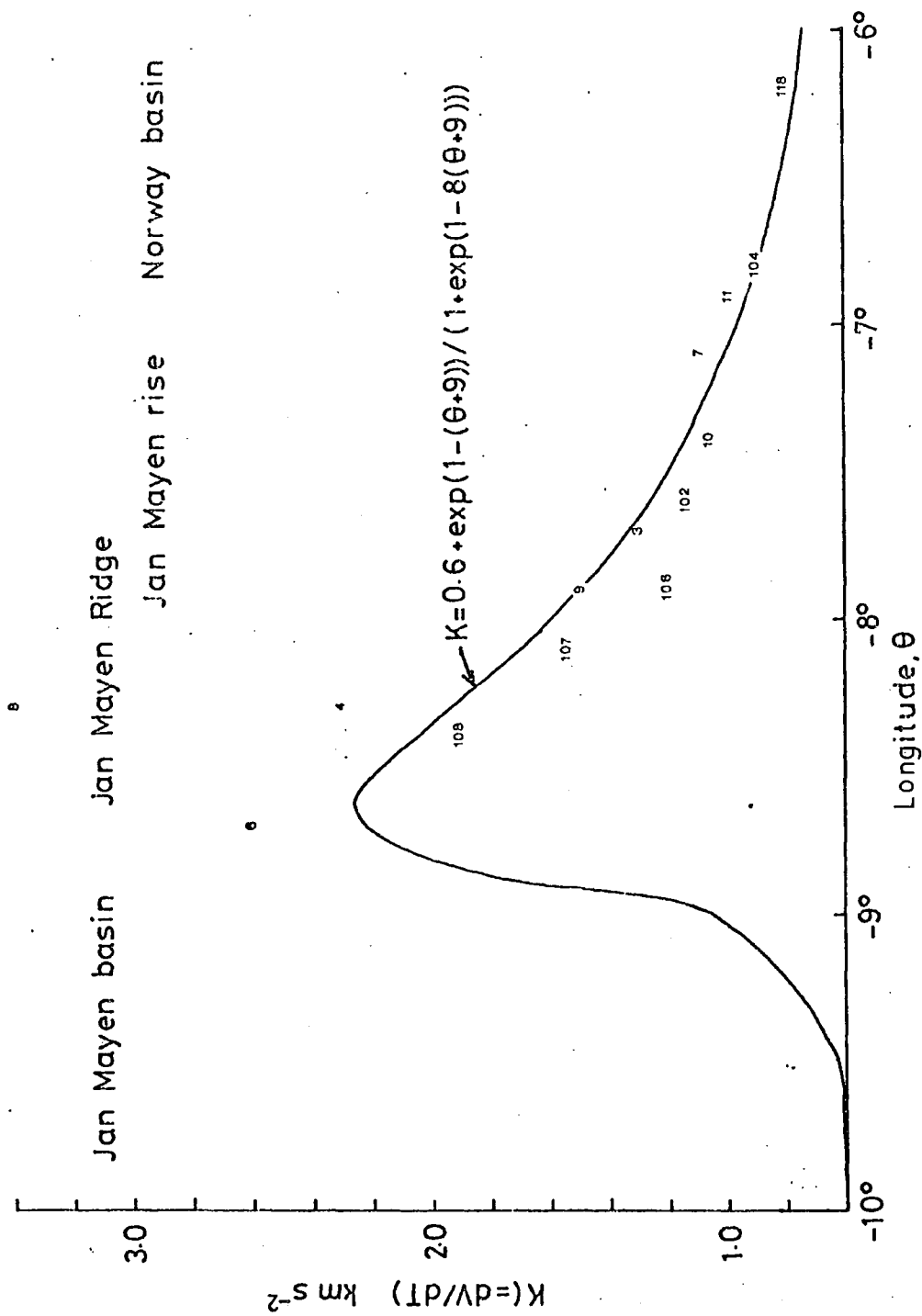


Figure 3.6 Average velocity-time gradients as a function of longitude, for the Jan Mayen Ridge (which is aligned N-S). Determined graphically from Figure 3.5. Also shown is an analytical curves which approximates the data and falls away over the Jan Mayen basin.

Mayen Ridge is aligned nearly due north-south, values of k , the velocity-time gradient, estimated from Figure 3.5 are plotted versus latitude in Figure 3.6. The values fall away exponentially eastward and appear to approach a value of the order of 0.6 km s^{-2} , which was established for the ocean basin sediments.

The very high values are mainly due to the effect of the Oligocene unconformity but further east the decrease in k is thought to arise from a progressive change from coarse, shallow water sediment to fine deep water sediment. The solid line in Figure 3.6 represents an analytic function which adequately fits the data and falls away to 0.6 km s^{-2} in both the Norway Basin and the Jan Mayen basin. Although no measurements are available for the latter basin it is expected that the sediments are fairly similar to those of the central Iceland Plateau.

Also shown in Figure 3.5 are two velocity profiles from the Hammerfest Basin, offshore northern Norway (Tryti, 1979). The sediments within the Hammerfest Basin are probably not dissimilar from those of the Jan Mayen Ridge, although most of the sequence is believed to be Cretaceous or older. The main purpose in presenting these profiles is to show how refraction velocities (v_r) are consistently higher than the interval velocities (v_i) which were determined along the same line. The average ratio (v_r/v_i) for the whole section is 0.9. This difference is probably due to the effects of seismic anisotropy (Sheriff, 1973). In Figure 3.7a sonic velocities measured parallel to the bedding, $v_{||}$, are plotted versus sonic velocities measured perpendicular to the bedding, $v_{\perp}$, for samples from DSDP hole 398, off the Iberian margin (Sibuet, Ryan et al., 1979). Although there is considerable scatter the line $v_{\perp} = 0.9v_{||}$ fits the data reasonably well.

It is likely that the refraction velocities measured over the Jan Mayen Ridge are also higher than the vertical interval velocities. The values of gradient, k are not determined accurately enough to make a 10% correction meaningful. However, if a more realistic value of 1.5 km s^{-1}

for the velocity of the surface sediments is substituted for the average value of 1.8 km s^{-1} as determined by refraction studies then a correction of the right order is effected at reasonable depths.

Comparison with data from other areas

Figure 3.7b shows the refraction data of Nafe and Drake (1957), principally from the coastal shelf of eastern North America and the adjacent ocean basins. Only deep water (greater than 1500 fathoms) and shallow water (less than 100 fathoms) stations are shown. The velocity gradient with depth is distinctly greater for shallow water sediments. This was attributed to the lower porosity of the shallow water sediments, which is probably an indirect effect of larger grain size and smaller clay content. Data from intermediate depth stations scatter between these extremes.

Superimposed on Figure 3.7b are velocity-depth curves derived, for differing values of k , from Equation 3.1d (with $v_0 = 1.8 \text{ km s}^{-1}$ as these are refraction data). Curves corresponding to values of k from 1.4 to 2.2 km s^{-2} (typical of the Jan Mayen Ridge, Figure 3.6) span most of the shallow water data, while the curve for $k = 0.6 \text{ km s}^{-2}$ lies just above most of the deep water data. This is reasonable because the deep water sediments of the relatively enclosed southern Norwegian-Greenland Sea are probably more terrigenous than those of the main Atlantic basins. The decrease in velocity gradient with depth shown by the curves is consistent with the trend of the data, providing justification for the use of a linear velocity-time function.

In Figure 3.7c velocity depth curves (for $v_0 = 1.5 \text{ km s}^{-1}$) are shown together with the interval velocity data of Faust (1951) which were derived from well surveys of shale and sandstone sections. Direct comparison is not possible because Faust's data are grouped by age, but the lower gradient curves ($k = 1.0$ to 1.4 km s^{-2}) cut Faust's curves in a manner consistent with the age-depth relationships of The Jan Mayen Rise.

Figure 3.7a shows the variation of seismic velocity measured parallel to bedding ($v_{\parallel}$) with seismic velocity measured perpendicular to bedding ($v_{\perp}$) for samples from DSDP hole 398, off the Iberian margin. Redrawn from Sibuet, Ryan et al. (1975, page 65). The high velocity sediments ($>3 \text{ km s}^{-1}$) are limestones, the others are mainly clays and shales.

Figure 3.7b shows velocities versus depth for deep water (large dots) and shallow water (small dots) seismic refraction profiles. Redrawn from Nafe and Drake (1957, Figure 3). Also shown are curves showing the variation of velocity with depth assuming a linear increase of velocity with time (gradient k) with a surface velocity of 1.8 km s^{-1} .

Figure 3.7c shows Faust's (1951) relationships between interval velocity and depth, for sand-shale sequences of various ages. Also shown are curves as in 3.7b, with a surface velocity of 1.5 km s^{-1} .

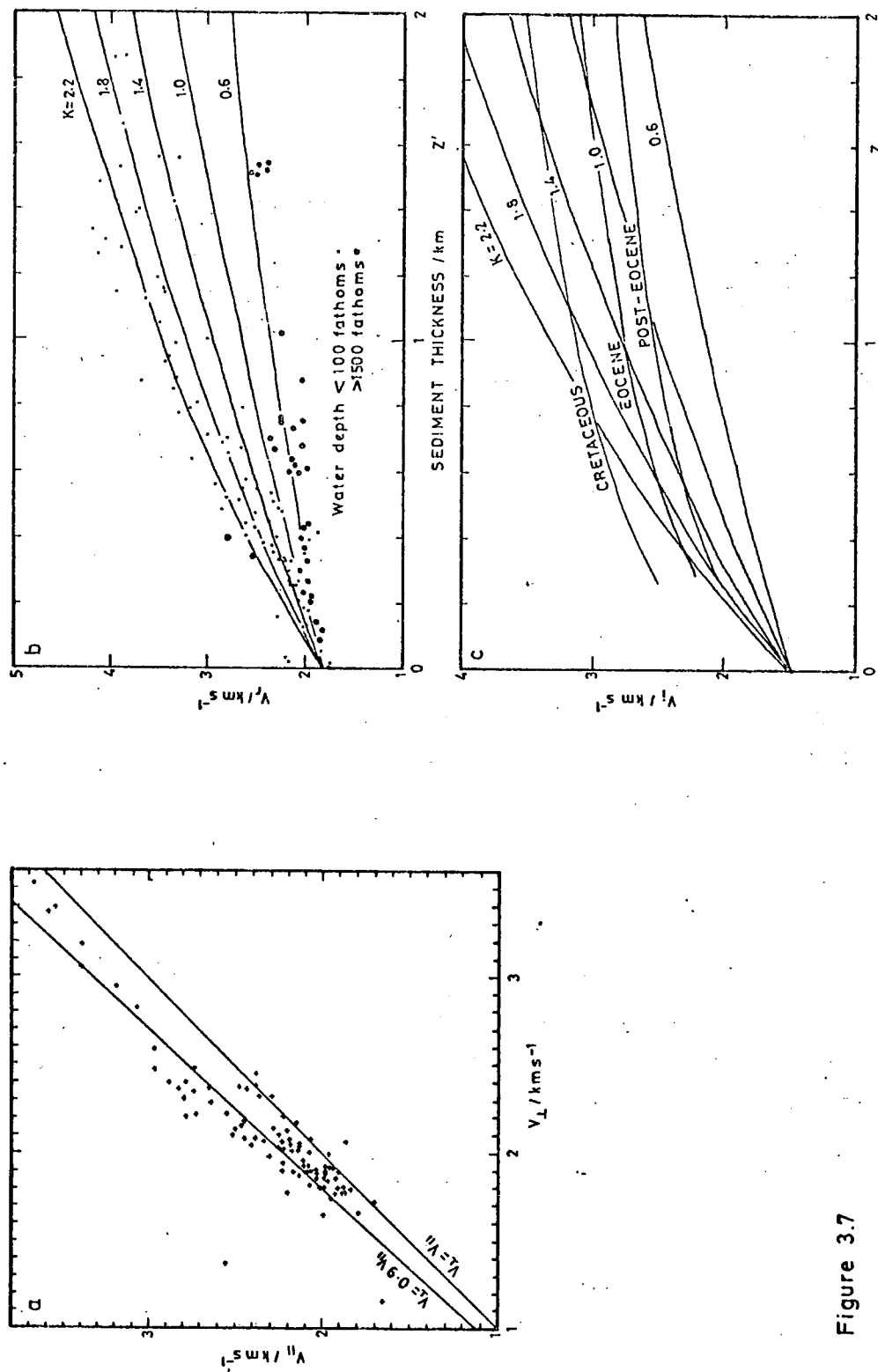


Figure 3.7

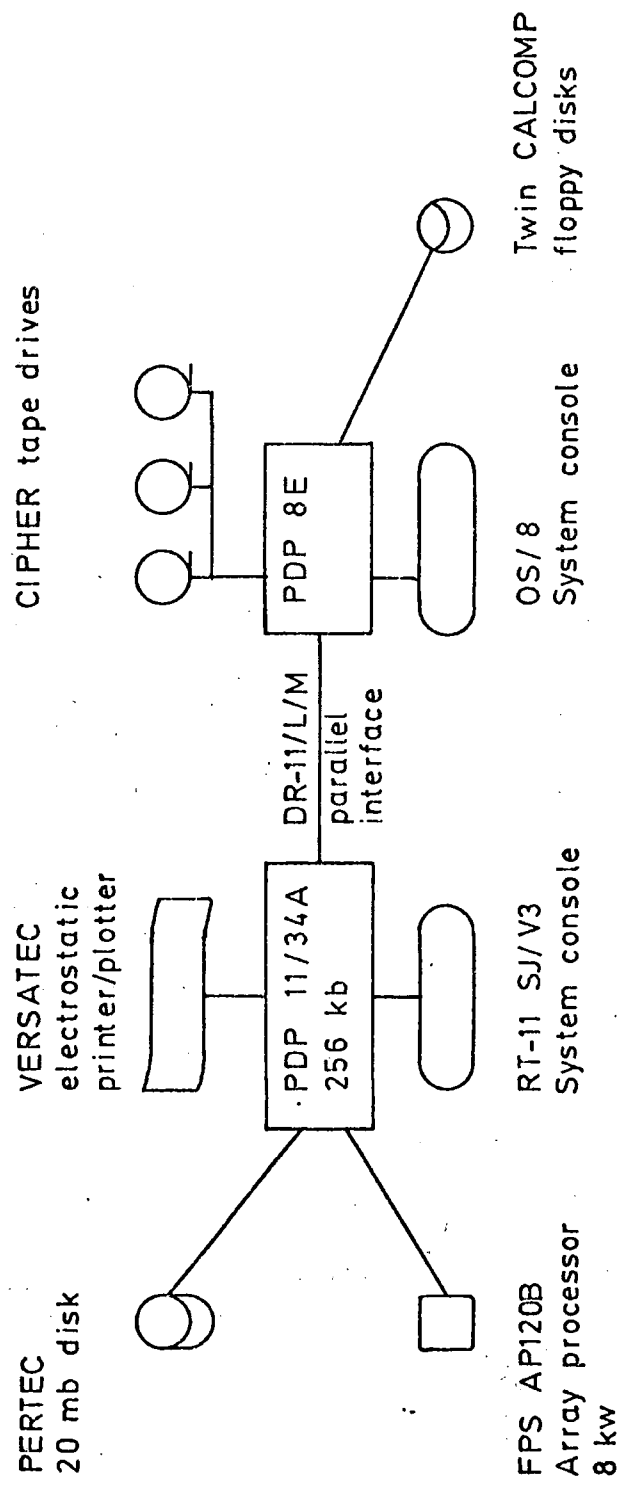


Figure 3.8 Block diagram of the Durham University Reflection Seismic Processing System (After a diagram by M.J.Poulter).

Summary

Available data show that the linear relationship

$$v = 1.5 + kT \quad 3.2$$

between velocity (in km s^{-1}) and two-way time (in s) adequately represents the interval velocities within the sediments of the survey area, where the gradient k , varies from about 0.6 km s^{-2} in the deepest environments to about 2.4 km s^{-2} over the crest of the Jan Mayen Ridge. The relationship

$$k = 0.6 + \exp(1-(\theta+9))/(1+\exp(1-8(\theta+9))) \quad 3.3$$

between longitude θ (degrees) and gradient k (km s^{-2}) was derived for the Jan Mayen Ridge sensu stricto, and can be applied to profiles 8, 9 and the Norway Basin traverse. In the absence of any other information a value of $k = 1.0 \text{ km s}^{-2}$ is assumed for profiles crossing the southern ridge complex.

Corresponding to Equation 3.2 the following equations relate depth and two way time and velocity and depth.

$$z = 0.75T + kT^2/4 \quad 3.4$$

$$v = (2.25 + 4kz)^{0.5} \quad 3.5$$

3.3 The Durham University Reflection Seismic Processing System

In this section a brief description is given of the Durham University Reflection Seismic Processing System as it presently stands. A fuller description will be given by M.J. Poulter (thesis in preparation). Figure 3.8 shows the basic hardware configuration. Work on the system was begun in October 1978 and has included interfacing of the various modules (including modification of the software systems provided) and the development of a basic set of seismic processing programs. Implementation of a batch stream system which will allow automatic "production line" processing of reflection seismic data is underway. The majority of the work has been undertaken by J.H. Peacock and M.J. Poulter, with the present author spending several months on the writing and initial implementation of interpolation, velocity

analysis and CDP stacking programs. This work is described in more detail in Section 3.4.

The system is built around a PDP11/34A 16 bit minicomputer which utilises the RT-11 operating system and is programmed in FORTRAN and Assembler. The PDP-11 has 256 kilobytes of storage in all (1 kilobyte = (kb) = 1024 characters), of which 56 kb are available in lower storage for program and data storage and 192 kb are available in virtual (extended) memory. Most of the processing programs in use occupy the majority of lower memory, so any large data arrays must be stored in virtual memory, where the access is slower, due to the indirect addressing system which is used. A total of 48 k REAL*4 values can be stored in virtual memory, although any single array is limited to 32 k.

An auxiliary PDP8E 12 bit minicomputer handles data transfer to and from tape. The capacity exists for reading and writing 9 track 800 BPI tapes in either blocked or unblocked formats. Thus the original field tapes (unblocked SEG format) can be read and in addition partially or fully processed data can be written back onto tape in blocked or unblocked format as required.

The Pertec 20 Megabyte disk pack provides storage which is intermediate in capacity and access time between the tape decks and the PDP11 virtual memory. About 5 Mbytes are used for storage of system and geophysical programs and the remainder can, in principle, be used for the storage of data although in practice the full capacity is not used.

For the purposes of rapid vector computations a floating point array processor, whose capacity is 8 k 38 bit data words, is linked to the PDP11. Control is generally via FORTRAN programs resident in the host PDP11 lower memory.

Transfer of data between the PDP11 memory and the array processor and disk pack is carried out efficiently using direct memory access (DMA) at both ends. DMA transfers and all processing by the AP can be overlapped with the operation of the host processor once the initial instructions have been

issued. The software for DMA transfers to and from virtual memory is not commercially available and was developed within the department by M.J. Poulter.

Plots and seismic sections are produced on a Versatec electrostatic printer plotter.

3.4 Velocity analysis and CDP stacking

The purpose of this section is to provide brief documentation of the velocity analysis, CDP stacking and other ancillary programs which were written for the processing system. These programs are listed in Appendix B.

The complementary techniques of CDP stacking (Mayne, 1962) and velocity analysis using the semblance function (Taner and Koehler, 1969) require as input a set of reflection seismic traces, acquired with different shot-receiver spacings such that the midpoint between shot and receiver is the same for all traces. In this case primary reflection events which are due to flat or undulating horizons can be shown (for example Taner and Koehler) to follow trajectories of two-way reflection time, T versus shot-receiver offset X , which are given by

$$T^2(X) = T^2(0) + X^2/V^2$$

where $T(0)$ is the time at which the event would be observed for coincident shot and receiver and V is the "stacking velocity" for the event. In the case of a horizontal reflector overlain by a number of beds of differing velocities, the stacking velocity closely approximates the rms velocity of Dix (1955).

A knowledge of the variation of stacking velocity with zero offset reflection time allows the effect of "normal moveout" to be removed from a set of traces, which can be then stacked together (that is, added) to give a single trace in which multiple reflectors are attenuated and primary reflections are enhanced relative to the noise level. The purpose of velocity analysis is to provide the velocity information necessary for CDP stacking.

Selection of traces for velocity analysis and stacking

Common mid-point acquisition at sea is conventionally achieved by steaming at a rate of $0.5\Delta X/\Delta S$ where X is the spacing of receivers along the array and ΔS is the shot interval. A common mid-point gather for shot 1 is then assembled by grouping together the records of shot 1 at receiver 1, shot 2 at receiver 2 etc., where receiver 1 is nearest the ship. If the ship is steaming at a greater rate R , then the midpoints for this gather will move forward by $R\Delta S - 0.5\Delta X$ each shot.

For example, consider the beginning of profile 11. The ship was moving 70.7 m between shots while $0.5\Delta X$ was only 50 m. If a conventional gather is used the midpoints for 11 channels extend over 207 m (Figure 3.9a) so the basic premise of CDP stacking and velocity analysis is not satisfied. This is not serious if all horizons are flat but if for example, the ship is steaming downdip, the hyperbolic arrival trajectories will be steepened so as to give the appearance of lower velocities.

It is desirable to gather the traces for velocity analysis and stacking so as to minimise this effect. This can be done by taking a trace from each receiver channel as usual, with the shot number increasing with channel number, until the midpoint has advanced by about $0.5\Delta X$. The shot number is then kept constant for the next trace, so that the midpoint moves back by $0.5\Delta X$ and successive traces are gathered with the shot number one less than the channel number, until the midpoint has again advanced by about $0.5\Delta X$, when the shot number is held constant, and so on. Figure 3.9a shows a gather designed in this manner for the beginning of profile 11, in which the first 4 traces are collected with the shot number equal to the channel number, the next three with the shot number equal to one less than the channel number and the final four with the shot number equal to two less than the channel number. The total mid-point spread is 65.6 m and the gather is symmetrical about the midpoint for shot 5 and channel 6.

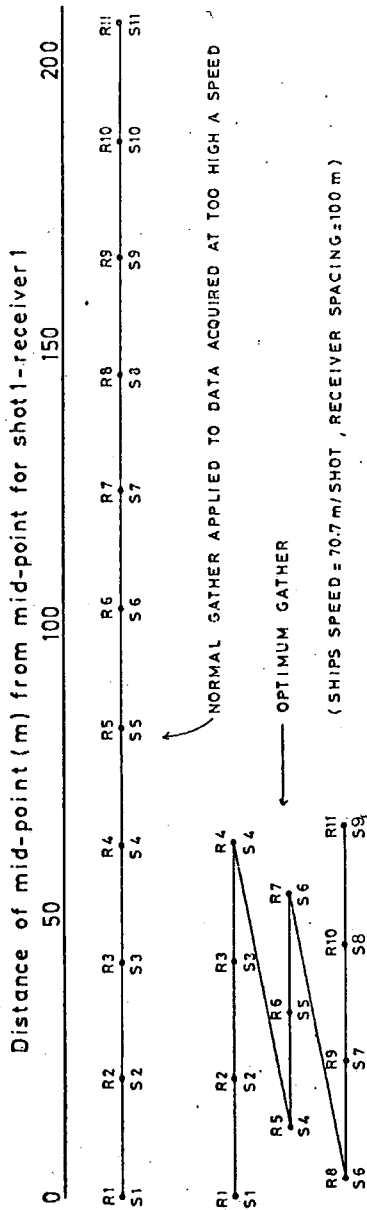


Figure 3.9a Geometry of standard and optimum CDP gathers, for 11 fold data acquired at a speed of 70.7 m/shot, with a receiver spacing of 100 m, as at the beginning of profile 11.

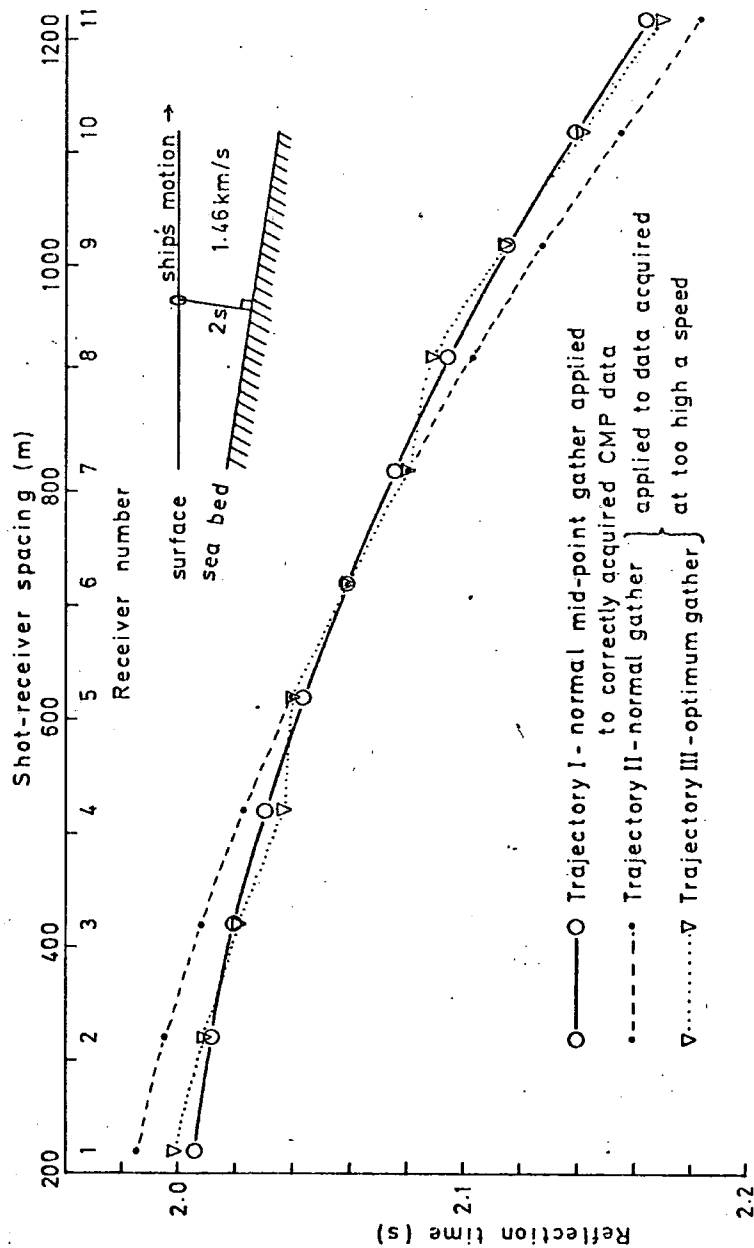


Figure 3.9b Sea bottom arrival time trajectories for the simple situation shown as an inset, with correct acquisition (I) and acquisition at too high a speed (as above), with normal (II) and optimum (III) gathers.

Figure 3.9b illustrates a comparison of sea bottom reflection time trajectories for a model based on profile 11 (about 10 km along track) in which the sea bottom dips 8.5° in the direction of the ship's motion and the two way reflection time at the origin is 2 s (sea water velocity of 1.46 km s^{-1}). Trajectory I is based on a true common mid-point gather centred over the origin, with an initial offset of 220 m and a receiver spacing of 100 m, as in the 1977 survey. The apparent velocity given by this trajectory is $1.46/\cos(8.5^\circ)$, only 1% different from the true velocity (Levin 1971). Trajectory II is based on the use of a normal gather, with a ship's speed of 70.7 m/shot. The midpoint for shot 6-receiver 6 is centred over the origin. T^2-x^2 analysis shows that this trajectory is not truly hyperbolic. Its mean velocity is significantly low at 1.3 km s^{-1} . Trajectory III is based on the more closely grouped gather described above, symmetrically placed about the origin. The trajectory is not smooth, but the points scatter about the correct curve in such a way that velocity analysis would give very nearly the correct velocity (assuming a finite pulse width).

Unfortunately, although the demultiplexing software was set up to allow any gather design, the trial section was processed using a normal gather. In consequence the velocity analyses cannot be used to give interval velocity information. However they do provide stacking velocities consistent with the gather used so that the final section is not impaired except for a certain amount of lateral averaging over about 200 m.

CDP stacking - theory

An ensemble of M seismic traces can be represented as an amplitude matrix A_{ij} . The j 'th column of this matrix is the seismic signal at shot-receiver spacing x_j , sampled at times

$$T_i = (i-1)\Delta T \quad i = 1, 2, \dots, L$$

after the shot instant. The ensemble of moveout corrected traces A^0_{ij} is given by

$$A^0_{ij} = \begin{cases} A_{Ij} & \text{for } I \leq L \\ 0 & \text{for } I > L \end{cases}$$

where $I = I(i, j)$ is the index of the sample point nearest to

the moveout trajectory and is given by

$$I = 1 + \text{INT}\{(T_i^2 + x_j^2/V_i^2)^{0.5}/\Delta T\}$$

In this expression V_i is the stacking velocity appropriate to time T_i and INT denotes the operation of rounding to the nearest integer (this can be done on the computer by adding 0.5 and truncating). The stacked trace S_i is formed simply as

$$S_i = M^{-1} \sum_{j=1}^M c_j A_{ij}^0$$

where c_j is a scaling factor, for example

$$c_j = \left\{ \sum_{i=1}^M A_{ij}^0{}^2 \right\}^{-1} \quad - \quad \text{constant energy stack}$$

$$c_j = \left\{ \sum_{i=1}^M A_{ij}^0{}^2 \right\}^{-2} \quad - \quad \text{diversity stack}$$

Velocity analysis - theory

A measure of stacking efficiency as a function of stacking velocity V and zero offset reflection time T is given by the semblance, $s(V, T)$ which is computed as follows. The moveout trajectory corresponding to V and T is evaluated and for each trace $N+1$ points are taken from a gate which is symmetrically placed with respect to the trajectory. The semblance indicates how well the signals from the M gates stack together and is calculated as the ratio of energy after stacking to the summed signal energies prior to stacking, normalised to the range (0,1); that is

$$s(V, T) = \frac{\sum_{k=-N/2}^{N/2} \left\{ \sum_{j=1}^M A_{I+k, j} \right\}^2}{M \sum_{k=-N/2}^{N/2} \sum_{j=1}^M A_{I+k, j}^2}$$

where $I = I(V, T) = 1 + \text{INT}\{(T^2 + x^2/V^2)^{0.5}/\Delta T\}$

A contour plot of $s(V, T)$ allows one to measure the stacking velocities corresponding to strong primary reflections, which show up as sharp peaks and then to construct a stacking velocity function $V(T)$.

Interpolation

In the methods described above the moveout trajectories are determined to the nearest sample point. In the case of the stacking technique it is not difficult to show that the moveout corrected signal suffers, in consequence of this imprecision, a proportional power loss at frequency f given on average by $1 - \{\sin(\pi f \Delta T) / (\pi f \Delta T)\}^2$. The lost power from all frequencies is distributed as white noise. Taking f as 62.5 Hz (the cut-off frequency of the antialiasing filter - see Figure 2.4b) and ΔT as 4 ms, the proportional power loss at the highest frequency of interest is an unacceptable 19%. A solution to this problem is to increase the sampling rate by interpolation prior to stacking. If, for example, the sampling rate is increased four times, the power loss at 62.5 Hz is only 1.3%. It is common commercial practice to increase the sampling rate sixteen times, resampling at 1/4 ms; in this case the power loss is less than 0.1%.

In the case of the velocity analysis technique the imprecision in determining the moveout trajectories causes the signals from the various gates to be slightly misaligned, on a random basis. This effect averages out over the semblance plot, and in practice little seems to be gained by resampling at an interval less than 4 ms, however the facility can be readily incorporated.

Lu and Gupta (1978) presented a Fourier domain technique for increasing the sampling rate of a time series an arbitrary number of times, without altering the shape of its spectrum. This method is based on the principle that a time shift, t , can be effected by multiplying the Fourier transform of a signal by the factor $\exp(-i2\pi ft)$. The algorithm is described in detail by Lu and Gupta, in terms which make its implementation straightforward on an array processor which has Fast Fourier Transform software.

Implementation

The first step in the CDP stacking program (CDP, Appendix B) is to set up a complex exponential array which is

permanently resident in the Array Processor (AP) memory for use in the interpolation. Traces are then brought down from disk one at a time into the PDP11 virtual memory. For each trace an index array is computed in the AP to enable the correct samples to be picked from the trace (after interpolation) to form the moveout corrected signal. The speed of the AP software is such that the square root in the moveout relationship can be calculated explicitly without the use of approximations which have hitherto been standardly employed (Waters, 1977). The trace is then transferred to the AP, where it has its mean removed and is appropriately scaled and muted (Waters, 1977) if required, following which it is interpolated and the interpolated trace is stored in virtual memory in a transposed form. A moveout corrected signal containing the same number of points as the original trace is picked from it and transferred back to the AP where it is added to a permanently resident stack. When all of the traces have been dealt with this stack is stored on disk.

A separate program (INT, Appendix B) was written to interpolate a set of traces independently and write them back to disk. This can be run prior to velocity analysis if required.

In the first part of the velocity analysis program (VEL, Appendix B) input parameters are examined for internal consistency and modified if necessary. For example, the maximum zero offset reflection time for which semblance is to be calculated will be decreased if the steepest moveout trajectory passes the end of the data set.

In the main part of VEL the quantity changed least rapidly is the zero offset reflection time, which steps upward by a constant increment. At each zero offset reflection time, the semblance is calculated for a range of stacking velocities, which also step upwards from a specified lower limit by a constant increment to an upper limit which can be made dependent on reflection time, so that semblances are not computed for high velocities at small times. The natural way to compute semblance is to iterate through the velocities, completing the semblance computation for each

velocity before moving to the next. However, given the size of the AP array memory, this system imposes undue restrictions on the number of channels which can be handled, if excessive amounts of data transfer are to be avoided. The system adopted is to iterate through the channels, performing partial semblance calculations which are completed once all channels have been dealt with. For each zero offset reflection time and channel number, there is a data window which covers all required trajectories together with a half gate width at each end. At the beginning of the partial semblance calculations for a particular channel this data window is resident in a buffer array in the host lower memory, having been transferred there from disk during the computations for the previous channel. In practice slightly more sample points are present in this buffer than are actually required, as data is copied from disk, block by block (128 words each). The data window is then transferred to the AP and the partial semblance computations are performed. For each velocity an appropriately moved-out gated signal is selected, from the data window, at the original sampling rate. Its energy is computed and added to a resident summed energy variable; the signal itself is added to a resident stack.

After the partial semblance computations have been performed for all the channels, iteration with respect to velocity yields the final semblance vector. For each velocity the energy of the corresponding stack is found and divided by the product of the number of channels and the respective summed energy, to give the required semblance. The array of semblance values appropriate to the original zero offset reflection time and range of velocities is then written to disk.

In order to facilitate the computations described above most of the important indices are computed in the AP, immediately after a new zero offset reflection time is chosen and before any data transfer or partial semblance computation takes place. For each channel the following indices are computed (the names refer to variables within VEL).

JBLK - the number of the disk block in which the

start of the data window is to be found

NWDS - the number of words to be transferred from disk to the lower memory buffer, in order to include the whole data window

JST - the address within the lower memory buffer at which the data window starts

NEL - the number of samples within the data window.

The array NMOFS contains the addresses (for each channel and each velocity) at which the gated signals start in the data window, as stored in the AP.

As an experiment a modified version of VEL was produced in which the gated signal is multiplied by a cosine tapered (Hanning) window, in the hope of producing a smoother semblance function. The effective gate width in this case is reduced to about $3/8$ of the actual value (corresponding to the ratio of the energy of a Hanning window and a unit amplitude window of the same width). Some synthetic examples are shown below.

In order to test CDP, INT and VEL a simple synthetic seismogram program was written (SYNSEI, Appendix B). SYNSEI generates a set of common midpoint traces which display a number of primary reflections, whose trajectories are specified by zero-offset time and stacking velocity. No account is taken of transmission and reflection effects, so the seismic pulse is the same for all traces and reflectors. The pulse is represented by a cosine tapered, second derivative Gaussian wavelet, described by the function

$$r(t) = \begin{cases} 0.5(1 + \cos(\pi f_m t)) (1 - 2(\pi f_m t)^2) \exp(-(\pi f_m t)^2) & \text{for } |t| \leq 1/f_m \\ 0 & \text{for } |t| > 1/f_m \end{cases}$$

where f_m is the frequency at which the spectral amplitude of the untapered wavelet is maximum. Figure 3.10a illustrates a pulse for which $f_m = 20$ Hz as in the synthetic seismograms illustrated below. The normalised power spectrum of this pulse (Figure 3.10b) shows that the spectral peak is not significantly shifted by the cosine taper.

Band limited Gaussian white noise is added independently to all traces. The Fourier transform of the noise is constructed with complex components of unit amplitude and

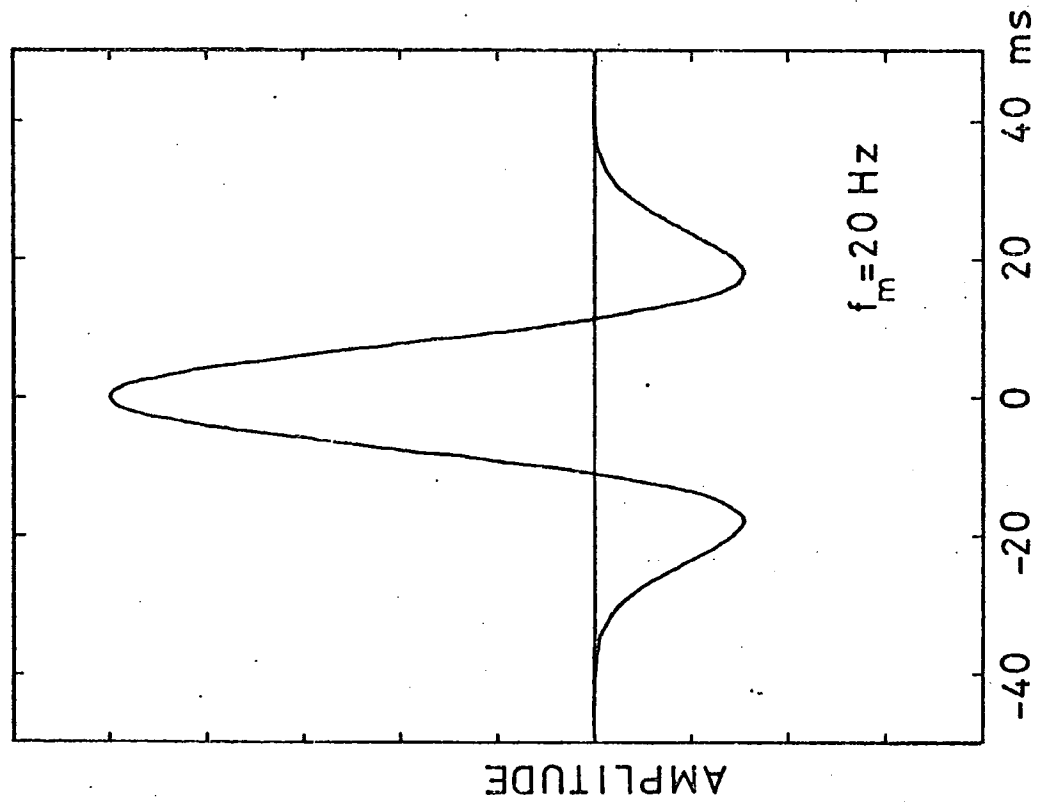


Figure 3.10 Pulse shape and power spectrum for cosine tapered second derivative gaussian wavelet which is used in the synthetic seismograms.

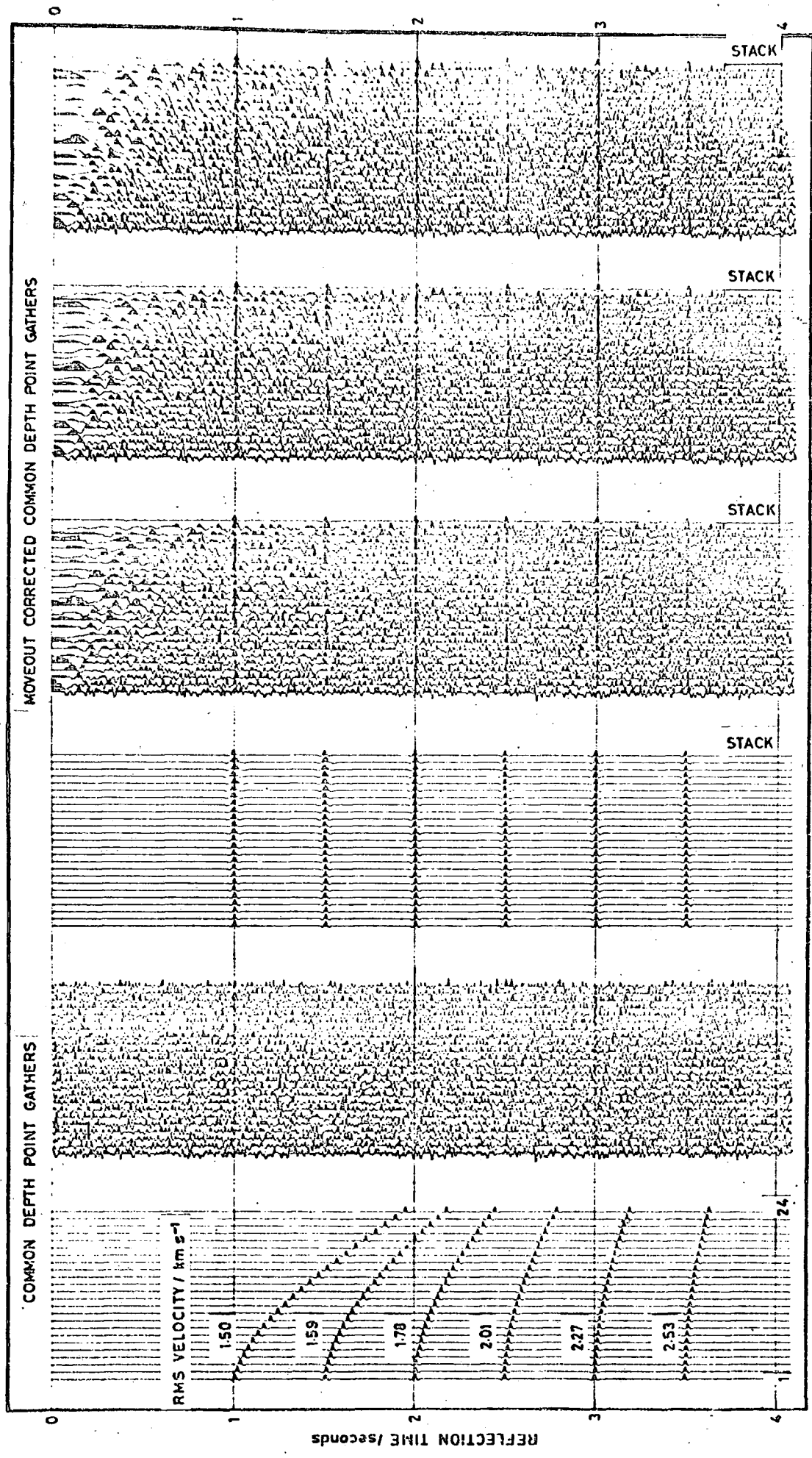
uniformly random phase, up to a specified cut-off frequency, and inverse transformation yields the requisite series.

Brief details of input for all of these programs are included in Appendix B. It is expected that the programs will undergo further modification when they are incorporated into the batch stream system, so final documentation will appear in a future work (M.J. Poulter, thesis to be submitted). None of the programs produces graphical output directly. Separate plotting programs were written by M.J. Poulter.

Synthetic Examples

The first frame of Figure 3.11 shows a synthetic seismogram with reflectors spaced at half second intervals below 1 s. The rms velocities are based on the rms equivalent of Equation 3.2, with $k=1.0 \text{ km s}^{-2}$. The shot receiver spacings are as for the 1977 survey, but there are 24 channels. The second frame shows the same seismogram with 0-62 Hz white noise added. The remaining four frames illustrate CDP stacking. In each case the first 24 traces display the moveout corrected signals, and the 25th displays the stacked signal. In the three gathers which have noise added, the resampling rate varies from 1 (that is no interpolation) through 4, to 16 times the original sampling rate. In all cases there is a very good attenuation of noise in the final gather, and the level of interpolation does not alter the quality of the stacked signal significantly. The normal moveout stretch (Dunkin and Levin, 1973) at small times and large offsets should be noted; this effect can be removed by muting (Waters, 1977). Synthetic examples with pulse frequencies as high as 62 Hz showed that 4 times interpolation gives significantly better results than no interpolation but that 16 times interpolation does not appear to be warranted.

Figure 3.12 shows some velocity analyses performed on the synthetic seismogram of Figure 3.11 (with noise). In all of the analyses the times and velocities of the maxima correspond well to the synthetic seismogram parameters. The

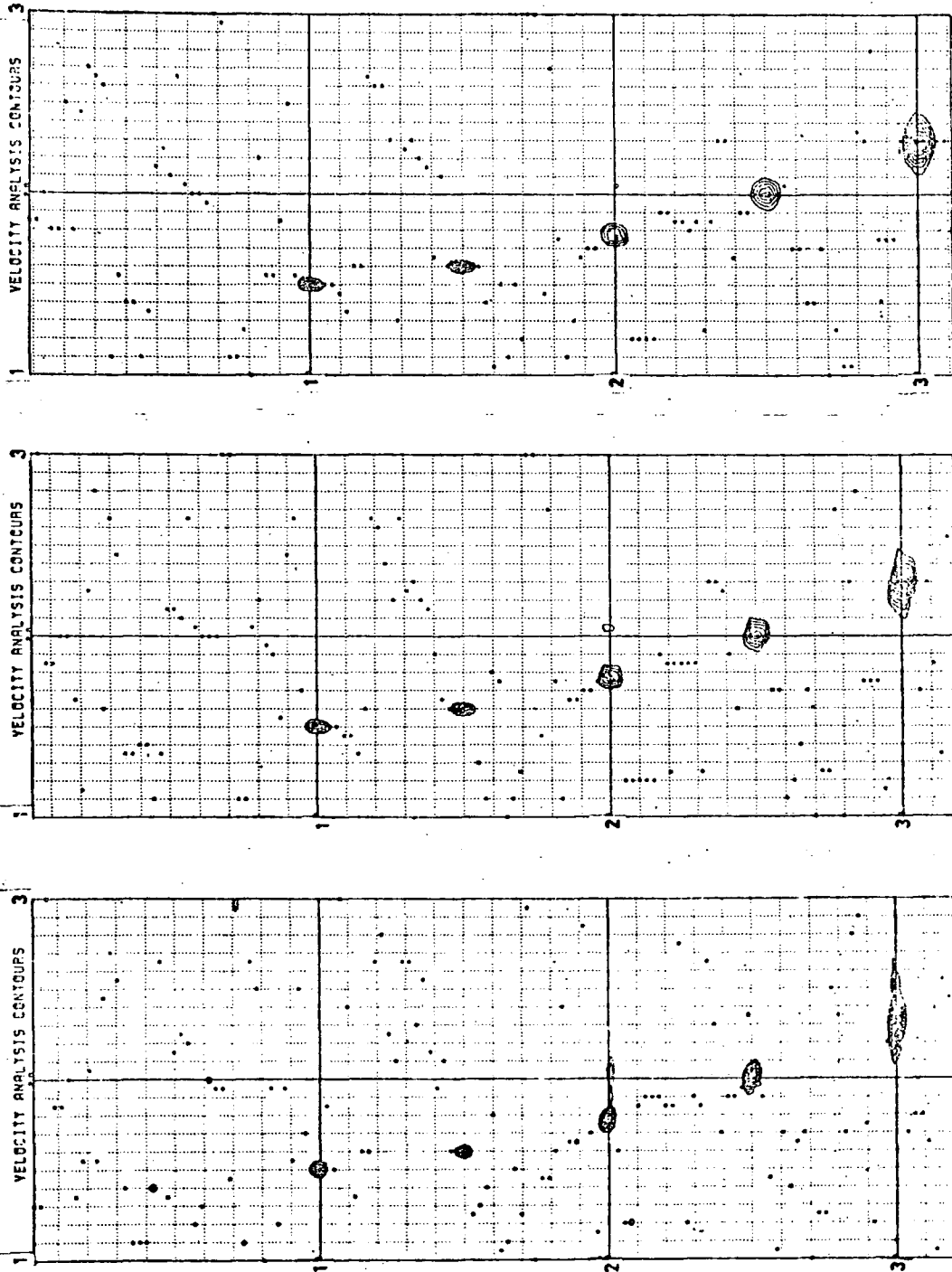


$X(1)=220\text{m}$ $XSTEP=100\text{m}$ PLUS 0-62 Hz WHITE NOISE
 SAMPLING RATE=4 ms PEAK SIGNAL/RMS NOISE=2.0

20 Hz WAVELET Figure 3.11 Synthetic seismograms illustrating CDP stacking with various levels of interpolation (plotting software provided by M.J.Poulter).

VELOCITY / km s⁻¹

VELOCITY STEP=0.05 km s⁻¹



ZERO-OFFSET REFLECTION TIME/seconds

TIME STEP=24 ms

24 ms GATE

64 ms GATE

168 ms GATE WITH HANNING WINDOW

Figure 3.12 Velocity analyses of the noisy synthetic seismogram of Figure 3.10, with no interpolation (plotting software provided by M.J.Foulter). Semblance contour interval of 0.1 with dots indicating maximum semblance for each velocity.

first analysis was performed with a gate width approximately equal to the width of the central lobe of the pulse. There is very good time resolution, but there is a tendency towards instability which degrades the velocity resolution. This is because a number of different trajectories can cross the central lobe to give roughly equal semblance. The second analysis was performed with a gate as wide as the whole pulse. This is more stable, but the time resolution is downgraded. The third analysis was performed with a Hanning window, such that the effective gatewidth was the same as in the second analysis. There is a marginal improvement in smoothness, but the analysis took much longer, as by introducing a shaped window, the number of computations performed in the most interior loop in the program is roughly doubled, and in addition the gate width has to be over twice as large. It is therefore doubtful whether the use of a window function is worthwhile, but tests on real data would be interesting. It was noted that the presence of random noise stabilises the velocity analysis. If no noise is present more spurious correlations between different pulse lobes are picked up, and in addition the semblance does not decrease until all possible trajectories are clear of the limits of the pulse by half a gate width.

Chapter 4

INTERPRETATION OF THE GRAVITY ANOMALY DATA FROM THE 1977
CRUISE

In this chapter the interpretation of gravity anomaly data acquired during the 1977 survey is described. Figure 4.1 shows the Durham profiles which have been interpreted, in relation to the free air gravity anomaly chart and interpreted profiles F, J and N of Gronlie et al. (1979).

The principal positive gravity anomalies, with the exception of anomaly "C" are related to basement ridges; the western crest of the Jan Mayen Ridge in the north and ridges 1 to 6 in the south. Relatively negative anomalies are associated with the troughs between the ridges. There is a particularly pronounced negative anomaly ("L") associated with the Jan Mayen trough but also extending further northeast.

It is of fundamental importance to know whether or not the free air gravity anomalies of the Jan Mayen block can be entirely accounted for by observed water depths and Tertiary sediment thicknesses. If the presence of pre-Tertiary sedimentary rocks could be established, the continental nature of the Jan Mayen crust would be proved. Gronlie et al. have inferred that substantial thicknesses of low density material underlie the acoustic basement of ridges 4, 5 and 6 and have identified this with pre-Tertiary sediment. This conclusion is not borne out by analysis of the Durham data.

In addition to the Jan Mayen profiles the relatively simple Norway Basin profile was also interpreted mainly for the purpose of learning something of the deep structure beneath the Norway Basin and its margins.

The gravity anomalies crossed during the 1977 cruise are generally elongate and can be considered as two-dimensional. A computer program based on Oldenburg's (1974) frequency domain gravity interpretation method was developed to strip

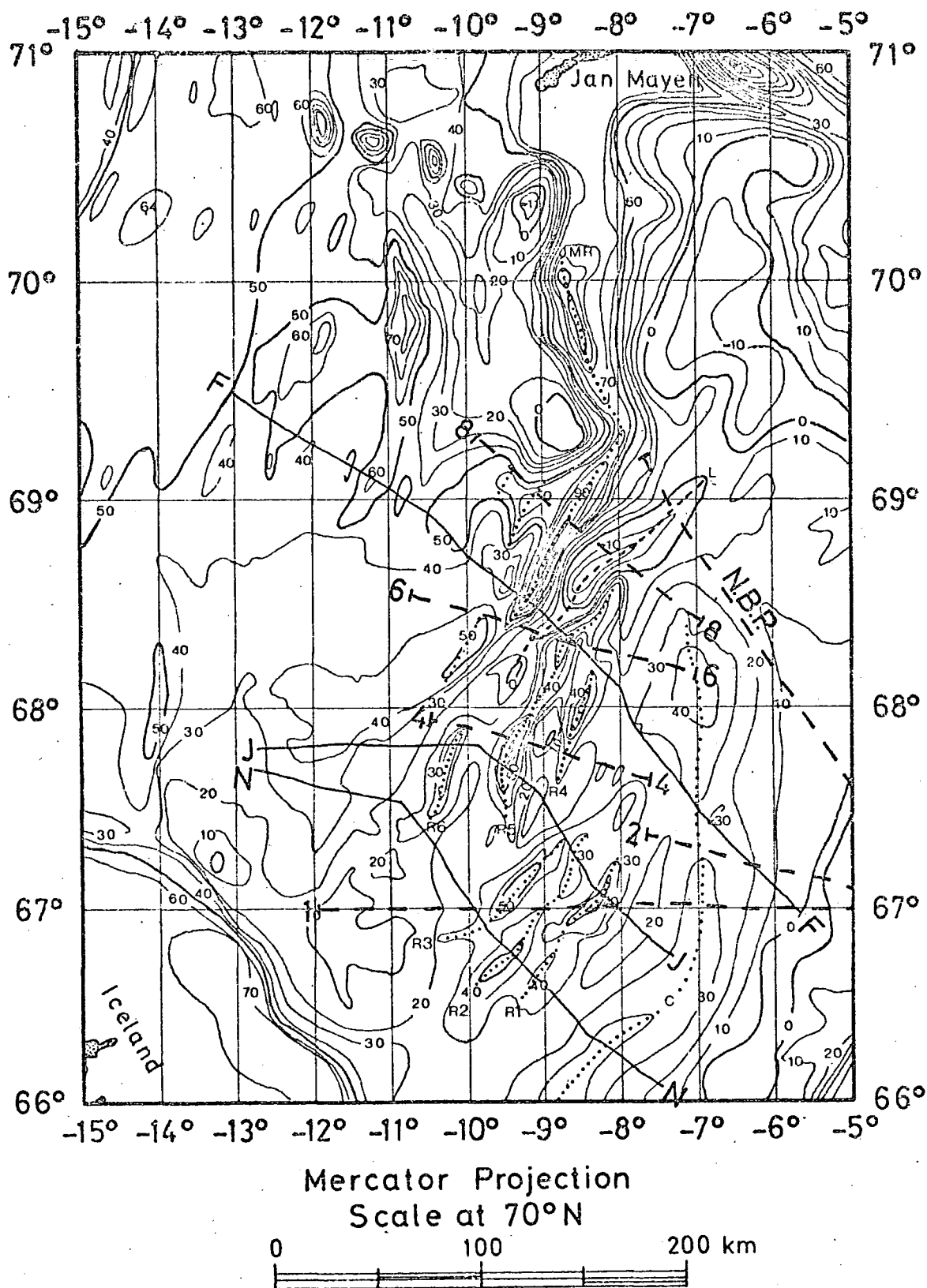


Figure 4.1. Free air gravity anomaly chart of the Jan Mayen block, showing anomalies L and C and ridges R1 to R6. Also showing the locations of the interpreted profiles F, J and N of Gronlie et al. (1979) and profiles 1 to 8 and the Norway Basin profile (NBP) of the present study (10 km intervals). Redrawn from Gronlie et al. (1979), see Figure 1.13 for further details.

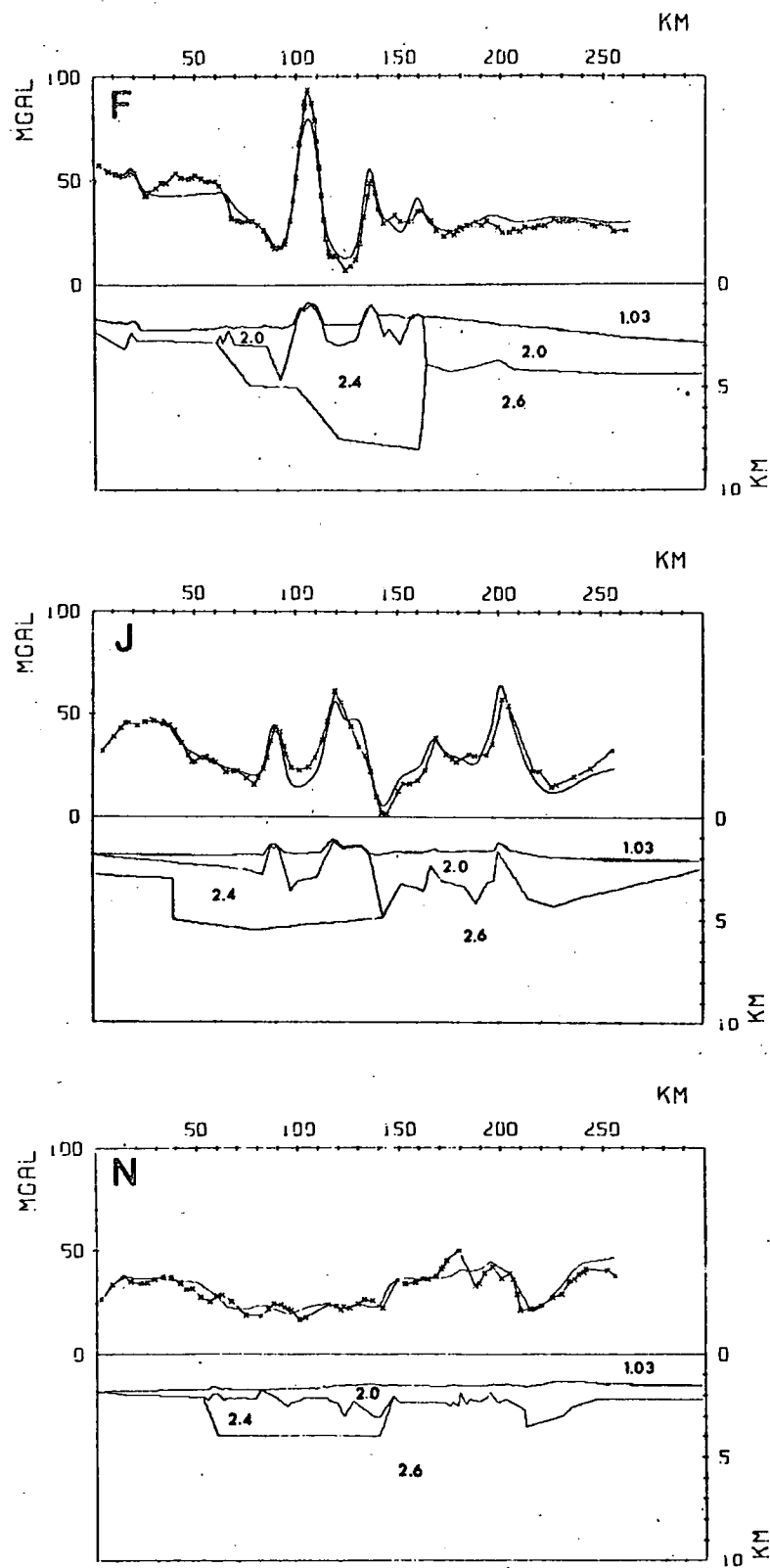


Fig. 8. Three gravity profiles with corresponding crustal models across the Jan Mayen Ridge area (for location of profiles, see Fig. 7). Observed free air gravity is shown as solid line with crosses, computed gravity effect as solid line.

Figure 4.2. Interpreted gravity profiles F, J and N of Gronlie et al. (1979). See Figure 4.1 for location of profiles.

the effect of known layers and model the residual anomalies in terms of interfaces at depth.

4.1 Frequency domain gravity interpretation

Parker (1972) showed how a series of Fourier transforms can be used to calculate the gravity or magnetic anomaly due to an uneven, non-uniform layer of material. A rearrangement of the formula for the gravity effect of a two-dimensional uneven layer leads to an iterative procedure for inverting the anomaly to give the shape of the perturbing body (Oldenburg, 1974). The theoretical discussion below mainly follows the original papers of Parker and Oldenburg.

The forward algorithm

Consider an x-z Cartesian coordinate system (Figure 4.3a, modified from Oldenburg to have z positive downwards). The gravity anomaly, $\Delta g(x)$ due to a two dimensional perturbing layer of constant density ρ , is measured at the surface, $z=0$. The perturbing layer is bounded below by the plane $z=z_0$ and above by the undulating surface $z = z_0 - h(x)$. Subject to restrictions on $h(x)$, which are satisfied in marine gravity problems, the Fourier transform of $\Delta g(x)$ is given by

$$F(\Delta g(x)) = 2\pi G \rho \exp(-|k|z_0) \sum_{n=1}^{\infty} (|k|^{n-1}/n!) F(h^n(x)) \quad 4.1$$

where G is the universal gravitational constant and the Fourier transform is defined as

$$F(f(x)) = \int_{-\infty}^{\infty} f(x) \exp(ikx) dx$$

where k is the wavenumber ($=2\pi/\text{wavelength}$). The reference level, z_0 , is arbitrary in the sense that the attraction from a uniform slab can be added or subtracted. Parker has shown that the fastest convergence results from placing the reference level midway between the highest and lowest points of the interface. The gravity anomaly $\Delta g(x)$ is evaluated by taking a finite number of terms of the series in Equation 4.1 followed by inverse Fourier transformation.

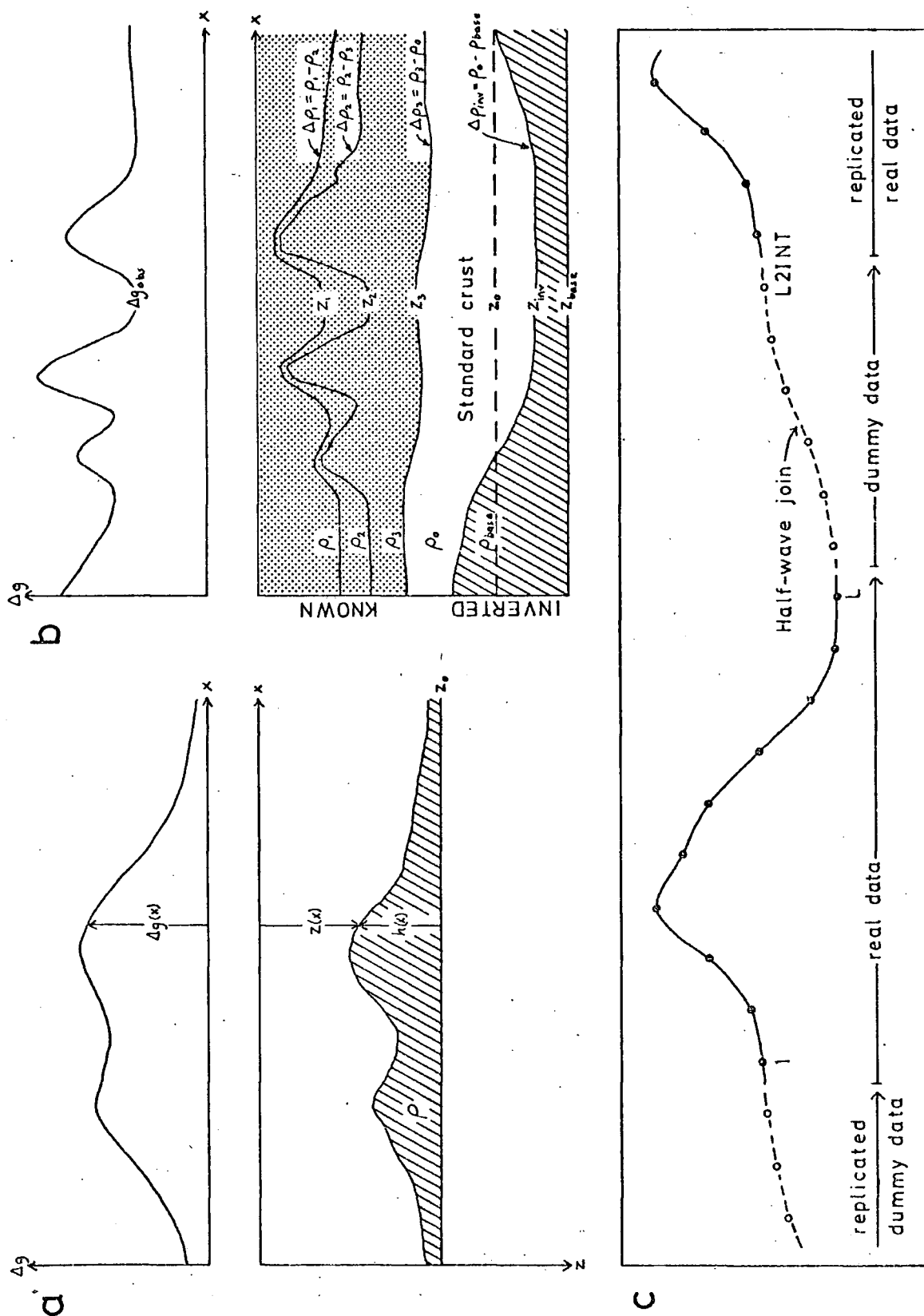


Figure 4.3a. The basic set up for the frequency domain gravity effect formula 4.1.

Figure 4.3b. The set up for gravity stripping and inversion as described in the text.

Figure 4.3c. The cosine-join data padding technique which copes with the false periodicity introduced by the use of the FFT.

The inverse algorithm

Transposing the first term of the infinite sum in Equation 4.1 and rearranging gives

$$h(x) = F^{-1}\{F[\Delta g(x)]\exp(|k|z_0)/2\pi G\rho\} - F^{-1}\left\{\sum_{n=2}^{\infty} (|k|^{n-1}/n!)F[h^n(x)]\right\} \quad 4.2$$

where F^{-1} represents the inverse Fourier transform. If ρ and z_0 are known (or assumed) this equation allows a new estimate of $h(x)$ to be generated from $\Delta g(x)$ and an old estimate of $h(x)$. Under favourable conditions, iteration in this manner, with a starting estimate $h(x)=0$, converges to give a function $h(x)$ which satisfies Equation 4.1.

The first term of Equation 4.2 represents the expansion, using the slab formula, of the equivalent density layer (Bott, 1973) which is obtained by downward continuing the original field to depth z_0 . This term is subject to short wavelength instability and must be filtered so that short wavelength noise in $\Delta g(x)$ does not become amplified excessively.

It is also necessary to filter every new estimate of $h(x)$, because the powers of $h(x)$ involved in the infinite sum in Equation 4.2 always cause leakage of energy into the short wavelength components. The explicit inversion scheme proposed by Oldenburg is

$$h_{i+1} = F^{-1}\{L(k)F[\Delta g(x)]\exp(|k|z_0)/2\pi G\rho - L(k)\sum_{n=2}^{\infty} (|k|^{n-1}/n!)F[h_i^n(x)]\} \quad 4.3$$

where $L(k)$ is a low pass filter. A suitable filter is the cosine tapered window defined by

$$L(k) = \begin{cases} 1 & |k| < 2\pi f_3 \\ 0.5[1 + \cos((k - 2\pi f_3)/2(f_4 - f_3))] & 2f_3 \leq |k| \leq 2\pi f_4 \\ 0 & |k| > 2\pi f_4 \end{cases}$$

This filter lets pass all wavelengths less than $1/f_3$, attenuates wavelengths greater than $1/f_3$ and less than $1/f_4$ and stops wavelengths greater than $1/f_4$.

The infinite sum in Equation 4.3 is absolutely and uniformly convergent for finite k , given reasonable restrictions on $h_i(x)$ (Oldenburg, 1974) but the iterative procedure as a whole may diverge. In practice, a

sufficiently severe low pass filter will ensure convergence. However the gravity anomaly calculated from the final estimate may not necessarily match the observed gravity anomalies to the desired degree of accuracy if the filtering has been too severe. Oldenburg discusses how variations of ρ , z_0 and the filter values affect the final solution.

Implementation

The FORTRAN program GRVINV incorporating the forward and inverse algorithms described above was written for the purpose of interpreting the 1977 cruise data (see listing in Appendix B).

Figure 4.3b illustrates the approach used. The gravity anomaly $\Delta g_{\text{obs}}(x)$ is known and is assumed to be caused by perturbation of a standard structure in which a density ρ_0 is maintained down to a level z_{base} , below which the density is ρ_{base} . The upper interfaces $z_1(x)$, $z_2(x)$ and $z_3(x)$, together with the densities ρ_1 , ρ_2 and ρ_3 are known or assumed and the problem is to determine the depth of the lower, compensating interface $z_{\text{inv}}(x)$.

Firstly the relative effect of the upper layers must be stripped from the observed anomaly. In order to do this a slightly modified form of Equation 4.1 was programmed (subroutine FORWD) to calculate the relative gravity effect of a layer extending from the surface to a variable depth $z(x)$, with a density contrast $\Delta\rho$ relative to the underlying material. In this case

$$z_0 = (\min(z(x)) + \max(z(x))) / 2$$

$$h(x) = z_0 - z(x)$$

and the required gravity effect is

$$g_{\text{calc}} = \Delta g(x) + 2\pi G \Delta\rho z_0$$

where $\Delta g(x)$ is as in Equation 4.1, with $\rho = \Delta\rho$.

This is a convenient formulation for gravity stripping. In the example of Figure 4.3b the total anomaly of all the upper layers, contrasted with the density ρ_0 , is given by the summed totals of the effect of interfaces $z_1(x)$, $z_2(x)$ and $z_3(x)$, computed with contrasts $\Delta\rho_1 = \rho_1 - \rho_0$, $\Delta\rho_2 = \rho_2 - \rho_0$ and

$\Delta\rho_3=\rho_3-\rho_0$ respectively. This is obvious when it is seen that the layer bounded below by $z_1(x)$ for instance, is included three times, with contrasts $\rho_1-\rho_2$, $\rho_2-\rho_3$ and $\rho_3-\rho_0$ so that its total contrast is $\rho_1-\rho_0$.

The next step is to subtract the combined effect of the upper layers from the observed anomaly to give a residual $\Delta g_{res}(x)$ which is modelled by a layer of density contrast $\Delta\rho_{inv}=\rho_0-\rho_{base}$ extending upwards from the base level z_{base} . The value of z_{base} can either be assumed from a knowledge of normal crustal structure or alternatively, if the mean level of the lower interface is known to be equal to z_{inv} over some part of the profile, z_{base} can be estimated as

$z_{inv} + \Delta\bar{g}_{res}/2\pi G\Delta\rho_{inv}$
 where $\Delta\bar{g}_{res}$ is the mean residual anomaly over the part of the profile for which the depth is known.

As the first term in Equation 4.3 involves downward continuation to the reference level z_0 , it is unsatisfactory from the point of view of numerical stability to have z_0 equal to z_{base} , if the latter is far removed from the expected level of the inverted topography. Accordingly, the mean level $\Delta\bar{g}_{res}$ (over the whole profile) is removed from the anomaly to be inverted and the reference level z_0 is put equal to $z_{base} + \Delta\bar{g}_{res}/2\pi G\Delta\rho_{inv}$.

Inversion according to Equation 4.3 (subroutine INV) then yields a topographic variation $h(x)$ which is subtracted from z_0 to give the desired interface $z_{inv}(x)$.

The last step is to compute the final calculated and residual anomalies. The gravity effect of a layer of depth $z_{inv}(x)$ and contrast $\Delta\rho_{inv}$ is found using FORWD and the slab anomaly $2\pi G\Delta\rho_{inv}z_{base}$ is subtracted from it to yield the required effect of the basal layer. This anomaly is added to the calculated anomaly due to the upper layers to give a calculated anomaly for the whole model and is subtracted from Δg_{res} to yield a final residual anomaly.

Both forward and inverse algorithms require the computation of the sum

$$\sum_{n=2}^{\infty} (|k|^{n-1}/n!) F[h^n(x)]$$

This is performed in subroutine SUMG. The chosen convergence criterion (slightly different from Oldenburg's) is

$T_N/T_2 < 0.0005$, where T_N is the magnitude of the largest real or imaginary component of $(|k|^{N-1}/N!) F[h^N(x)]$ over all k . Appropriate adjustments of levels are made before and after calling SUMG from INV so that $\max(h(x)) = -\min(h(x))$ on entry and rapid convergence is attained.

Overall convergence of the inverse procedure is assumed when the rms difference between successive estimates of topography is less than 5 m.

As suggested by Parker and Oldenburg the Fast Fourier transform (FFT) was used throughout for the computation of forward and inverse transforms. The use of the FFT introduces the following limitations

- (i) regularly spaced observations are necessary
- (ii) the Fourier integrals are represented as sums.

This is a serious defect only when the depth to an interface is less than the sample interval.

(iii) a false periodicity is introduced into the data as shown in Figure 4.3c. This has little effect if a border of appropriate dummy points is added.

The FFT algorithm used (modified from Claerbout (1976)) requires that the number of input data be an integral power of 2, L2INT. This is chosen to be the next power of 2 higher than 1.25L where L is the number of actual points, thereby ensuring at least a 25% border. In GRVINV the dummy data are constructed as shown in Figure 4.3c, by fitting a half cosine wave into the border, smoothly joining the last point to the first (in its replicated position).

Provision was made in GRVINV for inversion to give two interfaces. For instance, a substantial residual anomaly which remains after inversion to give a deep interface such as the Moho can be inverted to give the relief on an assumed shallower interface. The limitations of this approach are severe and are discussed in detail in Section 4.5 (profile

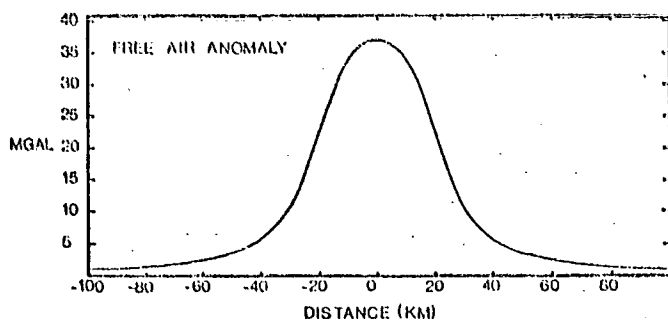
1). Alternatively, following inversion to give a shallow interface, the depth to a deeper interface may be found, to satisfy the overall mass balance down to the base level. In this instance the deeper interface must be assumed to be flat and its level could be found more simply.

Final output from GRVINV is a graphical display of the model and observed, calculated and residual anomalies together with a listing of ancillary information. Instructions for the use of the program are given in Appendix B.

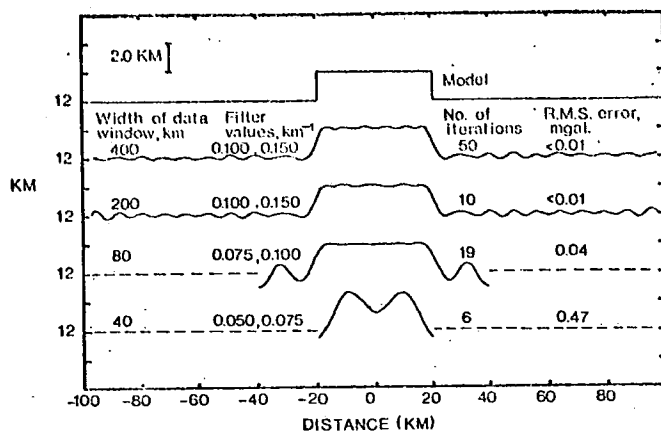
Synthetic examples of gravity anomaly inversion

Figure 4.4 shows the results of some simple tests designed to demonstrate how GRVINV copes with the anomaly due to a step in a model oceanic Moho. Given sufficiently noise free data (anomalies accurate to 0.01 mgal in this case) the method has very good resolution. Figure 4.4b shows that the method is not unduly affected by finite data length. Even a data window equal in length to the step width gives a reasonable estimate of the interface topography, although the example is artificial in that the base-level of the anomaly and topography are assumed to be known. A data window of 200 km is used in the examples of Figure 4.4c,d; the residuals are fairly evenly distributed over this range, therefore the rms residual is more meaningful than if a longer window is used. Variation of the filter parameters (Figure 4.4c) shows that the structure arising from use of the most severe filter (0.008, 0.016) gives rise to an rms residual of only 2.16 mgal (adjusted for a density contrast of -0.45 Mg m^{-3}). On account of the higher noise content of real data, which is compounded by inaccuracies in subtracting the effects of known layers, one could not expect to resolve a step in the real Moho much better than this, assuming that such a step could exist. As noted by Oldenburg (1974) the shortest wavelengths that one can resolve may not necessarily be much greater than the shortest wavelengths present at depth, in which case this limitation is not severe.

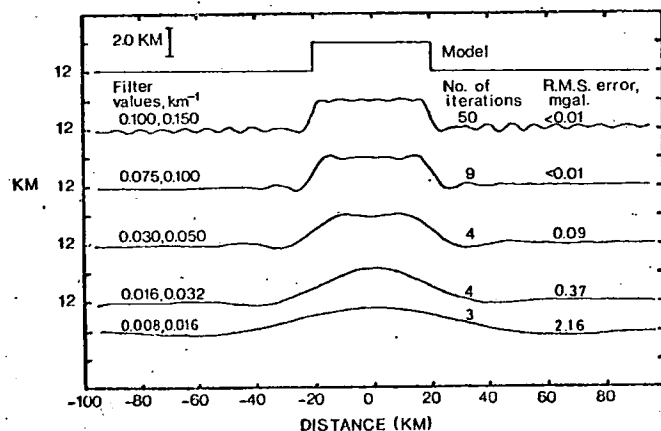
Figure 4.4d shows the effect of varying the base level



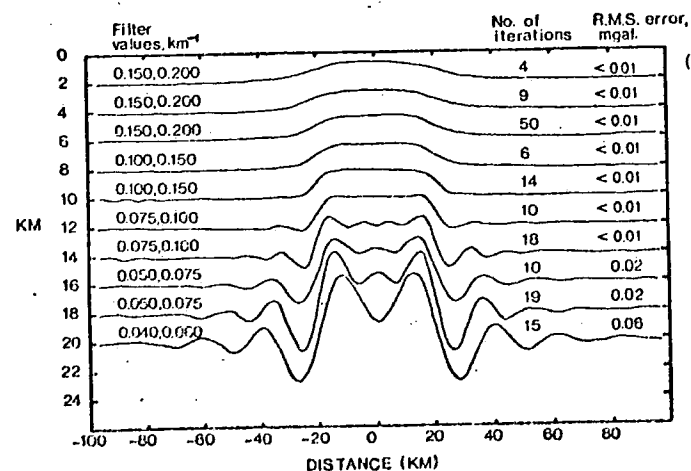
(a) Model anomaly due to a 40 km wide, 2 km high step in a "Moho" with a base level of 12 km (see below). The density contrast across the interface -3 "crust-mantle" is 0.65 Mg m⁻³ (too high to be realistic but this does not affect the tests). The anomaly was computed at 2.0 km intervals, from -400 km to +400 km, to a precision of 0.01 mgal using the program GRAVN (Bott 1975a).



(b) Illustration of the effect of finite data windows. In each case the correct density contrast and base level was used and the low pass filter was adjusted to give the maximum short wavelength content without causing divergence of the iterative process. The filter parameters are given in terms of spatial frequency (1/wavelength). In each case the smaller value refers to the upper limit of the all-pass band and the larger value refers to the lower limit of the all-cut band. Between these limits the Fourier domain filter is cosine-tapered. Convergence of the inverse procedure was assumed when the root-mean-squared (rms) difference between the results of successive iterations was less than 5 m. The number of iterations is shown, together with the rms difference between the original anomaly and the anomaly recomputed from the final estimate.



(c) Illustration of the effect of changing the severity of the low pass filter which is applied to the topography after each iteration. A data window of 200 km and the correct values of density contrast and base-level were used.



(d) Illustration of the effect of changing the base-level for inversion; for example, in the lowermost case the anomaly is assumed to be caused by perturbation of an interface which has a normal level of 20 km. A data window of 200 km and the correct value of density contrast was used. The filter values were adjusted as in (c).

Figure 4.4 Tests of the gravity inversion program GRVIN, which is based on the frequency domain method of Oldenburg (1974).

for inversion. If the assumed base level is shallower than the true base level the computed topography is less variable than the real topography. If the assumed base level significantly exceeds the true base level the inversion becomes unstable.

4.2 Data preparation for inversion programs

The system of data preparation was set up to allow any number of interfaces to be picked from the seismic records. However, using the monitor records, only the acoustic basement can be determined over any complete profile and where an intermediate interface can be picked there is usually insufficient information to allow separate seismic velocities and densities to be assigned above and below. Therefore, in practice, the total reflection time through the sediment layer (or the minimum reflection time in cases where the basement could not be seen) was picked from the monitor records at 10 minute intervals.

For each profile the reflection times were stored in a file which was input to the FORTRAN program TIMDEP (Appendix B). The purpose of this program is to create a condensed data file suitable for input to the inversion programs GRVINV and MAGINV (chapter 5). TIMDEP reads latitude, longitude and distance from the modified Merged-Merged Format working data file (Section 2.5) at 10 minute intervals and corrected depth, free air gravity anomaly and residual magnetic anomaly at 2 minute intervals. The two minute data are then filtered in the frequency domain to exclude all wavelengths shorter than 20 minutes (to avoid aliasing) and are resampled at 10 minute intervals. The reflection times are converted to depth using a constant velocity, or optionally, any specified time-depth function. In practice, Equation 3.4 was used throughout with $k=1.0 \text{ km s}^{-2}$ except for the Norway Basin profile and profile 8, for which the longitudinal variation of k given by Equation 3.3 was assumed. Depths to basement are then found by adding the sediment thicknesses to the filtered water depths (avoiding the static errors which would arise from measuring absolute reflection times from the EPC records) and all ten minute values (latitude, longitude,

distance, gravity, magnetics, depth to sea bottom and to basement) are written to the condensed data file.

A sampling rate of 10 minutes (roughly 2 km along track) was chosen for the following reasons. Firstly, the monitor records cannot be picked accurately at a closer interval. Calculations based on real examples show that topographic components with wavelengths less than 4 km contribute less than 1 mgal to the gravity anomaly measured at the surface, so a sample interval of 10 minutes is satisfactory provided that aliasing is avoided by visual smoothing. Secondly, the gravity data has a high frequency noise component (of uncertain origin) with a wavelength of about 7-8 minutes and a maximum amplitude of about 5 mgals (for example, see profile 12, Appendix A). This noise is completely eliminated, without impairing the geological information in the data, by the filtering and resampling performed by TIMDEP.

Whereas the data output by TIMDEP are equally spaced in time, the inversion programs require data equally spaced in distance. The FORTRAN program SPLINE (Appendix B) performs cubic spline interpolation of all the condensed data, outputting an equal number of data, with a constant sample interval equal to the total profile length divided by the number of intervals. In practice, because the ship's speed was constant or slowly varying along each profile, gravity and magnetic inversions using splined and unsplined data were found to be indistinguishable.

Details of the formats of the header block and body of the condensed data files are given in Appendix B. Two other small programs (not included) were written to modify these files, one to stratify the sediment layer into a number of constant density layers and the other to include the topography of an assumed upper-lower crustal interface. The choice of densities and crustal structures is described in the next section.

4.3 Discussion of densities and assumed crustal structures

Sediment densities

In the absence of other information sediment densities can be estimated from seismic compressional wave velocities using the empirical curve of Nafe and Drake (1963) which is shown in Figure 4.5a. The densities of low velocity rocks ($v=1.5 \text{ km s}^{-1}$) vary from 1.3 to 2.1 Mg/m^3 but for velocities above about 2.0 km s^{-1} density can be estimated subject to an error of about 0.1 Mg/m^3 (Bott, 1971). A convenient parametrisation of the Nafe-Drake curve is

$$v = 1.5 + 3.33(\rho - 1.5)^{2.25} \quad 4.4$$

$$\rho = 1.5 + ((v - 1.5)/3.33)^{4/9} \quad 4.5$$

for $v > 1.5 \text{ km s}^{-1}$ and $\rho > 1.5 \text{ Mg/m}^3$.

In Section 3.2 the empirical relationship

$$v = 1.5 + kT$$

was found to be satisfactory for the survey area with k varying between 0.6 km s^{-2} for deep basins and 2.4 km s^{-2} over the crest of the Jan Mayen Ridge. Assuming that $\rho > 1.5 \text{ Mg m}^{-3}$, Equations 3.5 and 4.4 together give the following relationship between depth and density

$$z = ((1.5 + 3.33(\rho-1.5)^{2.25})^2 - 2.25)/4k \quad 4.6$$

This relationship is plotted in Figure 4.5c (for $k = 0.6 \text{ km s}^{-2}$) together with density-depth profiles from DSDP holes 345 and 348 (Talwani, Udinstev et al., 1976) which were measured from cores using the GRAPE (Gamma Ray Attenuation Porosity Evaluator) technique. There is good agreement between measured and derived densities at depths greater than 300 m. At shallower depths, the measured densities are lower, except in the surficial glacial layer, which throughout the Norwegian-Greenland Sea is characterised by high densities relative to sonic velocities.

A stepped density-depth function which meets the smooth curve at 1.5 and 1.8 Mg m^{-3} but elsewhere lies below, as shown in Figure 4.5c corresponds to the real data and is convenient for gravity modelling. Extending this idea, a

method was devised for automatically laminating the sediment layer in gravity models. As illustrated in the model of Figure 4.6a, a sufficiently deep sedimentary basin is stratified into layers which have densities, from top to bottom, of 1.5, 1.8, 2.0, 2.2, 2.3 and 2.4 Mg m^{-3} , separated by interfaces at depths of $1/k$ times 0.18, 0.65, 2.53 and 3.70 km respectively; these depths being calculated from Equation 4.6. Unfortunately, no independent direct check of this density model is available for basins deeper than 800 m. However, in the case of the Norway Basin profile, gravity models which are laminated in this manner give rise to significantly smaller residual anomalies than those in which a constant density is assumed for the sediment layer.

Over some of the ridges of the Jan Mayen block, the depth to basement could not be picked from the monitor records and in these instances gravity inversions were performed to give some idea of the basement topography. In order to do this a single average density, $\bar{\rho}$ must be assigned to the sediment layer. For any given thickness of sediment, $\bar{\rho}$ is ideally equal to the total mass per unit area of the pile, divided by its thickness. If the thickness of sediment varies, the value of $\bar{\rho}$ appropriate to the deeper sections should be used. Assuming that Equations 4.5 and 3.2 hold, $\bar{\rho}$ can be evaluated in terms of kT , where T is the two way reflection time, or equivalently in terms of the seismic velocity v ($v = 1.5 + kT$), where these values apply to the base of the pile. The relationship between $\bar{\rho}$ and kT which results is shown in Figure 4.5b. One can infer, with reference back to Figure 1.12, that appropriate average densities for the sediments of the Jan Mayen Ridge range from about 2.0 Mg m^{-3} ($v = 3 \text{ km s}^{-1}$) through to at least 2.2 Mg m^{-3} ($v = 4.2 \text{ km s}^{-1}$).

Assumed oceanic crustal structure

It is now recognised that the variation of seismic velocity with depth in the basement rocks of the oceanic crust is more continuous than was hitherto supposed and that the oceanic crust probably does not consist of a small number

Figure 4.5a. The Nafe-Drake relationship between the seismic compressional wave velocity (v) and bulk density ρ of saturated crustal rocks. Redrawn from Nafe and Drake (1963). The given equation, which was determined from a log-log plot, fits the Nafe-Drake curve almost exactly for $1.5 < v < 5.0$.

Figure 4.5b. Average density $\bar{\rho}$ of a sediment pile, assuming the Nafe-Drake curve and the velocity-two way time dependence $v = 1.5 + kT$. Given in terms of the seismic velocity at the base of the pile, or equivalently, kT .

Figure 4.5c. Density-depth relationships for DSDP holes 345 (dots) and 348 (dashed line), redrawn from Talwani, Udinstev et al. (1976, pages 459 and 603). Also shown is a density-depth curve based on the Nafe-Drake relationship and a velocity-time function $v = 1.5 + 0.6T$ (dash-dot line) and a stepwise function which lies below this curve (solid line), as used in the gravity models.

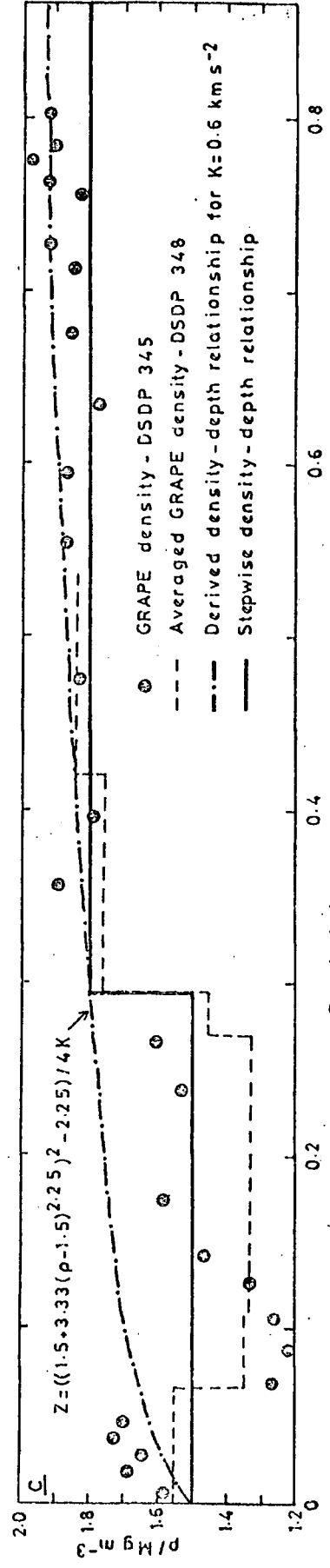
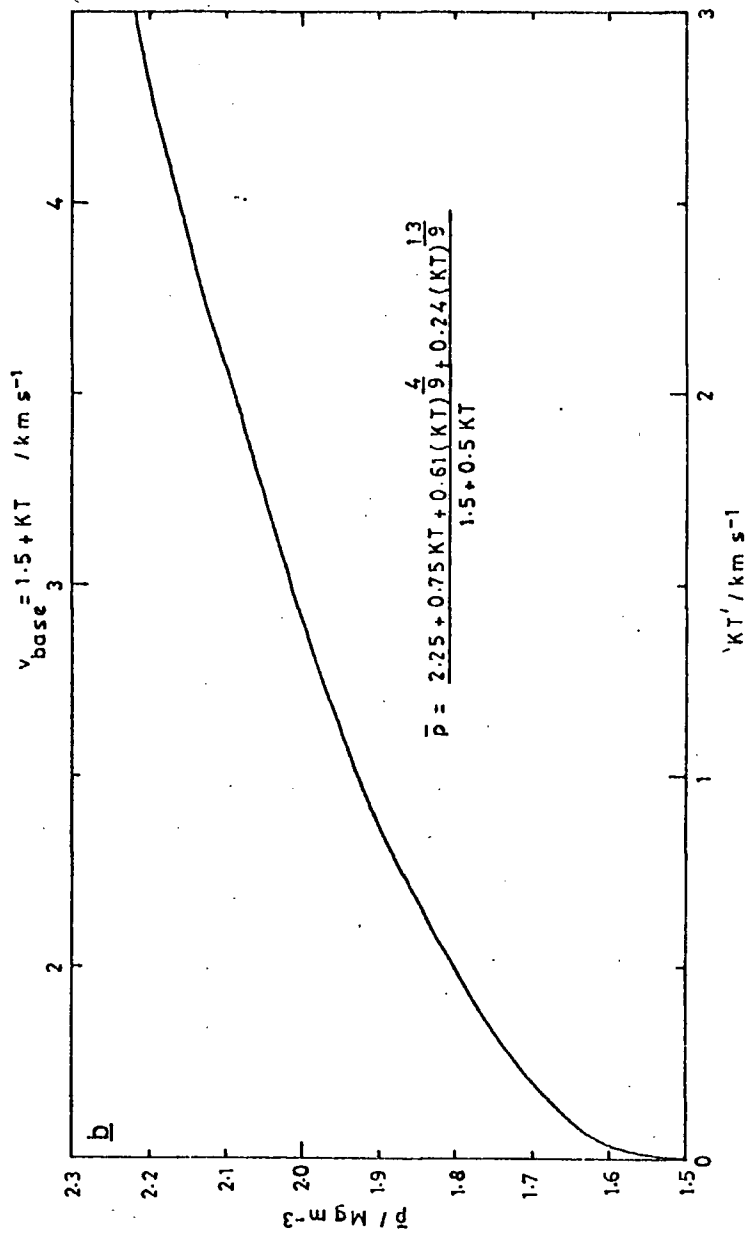
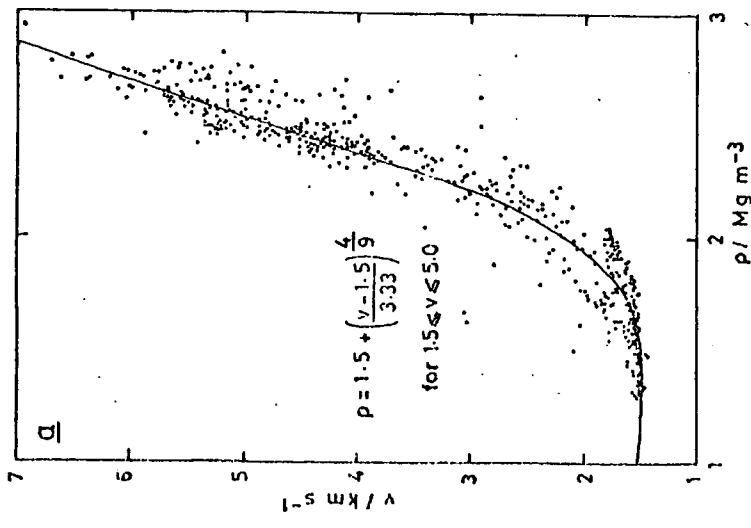


Figure 4.5

of constant velocity layers separated by first order discontinuities (Lewis, 1978). Nevertheless the original broad division into an upper layer 2 and a lower layer 3 (for example Raitt, 1963) is meaningful in the sense that layer 2 is a transitional zone in which seismic velocities increase rapidly downwards, while layer 3 is more nearly a constant velocity layer. The average thicknesses of layer 2 and 3 have been determined at about 1.5 and 5.0 km respectively (Christensen and Salisbury 1975, Woollard 1975) assuming a decomposition into, at most, four constant velocity layers (2A, 2B, 3A, 3B). Allowing for a continuous variation of velocity with depth increases the typical thickness of layer 2 to about 2 km (Lewis, 1978; Drury, 1979), but does not appreciably alter the thickness of layer 3.

Petrological models of the oceanic crust and mantle have been based on seismic velocities, marine dredge and core samples, studies of onshore ophiolite complexes and other considerations such as heat flow (Clague and Straley, 1977).

Layer 2 is generally considered to be basaltic and to be composed of pillow basalts and volcanic rubble, underlain by massive lava flows then a sheeted dyke complex. The increase of velocity with depth in layer 2 is believed to be due mainly to a downward decrease in porosity associated with which there is a downward increase in density.

The assignation of an average density to layer 2 for the purposes of gravity interpretation is a difficult problem. The chosen value, in this study, is 2.7 Mg m^{-3} , which, given a grain density of 2.95 Mg m^{-3} for oceanic basalts (Hyndman and Drury, 1976), corresponds to a porosity of 13%, which is in accord with Drury's (1979) estimate of the average porosity of layer 2. According to Christensen and Salisbury's (1975) non-linear relationship between density and velocity of DSDP basalts at 0.5 kbar, a density of 2.7 Mg m^{-3} corresponds to a seismic velocity of 5.3 km s^{-1} . A study of the profiles presented by Lewis (1978) shows that this is a realistic average velocity for Layer 2.

The refraction profile 3 of Hinz and Moe (1971) in the

western Norway Basin showed a lower average velocity of about 3.9 km s^{-1} , implying a considerably lower average density. This velocity is anomalously low with respect to crustal age (Houtz and Ewing, 1976) and it is not considered reasonable to adopt a lower density value on the basis of this single profile.

It is also necessary, for the purposes of gravity interpretation, to make some assumptions about the topography of the layer 2 - layer 3 interface, as it is not feasible to invert for both this interface and the Moho. In the models the interface was constructed by adding a constant 2 km to the basement depth, then smoothing in the frequency domain to remove all wavelengths shorter than 20 km.

There is general agreement (Clague and Straley, 1977) that the top part of layer 3 consists of gabbro or metagabbro. These rocks have grain densities of about 3.0 Mg m^{-3} (Hyndman and Drury, 1976) and the bulk density of the top part of the layer is expected to approach this value, as the typical seismic velocity of 6.8 km s^{-1} indicates negligible porosity (Christensen and Salisbury, 1975).

In ophiolite complexes the gabbroic layer is underlain by first a cumulate ultramafic layer, then a tectonised ultramafic layer. The interface between the latter two layers is thought to correspond to the boundary between differentiated crust and depleted upper mantle. Clague and Straley (1977), followed by Lewis (1978), have proposed that in the ocean crust, as in ophiolite complexes, the cumulate ultramafic layer and part of the underlying tectonised ultramafic layer are partly serpentised, lowering their seismic velocities to crustal values. In this model the Moho lies somewhat below the base of the differentiated crust and corresponds to the boundary between fresh mantle rocks and mantle rocks which have been converted to the crust by the process of serpentisation. According to Lewis (1978) varying degrees of serpentisation explain the different lower-crustal velocity profiles which have been obtained by detailed travel time inversion and seismogram

modelling. At one extreme, velocities sometimes increase uniformly throughout layer 3 to attain typical mantle velocities of circa 8.2 km s^{-1} with no discernable Moho discontinuity, at the other extreme, there may be a low velocity (6.5 km s^{-1}) channel at the base of the crust. In the Norway Basin, profile 3 of Hinz and Moe (1971) indicates the former velocity structure, although no extremal bounds are given.

The density chosen for layer 3 was 2.9 Mg m^{-3} , which is a realistic average value only if the lower crust is quite extensively serpentinised. If the lower crust is composed of fresh ultramafics (for example Drury, 1979) then layer 3 must have an average density of at least 3.0 Mg m^{-3} , possibly as high as 3.2 Mg m^{-3} . The chief virtue of the low value of 2.9 Mg m^{-3} , and the reason for which it was chosen, is that it is a conventional value (for example see interpretations from a variety of sources in Bott (1971)) and the overall mean density of 2.84 Mg m^{-3} (layer 2 and layer 3) corresponds to a commonly used "standard crustal density" (for example Worzel (1974)).

Assumed mantle density

A density of 3.35 Mg m^{-3} was chosen to represent the upper mantle; again a commonly assumed value, lying mid-way between the limits of 3.30 and 3.40 Mg m^{-3} (Bullen 1975). The effect of increasing the layer 3 density to 3.0 Mg m^{-3} would be to change the model Moho topography in a predictable manner, increasing the relative relief by a factor of 1.3.

Assumed crustal structure for the Jan Mayen block

Upper and lower crustal densities of 2.7 and 2.9 Mg m^{-3} can also be used for the supposed continental crust of the Jan Mayen block. A density of 2.7 Mg m^{-3} is appropriate for the basement beneath Horizon 0, whether this represents flood basalt (in which case an average density similar to that of oceanic layer 2 would be expected), indurated sediment or acidic crystalline basement (Dobrin 1976). It is also consistent with the mean refraction velocity of 5.2 km s^{-1}

observed below Horizon 0 (see Figures 1.12 and 4.5a).

In the models illustrated below, a single interface was used to represent the oceanic layer 2 - layer 3 boundary and the upper - lower crustal boundary beneath the Jan Mayen block, where it is maintained at a level of 7 km. The choice of this level was fairly arbitrary; it lies beneath all known sedimentary basins and in most of the models gives an average crustal density of about 2.84 Mg m^{-3} .

In general, a step had to be introduced between the oceanic and continental interfaces. This was placed at the apparent boundary between oceanic and continental crust as determined from the seismic records (Section 3.1) and was low pass filtered along with the layer 2 - layer 3 interface. In most models the upper crustal step, so positioned, is fortuitously coincident with the step in the model Moho and contributes in sympathy to the calculated gravity anomaly. In the case of profile 4 - profile 2 the upper crustal step lies to the west of the Moho step and simpler models would have been observed if it had been shifted. However the gravity effect of the upper crustal step is not large in any of the models.

4.4 Description of gravity models

Gravity models for the 1977 cruise data are shown in Figures 4.6 to 4.13. The Norway Basin profile will be discussed first, then the Jan Mayen profiles from north to south.

The Norway Basin profile (Figures 4.6 to 4.8)

The most satisfactory model which was obtained for the Norway Basin profile is illustrated in Figure 4.6a. The basement topography in this model, as in the others, was determined from the seismic monitor records. The sediment layer was stratified automatically and the upper-lower crustal interface was assumed to be at an average depth of 2 km below the basement under the Norway Basin and at a constant depth of 7 km under the Jan Mayen block, as

Figures 4.6 to 4.8. Gravity models for the Norway Basin profile (see Figures 2.2 and 2.3 for location of the profiles). Figure 4.6a shows the preferred model. The other models exemplify specific points as described in the text. The computer generated captions contain a listing of input and output parameters (symbols as used in the text, Section 4.1). Additional annotation has been added to aid interpretation. Annotated densities are in Mg m^{-3} .

Figure 4.6a Density stratified sediment layer.
Undulating layer 2-layer 3 boundary.
"Inverted" Moho, shortest wavelength 62 km.

Figure 4.6b Density stratified sediment layer.
Undulating layer 2 - layer 3 boundary.
Flat Moho, to illustrate residual due to deep structure.

Figure 4.7a Constant density sediment layer, 1.8 Mg m^{-3} .
Undulating layer 2-layer 3 boundary.
"Inverted" Moho, shortest wavelength 62 km.

Figure 4.7b Constant density sediment layer 2.2 Mg m^{-3} .
Undulating layer 2-layer 3 boundary.
"Inverted" Moho, shortest wavelength 62 km.

Figure 4.8a Density stratified sediment layer.
Flat layer 2-layer 3 boundary
"Inverted" Moho, shortest wavelength 62 km.

Figure 4.8b Density stratified sediment layer
Undulating layer 2-layer 3 boundary.
"Inverted" Moho, shortest wavelength 31 km

described in Section 4.3. The Moho topography was obtained by inversion of the residual anomaly which was left after the effect of the upper layers had been subtracted. The shortest wavelength present in the topography is 62.5 km and the shortest wavelength unaffected by filtering is 125 km (in this model as in all others $f_3 = 0.5f_4$, following Oldenburg 1974). The depth to the Moho was fixed at approximately 11 km over the central part of the basin, as indicated by the dotted line in Figure 4.6a, which resulted in a layer 3 thickness of 5 km. The base level for the inversion, 28.2 km is not of fundamental significance unless one can be sure of the densities chosen. However, in order for interpretations of nearby profiles to be consistent it is necessary to use the same base level.

The most obvious point to make about the interpretation is that most of the prominent anomalies are due to variations in water depth and sediment thickness. This is illustrated by Figure 4.6b in which the model Moho has been constrained to be flat so that the residual anomaly is that which is inverted in Figure 4.6a to give the Moho topography. This residual is relatively smooth and is only weakly correlated with the calculated anomaly due to the upper layers. In view of the fact that the basement topography was not expressly modified to give a good fit between observed and calculated anomalies, the rms residual of 2.02 mgal attained in Figure 4.6a is considered quite satisfactory.

The main anomalies along the line are the negative anomaly L at the western end, the positive anomaly C flanking it and the negative anomaly over the Aegir axis. Anomaly L is due to a deep sedimentary basin. The positive anomaly C is due to a combination of gradual effects rather than any single positive structure. The central negative over the Aegir axis is seen to be almost entirely due to the sediment thickness in the rift.

Figures 4.7a and 4.7b illustrate the need for density stratification within the sediment layer if a uniformly good fit is to be obtained for the whole profile. In the model of Figure 4.7a a constant sediment density of 1.8 Mg m^{-3} gives a

good fit over the central part of the basin, where the sediment is relatively thin, but a poor fit over the deep sedimentary basin at the western end of the line. If a higher density of 2.2 Mg m^{-3} is used (Figure 4.7b) the reverse applies. The overall fit obtained in the density stratified model of Figure 4.6a is substantially better than in either of these models.

The assumed topography of the upper-lower crustal interface does not contribute substantially to the calculated anomaly. In the model of Figure 4.8a this boundary was replaced by one which is flat, at a depth of 6 km. Both residual anomaly and the model Moho topography are very similar to those of Figure 4.6a.

In Figure 4.8b the model is identical to that of Figure 4.6a except that the shortest wavelength allowed in the Moho is 31 km. The rms residual is reduced to 1.15 mgal, but it is considered that most of the extra relief on the Moho is spurious, and results from the inversion of small anomalies which are due to inaccuracies in the upper part of the model.

An interesting feature of the model Moho topography in Figure 4.6a is the relative depression underneath the central part of the Norway Basin. This arises from the relatively negative residual of about 20 mgal, as shown in Figure 4.6b, and is persistent in all models. This might be due to a region of remanent low density associated with the extinct axis.

Profile 8 (Figure 4.9)

Figure 4.9a illustrates an attempt to model profile 8 using the same approach which was applied to the Norway Basin profile, with the base of the sediment layer being assumed to correspond to the deepest reflector which was picked from the reflection records. There are large residuals despite the undulations in the model Moho. The reason for this is clear if reference is made to the line drawing of Figure 3.1. The deepest reflector which could be reliably picked corresponds to acoustic basement only as far as 35 km along the track.

Figures 4.9 to 4.13. Gravity models for the Jan Mayen block. W-water layer, S-sediment layer, UC-upper crust, LC-lower crust, M-mantle. Densities indicated in Mg/m^3 .

Figure 4.9. Profile 8.

- (a) density stratified sediment layer but inadequate determination of basement, with inversion to give Moho.
- (b) inversion to give basement topography, no density contrast across "basement" as picked from seismic records

Figure 4.10. Profile 6.

- (a), (b) as for Figure 4.9.

Figure 4.11. Profile 4.

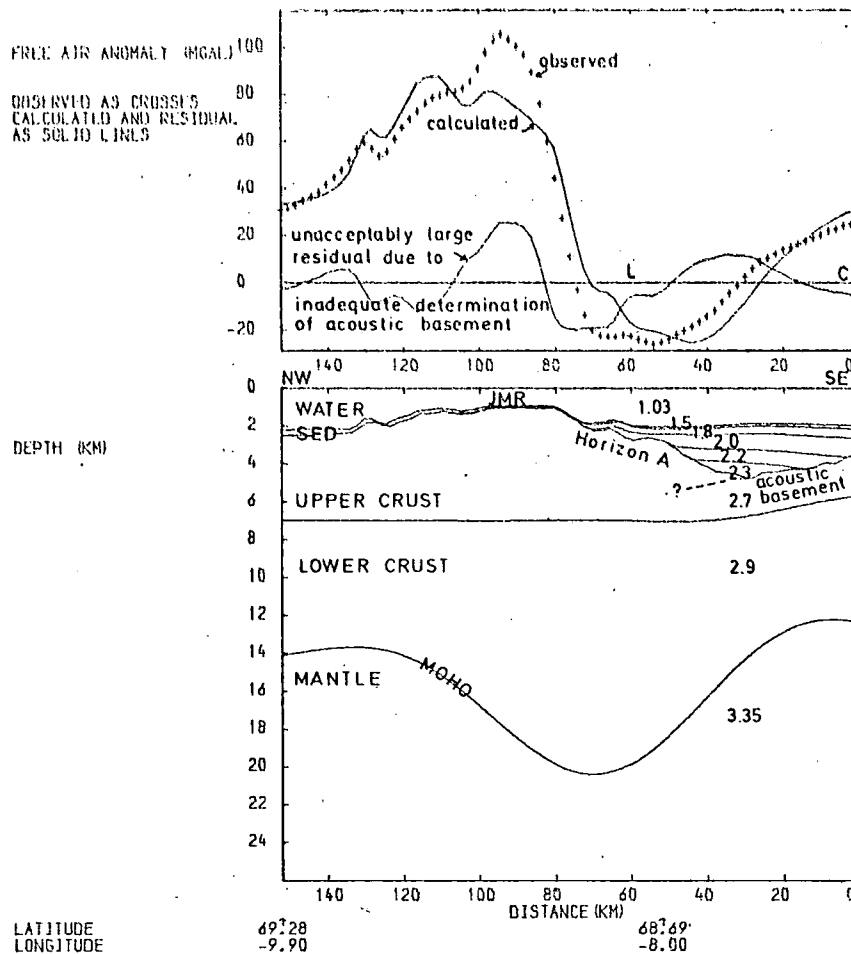
- (a), (b) as for Figure 4.9b, with differing sediment densities.

Figure 4.12. Profile 2

- (a) stratified sediment layer with adequate determination of basement. Inversion to give Moho, with minimum wavelength of 31 km.
- (b) as for (a) with minimum Moho wavelength of 62 km.

Figure 4.13. Profile 1

- (a) density stratified sediment layer. Inversion to give Moho. Unacceptable residuals because of low density "basement".
- (b) as in (a), with second inversion to give interface below "basement".



PROFILE 8

AVERAGE BEARING
OF PROFILE= 130.00°

SAMPLE SPACING= 2.003 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES,

$\Delta\rho_1 = -0.470 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_2 = -0.300 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_4 = -0.200 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_5 = -0.100 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_6 = -0.400 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_7 = -0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1

MEAN DEPTH= 15.9 km

S.D. OF DEPTH= 2.7 km

$\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

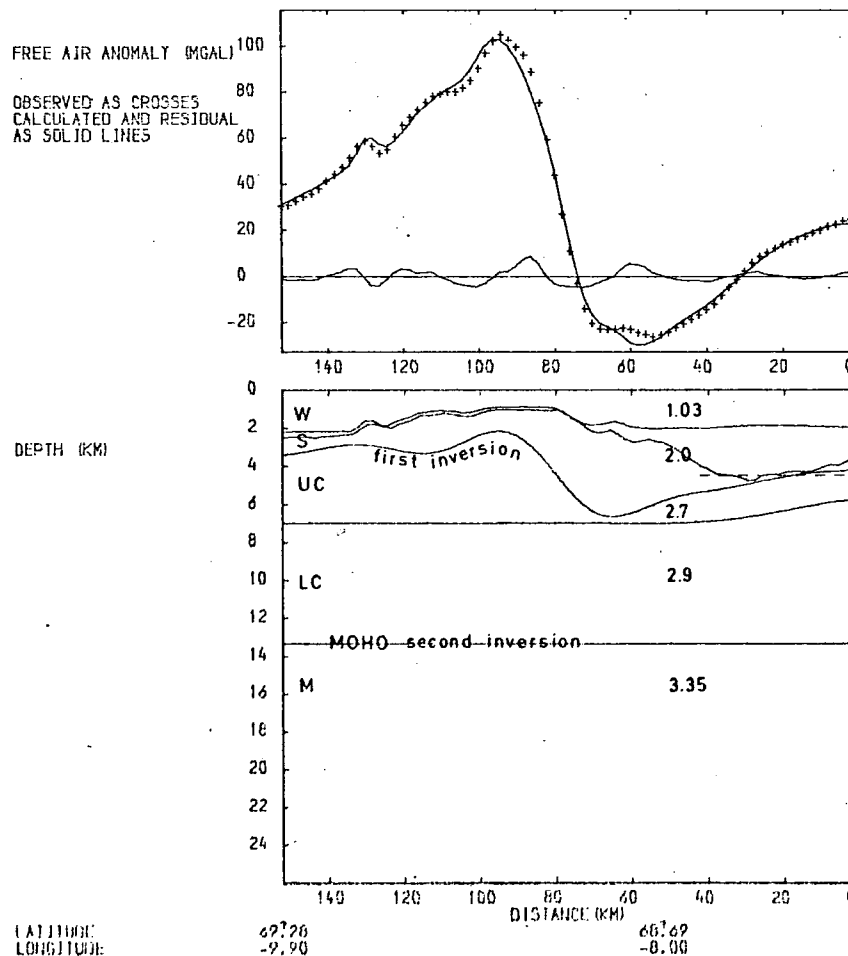
F3= 0.0080 km⁻¹

F4= 0.0160 km⁻¹

NO. OF ITERATIONS= 4

RMS RESIDUAL= 11.10 mgal

BASE LEVEL= 28.2 km



PROFILE 8

AVERAGE BEARING
OF PROFILE= 130.00°

SAMPLE SPACING= 2.003 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES,

$\Delta\rho_1 = -0.970 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_2 = 0.000 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1

MEAN DEPTH= 4.2 km

S.D. OF DEPTH= 1.3 km

$\Delta\rho = -0.700 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

F3= 0.0200 km⁻¹

F4= 0.0400 km⁻¹

NO. OF ITERATIONS= 5

RMS RESIDUAL= 2.80 mgal

INVERTED LAYER 2

MEAN DEPTH= 13.4 km

S.D. OF DEPTH= 0.0 km

$\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

F3= 0.0000 km⁻¹

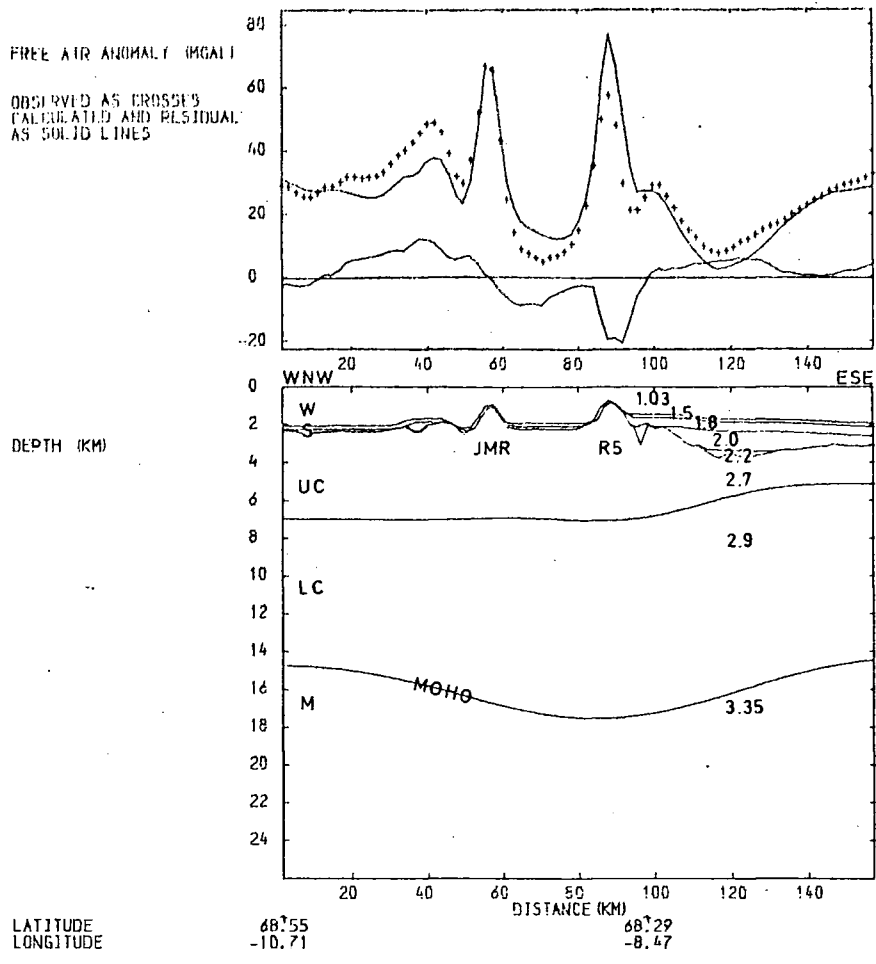
F4= 0.0000 km⁻¹

NO. OF ITERATIONS= 1

RMS RESIDUAL= 2.80 mgal

BASE LEVEL= 28.2 km

Figure 4.9



PROFILE 6 a

AVERAGE BEARING
OF PROFILE = 105.00°

DATA PROJECTED ONTO
LINE BEARING 115.00°

PROJECTED
SAMPLE SPACING = 1.895 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES,

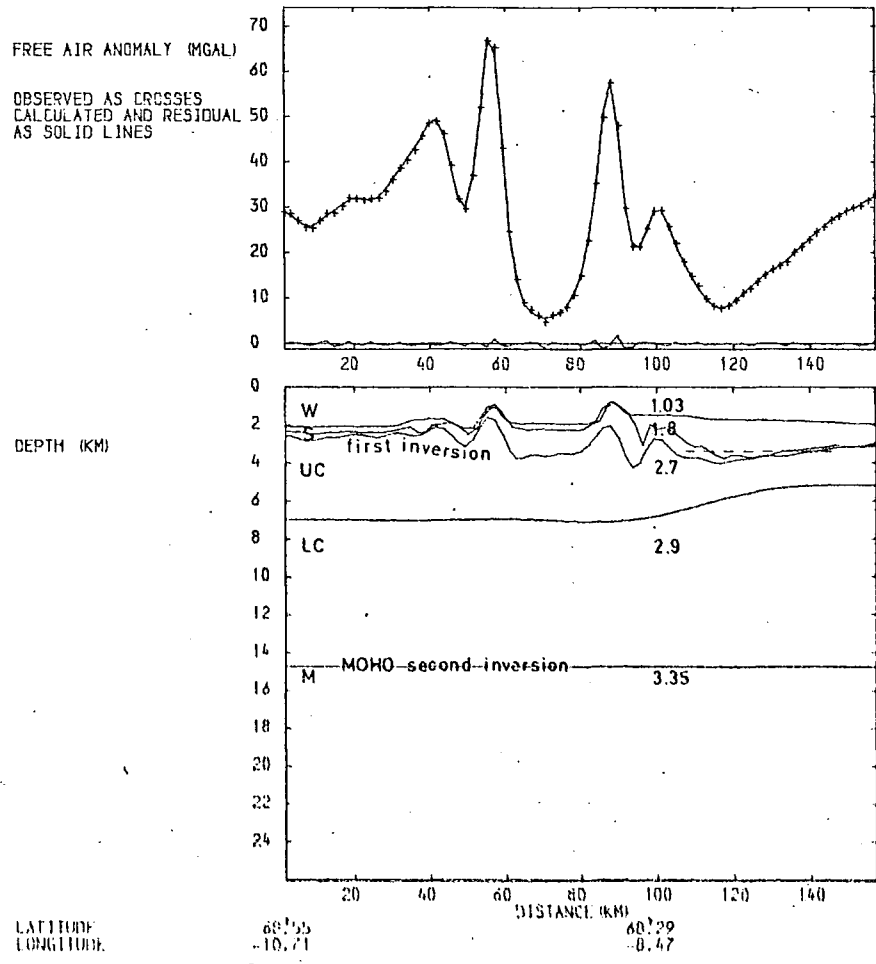
$\Delta\rho_1 = -0.470 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_2 = -0.300 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_4 = -0.200 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_5 = -0.500 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_6 = -0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1
 MEAN DEPTH = 16.1 km
 S.D. OF DEPTH = 1.0 km
 $\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES
 $F3 = 0.0080 \text{ km}^{-1}$
 $F4 = 0.0160 \text{ km}^{-1}$

NO. OF ITERATIONS = 3
 RMS RESIDUAL = 6.71 mgal

BASE LEVEL = 28.2 km



PROFILE 6 b

AVERAGE BEARING
OF PROFILE = 105.00°

DATA PROJECTED ONTO
LINE BEARING 115.00°

PROJECTED
SAMPLE SPACING = 1.895 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES,

$\Delta\rho_1 = -0.770 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_2 = 0.000 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1
 MEAN DEPTH = 3.1 km
 S.D. OF DEPTH = 0.6 km
 $\Delta\rho = -0.900 \times 10^3 \text{ kg/m}^3$

FILTER VALUES
 $F3 = 0.0000 \text{ km}^{-1}$
 $F4 = 0.1600 \text{ km}^{-1}$

NO. OF ITERATIONS = 8
 RMS RESIDUAL = 0.39 mgal

INVERTED LAYER 2
 MEAN DEPTH = 14.7 km
 S.D. OF DEPTH = 0.0 km
 $\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES
 $F3 = 0.0000 \text{ km}^{-1}$
 $F4 = 0.0000 \text{ km}^{-1}$

NO. OF ITERATIONS = 1
 RMS RESIDUAL = 0.39 mgal

BASE LEVEL = 28.2 km

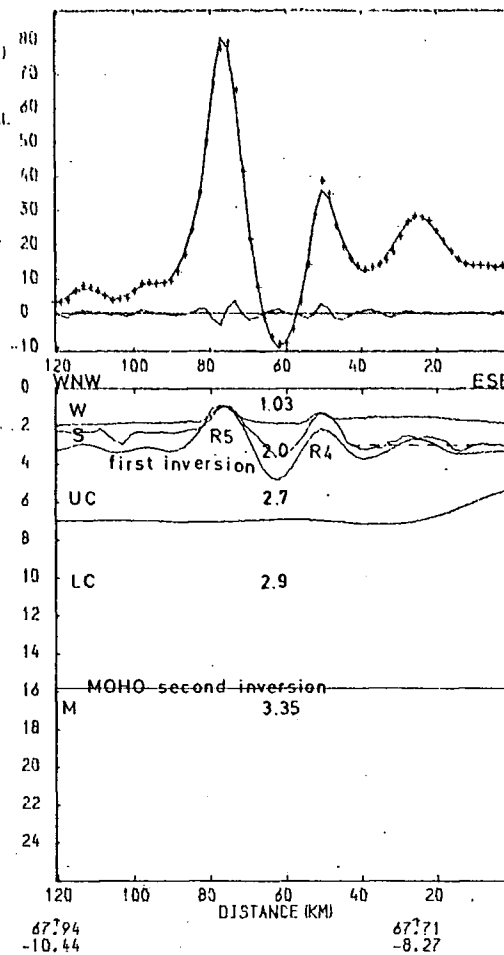
Figure 4.10

FREE AIR ANOMALY (MGAL)

OBSERVED AS CROSSES
CALCULATED AND RESIDUAL
AS SOLID LINES

DEPTH (KM)

LATITUDE
LONGITUDE



PROFILE 4

AVERAGE BEARING
OF PROFILE= 104.00°

DATA PROJECTED ONTO
LINE BEARING 120.00°

PROJECTED
SAMPLE SPACING= 1.840 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES,

$\Delta\rho_1 = -0.470 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_2 = 0.000 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1

MEAN DEPTH= 3.1 km

S.D. OF DEPTH= 0.8 km

$\Delta\rho = -0.700 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

F3= 0.0500 km⁻¹

F4= 0.1000 km⁻¹

NO. OF ITERATIONS= 8

RMS RESIDUAL= 1.04 mgal

INVERTED LAYER 2

MEAN DEPTH= 15.8 km

S.D. OF DEPTH= 0.0 km

$\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

F3= 0.0000 km⁻¹

F4= 0.0000 km⁻¹

NO. OF ITERATIONS= 1

RMS RESIDUAL= 1.04 mgal

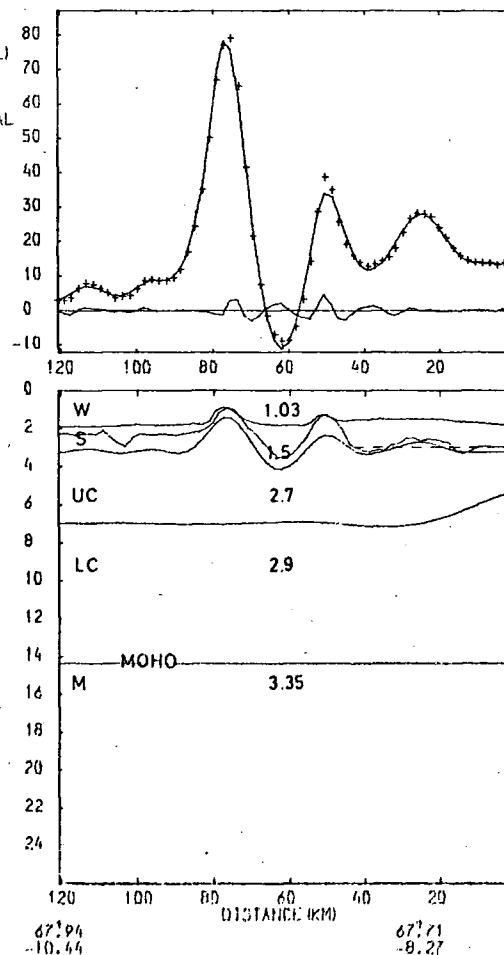
BASE LEVEL= 28.2 km

FREE AIR ANOMALY (MGAL)

OBSERVED AS CROSSES
CALCULATED AND RESIDUAL
AS SOLID LINES

DEPTH (KM)

LATITUDE
LONGITUDE



PROFILE 4

AVERAGE BEARING
OF PROFILE= 106.00°

DATA PROJECTED ONTO
LINE BEARING 120.00°

PROJECTED
SAMPLE SPACING= 1.840 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES,

$\Delta\rho_1 = -0.470 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_2 = 0.000 \times 10^3 \text{ kg/m}^3$

$\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1

MEAN DEPTH= 3.0 km

S.D. OF DEPTH= 0.5 km

$\Delta\rho = -1.200 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

F3= 0.0500 km⁻¹

F4= 0.1000 km⁻¹

NO. OF ITERATIONS= 6

RMS RESIDUAL= 1.26 mgal

INVERTED LAYER 2

MEAN DEPTH= 14.4 km

S.D. OF DEPTH= 0.0 km

$\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

F3= 0.0000 km⁻¹

F4= 0.0000 km⁻¹

NO. OF ITERATIONS= 1

RMS RESIDUAL= 1.28 mgal

BASE LEVEL= 28.2 km

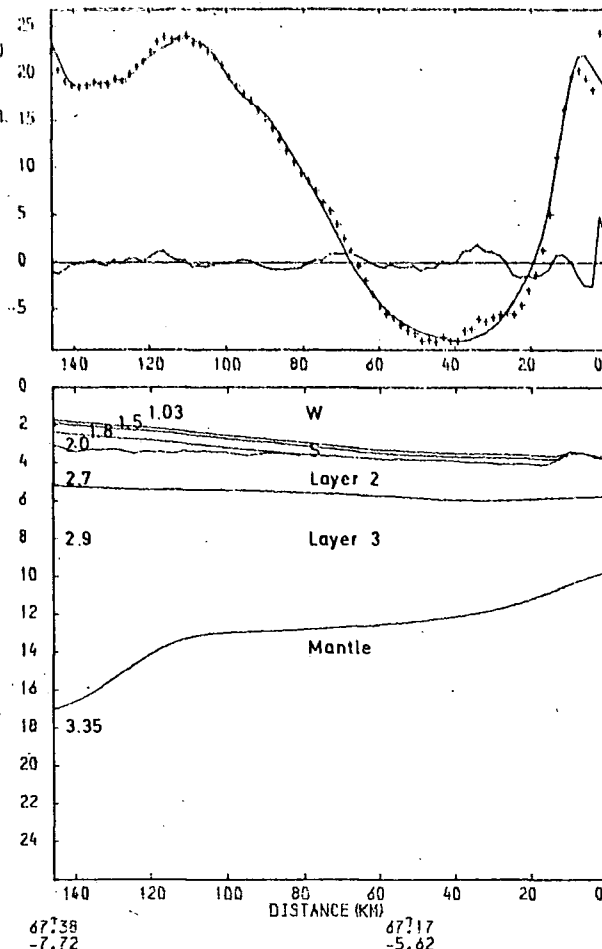
Figure 4.11

FREE AIR ANOMALY (MGAL)

OBSERVED AS CROSSES
CALCULATED AND RESIDUAL
AS SOLID LINES

DEPTH (KM)

LATITUDE
LONGITUDE



PROFILE 2

AVERAGE BEARING
OF PROFILE= 105.00°

SAMPLE SPACING= 1.869 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES:

$\Delta\rho_1 = -0.470 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_2 = -0.300 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_4 = -0.700 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_5 = -0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1
 MEAN DEPTH= 12.9 km
 S.D. OF DEPTH= 1.7 km
 $\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES
 $F3 = 0.0160 \text{ km}^{-1}$
 $F4 = 0.0320 \text{ km}^{-1}$

NO. OF ITERATIONS= 6
 RMS RESIDUAL= 1.03 mgal

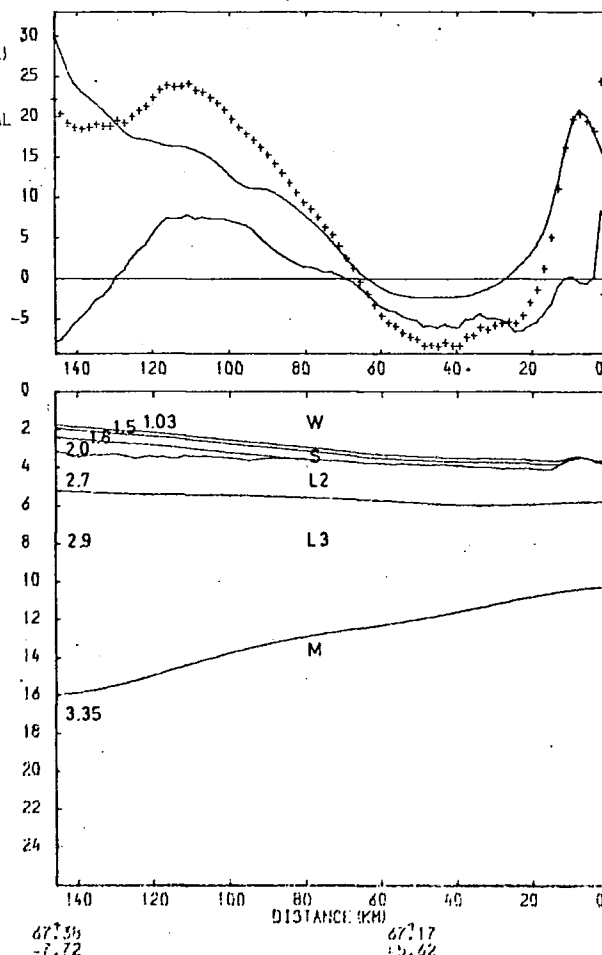
BASE LEVEL= 28.2 km

FREE AIR ANOMALY (MGAL)

OBSERVED AS CROSSES
CALCULATED AND RESIDUAL
AS SOLID LINES

DEPTH (KM)

LATITUDE
LONGITUDE



PROFILE 2

AVERAGE BEARING
OF PROFILE= 105.00°

SAMPLE SPACING= 1.869 km

DENSITY CONTRAST ACROSS
KNOWN INTERFACES:

$\Delta\rho_1 = -0.470 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_2 = -0.300 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_3 = -0.200 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_4 = -0.700 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_5 = -0.200 \times 10^3 \text{ kg/m}^3$

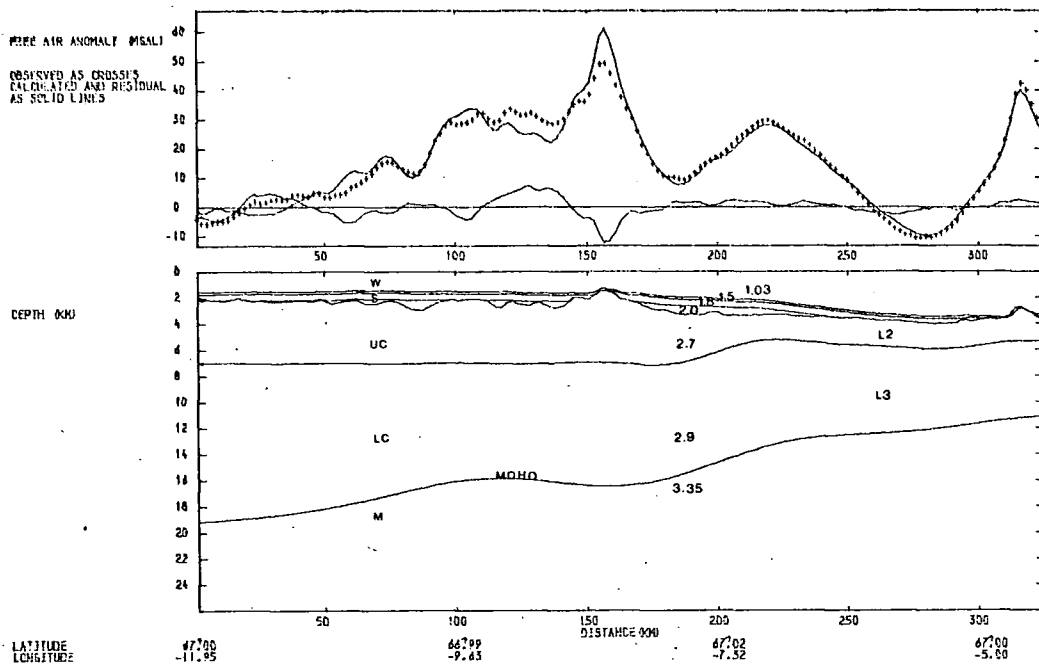
INVERTED LAYER 1
 MEAN DEPTH= 12.9 km
 S.D. OF DEPTH= 1.7 km
 $\Delta\rho = -0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES
 $F3 = 0.0080 \text{ km}^{-1}$
 $F4 = 0.0160 \text{ km}^{-1}$

NO. OF ITERATIONS= 4
 RMS RESIDUAL= 4.93 mgal

BASE LEVEL= 28.2 km

Figure 4.12



PROFILE 1

AVERAGE BEARING
OF PROFILE: 90.00°

DATA PROJECTED ONTO
LINE BEARING 115.00°

PROJECTED
SAMPLE SPACING: 1.030 km

DENSITY CONTRASTS ACROSS
KNOWN INTERFACES:

$\Delta\rho_{1-2} = 0.470 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{2-3} = 0.300 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{3-4} = 0.200 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{4-5} = 0.700 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{5-6} = 0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1

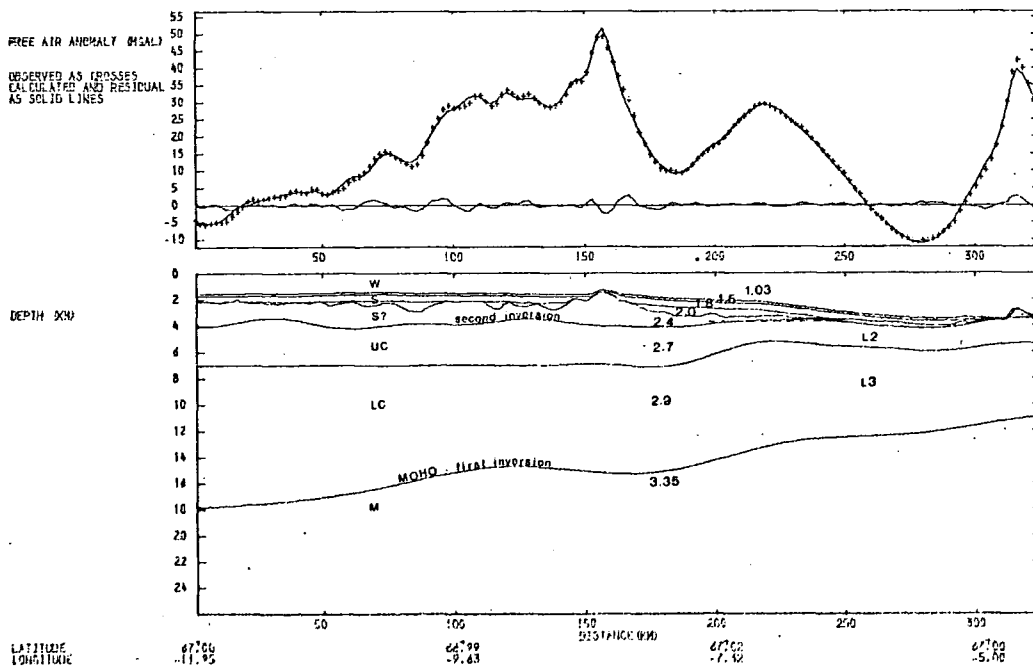
MEAN DEPTH: 15.2 km
 S.D. OF DEPTH: 2.4 km
 $\Delta\rho = 0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

$F_1 = 0.000 \text{ km}^{-1}$
 $F_2 = 0.0160 \text{ km}^{-1}$

NO. OF ITERATIONS: 5
 RMS RESIDUAL: 3.02 mgal

BASE LEVEL: 20.2 km



PROFILE 1

AVERAGE BEARING
OF PROFILE: 90.00°

DATA PROJECTED ONTO
LINE BEARING 115.00°

PROJECTED
SAMPLE SPACING: 1.030 km

DENSITY CONTRASTS ACROSS
KNOWN INTERFACES:

$\Delta\rho_{1-2} = 0.470 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{2-3} = 0.300 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{3-4} = 0.200 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{4-5} = 0.400 \times 10^3 \text{ kg/m}^3$
 $\Delta\rho_{5-6} = 0.200 \times 10^3 \text{ kg/m}^3$

INVERTED LAYER 1

MEAN DEPTH: 14.5 km
 S.D. OF DEPTH: 2.0 km
 $\Delta\rho = 0.450 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

$F_1 = 0.000 \text{ km}^{-1}$
 $F_2 = 0.0160 \text{ km}^{-1}$

NO. OF ITERATIONS: 5
 RMS RESIDUAL: 2.40 mgal

INVERTED LAYER 2

MEAN DEPTH: 4.6 km
 S.D. OF DEPTH: 0.7 km
 $\Delta\rho = 0.300 \times 10^3 \text{ kg/m}^3$

FILTER VALUES

$F_1 = 0.000 \text{ km}^{-1}$
 $F_2 = 0.0400 \text{ km}^{-1}$

NO. OF ITERATIONS: 5
 RMS RESIDUAL: 0.66 mgal

BASE LEVEL: 20.2 km

Figure 4.13

Further west, the picked reflector corresponds to Horizon A. Comparison with profile CP129 (Figure 1.9) suggests that about 2.5 s of sediments are present under the eastern flank of the ridge, thinning to the west as Horizon 0 rises nearer to the surface. As shown in Figure 3.1, indistinct reflectors are visible below Horizon A in the interval 40 to 90 km along the profile 8, and the deepest of these were tentatively correlated with Horizon 0. At its deepest point (about 65 km along track), Horizon 0 appears to lie about 2.8 s below the sea surface. As in CP129, Horizon 0 culminates about halfway across the flat part of the ridge crest. Further west the evidence is unclear, but judging from CP129, a small basin approaching 1 s in thickness is likely under the western part of the ridge. In view of the amount of sediment which has been neglected in the model of Figure 4.9a the large residuals are not surprising.

Rather than incorporating these inferences about the depth to Horizon 0 in an attempt to improve the model of Figure 4.9a it was decided to invert directly for the base of the sediment layer. The resulting model is shown in Figure 4.9b. In this model a zero density contrast was input across the interface corresponding to the deepest clear reflecting horizon. Consequently this interface did not contribute to the observed anomaly and was included for the purposes of comparison only. The average density of the sediment layer was chosen to be 2.0 Mg m^{-3} . This choice will be discussed below. The residual anomaly which resulted from the subtraction of the effect of the water layer and the assumed upper-lower crustal interface was inverted to give the relief on the base of the sediment layer. The base level for this inversion was chosen so that the inverted interface was nearly coincident with the acoustic basement at the eastern end of the profile. The depth to the model Moho was then determined such that the overall base level for the model was the same as for the Norway Basin profile.

The shape of the model sediment-basement interface in Figure 4.9b is consistent with the inferred topography of Horizon 0 under the Jan Mayen Ridge. At its thickest the sediment in the model is 5.0 km thick; this is consistent

with a reflection time of 2.8 s given a reasonable velocity gradient of 1.5 km s^{-2} . This sedimentary basin appears to be a continuation of the basin at the western end of the Norway Basin profile and is likewise associated with the anomaly L which achieves its maximum anomaly in the vicinity of profile 8 (Figure 4.1).

A problem associated with the model of Figure 4.9b is that the chosen sediment density of 2.0 Mg m^{-3} is almost certainly too low to accurately represent the average density of the deep sediment basin. Judging from Figure 4.5c, a value of at least 2.2 Mg m^{-3} would be more appropriate. However if a higher density is used the slopes of observed and model basements do not match at the eastern end of the profile and the depth to the base of the sediment basin becomes unrealistically large in the model. One solution to this dilemma is that the upper crustal density under profile 8 is higher than 2.7 Mg m^{-3} , say 2.9 Mg m^{-3} ; allowing the density contrast across the basement interface to remain at -0.7 Mg m^{-3} if the sediment density is increased to 2.2 Mg m^{-3} . Another solution, which is considered more likely, is that the Moho is not flat at the eastern end of the profile, but dips to the west, flattening out at about 60 km along the profile, thereby accounting for some of the westerly gravity gradient between anomaly C and anomaly L and allowing the remaining anomaly to be accounted for by a sediment layer of similar shape to that in Figure 4.9b, but of greater average density.

Profile 6 (Figure 4.10)

Profile 6 crossed the southern end of the Jan Mayen Ridge, the Jan Mayen trough and the northern end of ridge 5. As in the case of profile 8, substantial residuals arise if the base of the sediment layer is assumed to follow the deepest picked reflector (Figure 4.10a). Comparison with CP112 and CP108 suggests that Horizon 0 is shallow under the southern part of the Jan Mayen Ridge, hence there is not much error in the interpretation at the western end of the profile. Under Ridge 5 however, Horizon 0 probably dips to the east at about 1.5 s below the interface shown in Figure

4.10a.

Figure 4.10b shows the model obtained by inverting directly for the basement topography, assuming in this case a sediment density of 1.8 Mg m^{-3} . There is a good correspondence between calculated and observed basement topography outside of a central region between 50 and 100 km along the profile. The model basement topography under Ridge 5 is as expected from the Capan profiles, with a sediment depth of about 2 km. Of interest in the model is the sedimentary thickness of 2 km beneath the Jan Mayen trough, corresponding to the southern extension of anomaly L. This supports the contention (Gairaud et al., 1978) that the acoustic basement of the trough is a fairly thin basaltic layer overlying further sediments. At the western end of the trough, Horizon 0 appears to underlie the basalt at a depth of about 1.6 s below the sea bottom (CP108, Figure 1.9).

The sediment density of 1.8 Mg m^{-3} was chosen to give the best fit between observed and model basements at the eastern end of the profile. Although the sediments are relatively thin a higher density of perhaps 2.0 Mg m^{-3} would be more realistic. As in the case of profile 8 it is possible that the Moho dips to the west under the eastern end of the profile. Incorporation of a dipping Moho into the model would also allow a better fit between observed and model basements between 100 and 120 km along the track.

Profile 4 (Figure 4.11)

Profile 4 crosses ridges 4 and 5 of the southern ridge complex. As in the case of profiles 6 and 8 a model in which the base of the sediment is assumed to correspond to the deepest reflector observed on the single channel seismic records gives rise to unsatisfactorily large residuals. Figures 4.11a and 4.11b illustrate the models in which the basement topography was obtained by direct inversion with input sediment densities of 2.0 and 1.5 Mg m^{-3} respectively. Both models show a good correspondence between observed and computed basement relief east of about 45 km along the profile. Beneath ridge 4 a sedimentary thickness of about

1 km is indicated, in reasonable agreement with observations on profiles CP108 and CP103.

In both models the computed basement lies at a depth of about 3 km below sea level under the southern extension of the Jan Mayen trough to the west of ridge 5 and is about 1 km deeper than the observed basaltic acoustic basement in this region. It is difficult to account for the magnitude of the gravity anomaly over the ridge if no sediment lies beneath the acoustic basement to the west unless the basement density is at least 3.0 Mg m^{-3} . It is perhaps more likely that the basalt is relatively thin and overlies more sediment, as is believed to be the case under the Jan Mayen trough further north. The occurrence of holes in the opaque layer, at about 110 km along profile 4, and at the southeastern end of profile 5 (Figure 3.1) lends support to this idea. Alternatively a deeper density deficiency may be biasing the gravity anomaly to the west of ridge 5.

Profile 2 (Figure 4.12)

Profile 2 forms an eastern extension of profile 4, offset 20 km to the south. The profile is underlain by oceanic basement, the depth to which was reliably determined from the single channel seismic record. The models in Figure 4.12 were obtained by inverting for the basement topography, as described for the Norway Basin profile. Relatively short wavelengths (30 km) need to be admitted into the Moho topography in order to obtain a good fit (compare Figures 4.12a and 4.12b). Anomaly C appears to be largely accounted for by variation in the depth to the Moho.

Profile 1 (Figure 4.13)

The depth to the base of the sediment layer was well determined along profile 1, which crosses the southern ridge complex at 67°N although as indicated in Figure 3.2, there are diffraction patterns originating from beneath the acoustic basement under the eastern part of the profile and indications of further sediment underlying the basalt which forms the basement of ridge 1.

In the model of Figure 4.13a, the depth to the Moho has been determined as for the Norway Basin profile. There are substantial residuals which correlate with the basement topography. Given that the sediment densities used are fairly realistic it is apparent that the density of the material beneath the sediment must be only about 2.4 Mg m^{-3} so that another interface is required in the model.

At the same time there are large variations in the depth to the base of the model Moho, which appear quite "genuine" and do not correlate overtly with the residual anomalies (compare Figure 4.9a). Therefore the approach used for the more northern lines, in which a flat Moho was assumed, is not appropriate.

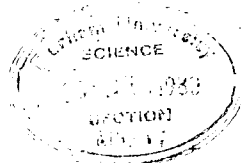
The model in Figure 4.13b was obtained by a two stage inversion procedure. Firstly the Moho was determined as in Figure 4.13a, except that a density contrast of -0.4 Mg m^{-3} instead of -0.7 Mg m^{-3} was input across the base of the sediment layer. Secondly the residual so obtained was inverted to find the topography on the base of a layer of density 2.4 Mg m^{-3} which was constrained to the same depth as the oceanic crust over the eastern part of the profile. This procedure was repeated a few times, changing the base level for the first inversion, until an overall base level of 28.2 km was obtained. The model is quite simple in that very little variation on the new interface is necessary to substantially decrease the rms residual to 0.86 mgal.

In general, using a finite number of gravity observations and assuming a knowledge of the density contrast, it is possible to determine the subsurface of one interface only (Bott, 1973). Using frequency domain methods one can invert for two interfaces provided that it is assumed that they contribute to disjoint parts of the gravity anomaly spectrum. This assumption is rarely justified, for while it is possible to show that a deep interface, such as the Moho, cannot contribute substantially to the short wavelength parts of the spectrum, it is possible for an undulating shallow interface to contribute to the long wavelength part. In the model of Figure 4.13b all of the long wavelength variations

in the gravity field are taken up by the model Moho, so that the second inverted interface has only relatively short wavelength variations superimposed upon an independently estimated mean level. Clearly an equivalent fit would be obtained if compensating undulations were introduced into the Moho and the upper interface. However, as the upper interface cannot rise by more than 1-2 km the extreme depth of the model Moho is limited. In the models for profiles 4, 6 and 8, the opposite assumption was made concerning the long wavelengths, in that they were all assigned to the upper interface, and similar comments obviously apply.

DSDP hole 350 showed the basement of ridge 1 to be basalt of age 41 Ma (Section 1.8). According to Gairaud et al. this basalt overlies further sediment, then Horizon 0. On the basis of diffraction patterns observed on the monitor records it is suggested that the faulted acoustic basement west of Ridge 1 is also flow basalt underlain by sediment. The inferred layer of density 2.4 Mg m^{-3} is therefore identified with a layer of Tertiary sediment of post opening age. One would imagine that the true basement underlying this layer is affected by the same faults which affect the acoustic basement. The exact topography of the base of this layer is not well determined from the gravity data; it could vary rather more than is shown in Figure 4.13b.

The first 40 km of profile 1 overlies what is probably oceanic crust of the Iceland Plateau. One might expect the model Moho to rise to a depth of about 11 km under this region, but instead it deepens to about 18 km. Slab formula computations based on R.V.Vema data (V2703, 27 July 1969, in Talwani 1974) show that a similar apparent crustal thickness is maintained right across the Iceland Plateau to the Greenland shelf. Possible reasons for this anomalously deep model Moho will be discussed in the concluding section.



4.5 Discussion and conclusions

Upper crustal structure

Gravity interpretation of the Norway Basin profile showed that the majority of the prominent gravity anomalies along the profile can be satisfactorily accounted for by observed variations in water depth and sediment thickness, leaving a long wavelength residual anomaly thought to be due to deeper structures.

The interpreted profiles over the Jan Mayen block cross most of the major short wavelength anomalies in the region (Figure 4.1) and the available evidence is consistent with an overall model in which these anomalies are caused by the water layer and post Horizon 0 sediments, contrasting with a basement of density 2.7 Mg m^{-3} , or possibly higher. This conclusion was based on a rather roundabout interpretational procedure because of the inadequacy of the seismic monitor records. The use of fully processed data would allow stronger conclusions about upper crustal densities and the variation of deeper interfaces, particularly if closely spaced velocity analyses were performed.

The gravity models F, J and N of Gronlie et al. (1979) incorporate a substantial block of low density material underlying the southern Jan Mayen Ridge and ridges 4 to 6 of the southern ridge complex. This material is interpreted by them as pre-Tertiary sediment. The interpretation for profile F is not unlike that for profile 1, (Figure 4.13b) although in the model of Gronlie et al. (1979) the low density layer does not underlie ridges 1 to 3, in accordance with their contention that these ridges are oceanic in nature. Profile F intersects profile 6, yet the respective models are quite different. In the case of profile 6 no necessity was seen for incorporating such a substantial low density layer and indeed it would seem as if an assumed density of 2.4 Mg m^{-3} below the acoustic basement would actually preclude a good fit, as the observed anomaly amplitudes would not match. As the Airy isostatic Moho assumed by Gronlie et al. is not included in their figure,

it is not clear how this influences the calculated anomaly. The interpretations presented in the present study do not provide any evidence of pre-Tertiary sediments beneath the Jan Mayen block. However the existence of dense pre-Tertiary sediments is not precluded.

Deep structure

Under the Norway Basin the model Moho deepens gradually away from the centre of the basin, then plunges under the Jan Mayen and Norwegian rises, appearing to level out at a depth of about 16 to 18 km.

Along the short profiles crossing the Jan Mayen block the Moho was assumed to be flat and all that could be determined was its level. In the case of profiles 8 and 6 the depths of 13.5 and 14.5 km respectively are probably biased by the choice of low densities for the sediment layer. The use of more realistic densities would increase the model Moho depth to about 16 km but would impair the anomaly fit unless the model Moho were allowed to rise at the eastern ends of the profiles.

Along profile 1 the model Moho has a minimum depth of 11 km under the Norway Basin, deepens to about 16 km under the Jan Mayen block then deepens further to 18 km under the oceanic Iceland Plateau.

It is difficult to assess how closely the model Moho topography described above accords with reality. The interpretations were made under the assumption of lateral density uniformity within the crust and mantle. In general terms a deep model Moho is indicative of a relative mass deficiency below the upper crust and it is by assumption that this deficiency is accounted for by crustal thickening.

One alternative possibility is that the crust is less dense in those regions where the model Moho is deep. In the present models the crust has an average density of about 2.84 Mg m^{-3} and is 11 to 14 km thick under these regions (excluding the sediment layer). In order for the true Moho

to be 5 km shallower at about 11 km depth the average crustal density would have to be about 2.5 Mg m^{-3} , assuming lateral uniformity within the mantle. Such a low density is not considered likely and it is probable that either the Moho depths in the models are fairly realistic or that there are compensating mass deficiencies occurring below the base of the crust.

The relatively negative residual anomalies which give rise to deep model Mohos arise in general from shallow water depths rather than from relatively negative free air anomalies. The problem of determining whether these negative residuals are due to thick crust or to mass deficiencies within the mantle is closely linked to the problem of deciding whether the anomalously elevated crust of the southern Norwegian-Greenland Sea is isostatically compensated for at a shallow or deep level.

It is possible to discriminate between shallow and deep compensation on the basis of extensive gravity data. An area of elevated crust compensated for at shallow depth produces only a slight gravity anomaly at the surface, whereas deep compensation gives rise to a long wavelength negative anomaly because the gravity effect of the mass deficiency is of diminished amplitude and is spread over a wider area.

Cochran and Talwani (1978) produced a satisfactory model of the gravity field around Iceland in which about 500 m of the excess elevation over the whole area is compensated for by a reduced density in the lithosphere and the asthenosphere, extending to a few hundred kilometres depths and the remainder is compensated for locally by mass deficiencies in the upper lithosphere, most probably by crustal thickening. A feature of this model is that the excess elevation ascribed to the mantle is believed to be fairly constant temporally and spatially, over a wide area, possibly extending southward to the Azores. Cochran and Talwani suggested that the deep density deficiency is about 0.006 Mg m^{-3} , which could be produced by a rise in temperature of about 60°C . An increased mantle temperature could induce deeper melting, and would therefore produce a

thicker crust. This is an attractive model in that abrupt changes in crustal thickness are easier to explain than abrupt changes in the mean density of the upper mantle.

This model is adopted in the present study. The postulated regional deficit in mantle density can be assumed to produce the same effect over all the profiles and to be taken account of by the chosen base level. The assumption of a normal crustal thickness in the centre of the Norway Basin is consistent with the residual depth anomaly of 500 m over the Aegir axis.

However the gravity interpretations presented above are not adequate in the respect that account was not taken of the normal variations of upper mantle density with age. In Cochran and Talwani's model the variations in residual depth anomaly are due to shallow sources and the lithosphere is assumed to be normally developed with age. This is a reasonable assumption because the physical parameters which determine the rate of heat flow from the mantle and the amount of densification that takes place are known or inferred to be relatively invariable. Furthermore any limitation of the thickness of the lithospheric boundary layer (by convection at shallow depths for instance) would necessarily increase the heat flow through the layer. Although Langseth and Zielinski (1974) averred that heat flows are anomalously high over the whole Norwegian-Greenland Sea, they do not differ greatly from the revised North Pacific averages given by Parsons and Sclater (1977), who found that heat flow q , varies with age in the manner

$$\begin{aligned} q(t) &= 11.3t^{0.5} \quad (\mu\text{cal cm}^{-2}\text{s}^{-1}) \\ &= 0.47t^{0.5} \quad (\text{W m}^{-2}) \end{aligned}$$

In fact the mean percentage difference between the 26 reported measurements in the southern Norwegian-Greenland Sea (Gronlie and Talwani, 1978) and the values derived from this expression (with anomaly identifications as described in Chapter 6) is $-10\% \pm 27\%$ (that is the measured values are low). This result is biased by two sets of 5 closely grouped measurements north of the Denmark Strait which are all low. If these are averaged first the mean percentage difference is $-2\% \pm 26\%$. The close correspondence of measured heat flow in

* $\pm$ one standard deviation

the southern Norwegian-Greenland Sea to the values in the North Pacific provides a strong indication that the development of the lithosphere beneath the region is normal.

Account can be taken of the variations in upper mantle density with age without explicitly involving models with lateral density variations. Consider the Norway Basin profile. At the centre of the basin the lithosphere is about 30 Ma old, at the margins about 50 Ma old. Between 30 Ma and 50 Ma normal oceanic crust subsides by 560 m. This subsidence produces a relative gravity anomaly of -54 mgal on account of the replacement of 560 m of mantle material by 560 m of sea water. Over normal 30 - 50 Ma crust this anomaly is nearly balanced by the gravity effect of lithospheric changes. Cochran and Talwani have determined an empirical free-air gravity age relationship for normal oceanic crust. Over 50 Ma crust the average anomaly is zero, over 30 Ma crust, 3 mgal. Hence the lithospheric contribution over 50 Ma relative to 30 Ma crust is 51 mgal. The effect of taking this contribution into account would be to increase the depth of the model Moho by 2.7 km at the margins of the Norway Basin, with of course, a smooth increase from the axis. If Cochran and Talwani's model is correct the oceanic crust must thicken by nearly 6 km towards the margins.

It is also reasonable to add about 3 km onto the model Moho depths for the Jan Mayen block and Norwegian margin as the lithosphere underlying these regions must be at least as old as 50 Ma. If this is done the relative Moho topography shown in the models for the Norway Basin margins is preserved. The material beneath the Jan Mayen block and the Norwegian margin can be tentatively interpreted as thinned continental crust.

In the case of the Iceland Plateau the oldest oceanic lithosphere adjacent to the Jan Mayen block at the western end of profile 1 is of similar age to that in the centre of the Norway Basin. Hence the model Moho depth of 18 km does not need alteration. It is consistent with the crustal depth of about 20 km found by Evans and Sacks (1979). Similar

apparent depths can be calculated across the whole Iceland Plateau, but taking into account the changes in the lithosphere the crust must thin towards the Kolbeinsey Ridge.

Accepting Cochran and Talwani's shallow compensation model it appears that the oceanic crust is anomalously thick near the margins of both the Norway Basin and the Iceland Plateau. Some of the thickening may be related to the "Iceland phenomenon" but apart from this one would expect thick oceanic crust in any sufficiently young ocean. At the inception of continental rifting, any magma generated must have sufficient hydrostatic head to reach the surface because basaltic magma is less dense than continental crust. Thus one would expect the initially accreted crust to be as thick as the continental crust, and it is difficult to see how the change to normal oceanic thickness could be anything but gradual.

Chapter 5

INTERPRETATION OF MAGNETIC ANOMALY DATA ALONG THE
NORWAY BASIN PROFILE5.1 Frequency domain magnetic anomaly interpretation

As stated in Section 4.1, Parker (1972) showed how a series of Fourier transforms can be used to calculate the gravity or magnetic anomaly due to a non-uniform layer of material. As an example he gave the formula for the total field magnetic anomaly due to a magnetised layer with an arbitrary two-dimensional topography but constant vertical thickness, in which the magnetisation vector has a varying intensity but constant direction. Parker and Huestis (1974) showed how this formula could be rearranged to yield a procedure for the inversion of magnetic anomalies in the presence of topography.

This procedure can be applied to the interpretation of sea floor spreading magnetic anomalies if the following assumptions are made

- (i) the anomalies, the oceanic basement topography and the magnetisation distribution within the oceanic crust are all two-dimensional
- (ii) significantly magnetised rocks are confined to a layer of constant thickness whose upper surface is coincident with that of the oceanic basement
- (iii) within the magnetised layer the direction of magnetisation is constant (except for possible 180° reversals)
- (iv) the intensity of magnetisation within the layer varies horizontally but not vertically.

Under these assumptions the horizontal variation of magnetic intensity within the oceanic crust can be inferred from a knowledge of the variation in the total field anomaly and depth to basement along a profile which is perpendicular to strike, provided that the thickness of the magnetised layer and the direction of magnetisation are independently estimated.

Description of theory and implementation

Consider an x-z Cartesian coordinate system (Figure 5.1) such that the z-axis is vertically downwards and the x-axis has a bearing β relative to true north. The total field anomaly $A(x)$ due to a two-dimensional magnetised layer which strikes perpendicularly to the x-axis is measured at the surface, $z=0$. The magnetised layer is bounded above by the surface $z(x)$ and has a constant vertical thickness, h_0 . The magnetisation vector within the layer has a constant declination, D_m and inclination, I_m but a variable intensity, $M(x)$ which can be either positive or negative. The earth's field is assumed to have a constant declination, D_0 and inclination, I_0 over the whole profile. Subject to the restriction that $A(x)$ is small relative to the magnitude of the earth's field, the Fourier transform (see Section 4.1) of the total field anomaly is given by

$$F[A(x)] = (\mu_0/2) \exp(-|k|z_0) (1 - \exp(-|k|h_0)) V(k) \cdot \sum_{n=0}^{\infty} (|k|^n/n!) F[M(x) (z_0 - z(x))^n] \quad 5.1$$

where z_0 is an arbitrary level and

$$V(k) = (\sin I_0 - i(k/|k|) \cos I_0 \cos(\beta - D_0)) \cdot (\sin I_m - i(k/|k|) \cos I_m \cos(\beta - D_m))$$

Equation 5.1 can be rearranged to give

$$M(x) = F^{-1} \left\{ \frac{F[A(x)] \exp(|k|z_0)}{(\mu_0/2) (1 - \exp(-|k|h_0)) V(k)} \right\} - F^{-1} \left\{ \sum_{n=1}^{\infty} (|k|^n/n!) F[M(x) (z_0 - z(x))^n] \right\} \quad 5.2$$

This equation forms the basis of an inversion procedure in which a new, hopefully better, estimate of magnetisation can be generated from an old estimate. The following points need to be considered in a practical implementation.

(i) As in the case of the gravity inversion procedure this scheme is subject to short wavelength instability. Thus in order to ensure convergence it is necessary to filter each estimate to remove undesirably high spatial frequencies.

(ii) In addition, the scheme is subject to long wavelength instability arising from the denominator term $(1 - \exp(-|k|h_0))$. As pointed out by Bott and Hutton (1970a), "errors in observation of quite small amplitude but long wavelength such as may be caused by a lack of correction for diurnal variation, produce large amplitude fluctuations of the same

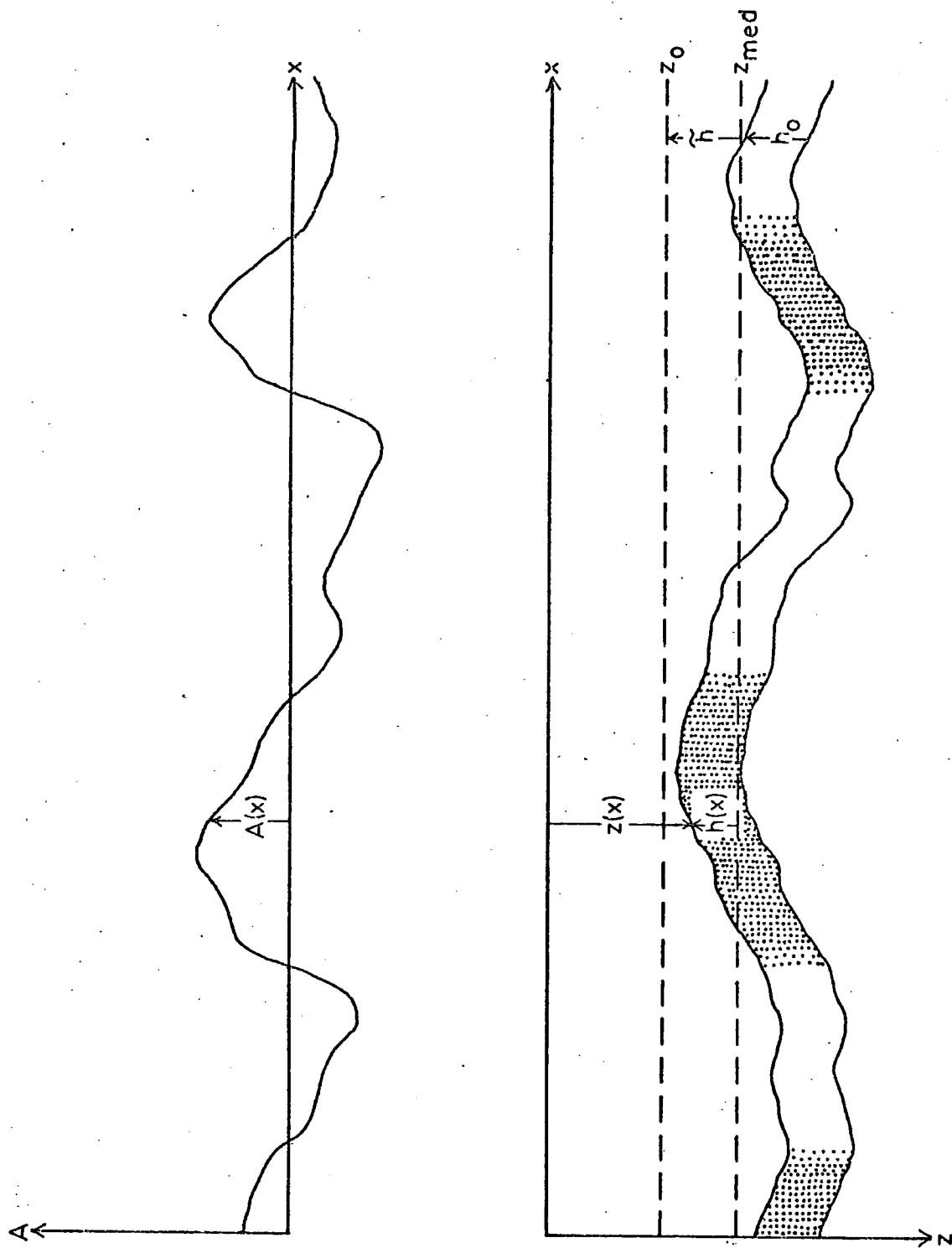


Figure 5.1. Set up for equations 5.1 to 5.3.

wavelength in the resulting magnetisation distribution". As there is no leakage of energy into the long wavelength part of the magnetisation spectrum during iteration, this instability can be avoided by removing the mean value and long wavelengths from the anomaly prior to iteration.

(iii) The inverse procedure converges most rapidly if the reference level z_0 is situated just above the highest topographic point, while the infinite sum converges most rapidly if the reference level is midway between the highest and lowest points in the surface $z(x)$. This problem is overcome by an appropriate adjustment of levels.

A practical inversion scheme is

$$M_{i+1} = M_0(x) - F^{-1} \left\{ L(k) \left(\exp(-|k|\tilde{h}) \sum_{n=1}^{\infty} (|k|\tilde{h})^n / n! \right) F[M_i(x) h^n(x)] - F[M_i(x)] \right\} \quad 5.3$$

with a starting estimate $M_1(x) = 0$, where

$$F[M_0(x)] = \begin{cases} 0 & \text{for } k=0 \\ \frac{L(k)H(k)F[A(x)]\exp(-|k|z_0)}{(\mu_0/2)(1-\exp(-|k|\tilde{h}))V(R)} & \text{for } |k|>0 \end{cases}$$

where $L(k)$ is the low pass filter of Equation 4.3 and $H(k)$ is a similarly designed cosine-tapered high pass window, which stops wavelengths less than $1/f_1$, attenuates wavelengths between $1/f_1$ and $1/f_2$ and lets pass wavelengths greater than $1/f_2$ (see Figure 5.4). The topographic variation $h(x)$ is given by

$$h(x) = (\max(z(x)) + \min(z(x))) / 2 - z(x)$$

and the quantities $\tilde{h}$ and z_0 by

$$\tilde{h} = (\max(z(x)) - \min(z(x))) / 1.98$$

$$z_0 = (\max(z(x)) + \min(z(x))) / 2 - h$$

Parker and Huestis (1974) pointed out that the problem of determining the magnetisation distribution $M(x)$ does not have a unique solution. In the case of an infinite layer with topographic variation, there exists a magnetisation distribution called the magnetic annihilator, which gives rise to no magnetic anomaly above the layer. Any multiple of the annihilator can be added to a determined distribution $M(x)$ without affecting the external magnetic field. In the case of a layer of finite extent, no exact annihilator exists but there is a distribution which gives rise to a minimal

anomaly above the layer. A band limited form of the annihilator, valid for all ambient field and magnetisation directions may be found from Equation 5.3 with the substitution $M_0(x)=1$ (the factor $\exp(-|k|h)$ can be ignored in this case). In contrast to $M(x)$, which is constructed with a mean level of zero, the annihilator distribution has a mean level of unity. A multiple of the annihilator distribution can be added to $M(x)$ to give a distribution which has a desired mean value. However, for profiles crossing approximately equal amounts of normally and reversely magnetised crust a mean magnetisation of about zero would be expected.

The forward and inverse algorithms were implemented in the FORTRAN program MAGINV (Appendix B). The function of the program is to compute firstly the magnetisation distribution $M(x)$ and then, using the forward algorithm, the calculated anomaly due to $M(x)$. MAGINV is fully commented and many of the details of implementation, such as the use of the Fast Fourier Transform are similar to those of GRVINV. The convergence criterion for computation of the infinite sum in Equation 5.3 is similar to that described for GRVINV and the overall convergence of the inverse procedure is assumed when the rms difference between successive estimates of magnetisation is less than 0.005 A m^{-1} .

An example of the graphical output of MAGINV is shown in Figure 5.2. The system of data preparation has been described in Section 4.2. Instructions for the use of the program are given in Appendix B.

5.2 Magnetic anomaly inversions over the Norway Basin profile

A number of inversions of the total field magnetic anomaly data which was acquired along the Norway Basin profile are shown in Figures 5.2, 5.3 and 5.4. The anomaly identifications which are shown are based on the results of Chapter 6. Note that anomalies 24A and 24B are distinct. At the western end of the profile, anomaly 21 is believed to be duplicated by offset across a fracture zone. Figure 5.3b illustrates the inversion of an artificial noise data set.

Figure 5.2. Magnetic anomaly inversions over the Norway Basin profile, showing the calculated magnetisation distribution which gives rise to a computed anomaly (++++ line) which very nearly matches the observed anomaly (—), on the assumption that the magnetic source layer is of constant thickness and lies immediately below the acoustic basement. Also shown is the annihilator distribution which gives rise to no external magnetic field. Figure 5.2a shows anomaly identifications (see Chapter 6). In the model of Figure 5.2a, the layer thickness is 0.5 km, in that of Figure 5.2b, 3.0 km. All other parameters are the same and are given in detail in the computer drawn captions. Vertically hatched sections of the assumed source layer correspond to positive computed magnetisation.

Figure 5.3a shows an unrealistic inversion in which the long wavelength noise was not removed from the anomaly profile and has become excessively amplified in the magnetisation distribution.

Figure 5.3b shows an inversion of a white noise signal, assuming the basement topography of the Norway Basin profile.

Basement topography

In the models of Figures 5.2 and 5.3 the basement topography is as determined from the seismic monitor records, with conversion from time to depth according to Equations 3.3 and 3.4. The basement topography in Figure 5.4, which was computed using a constant seismic velocity of 2.0 km s^{-1} for the sediment layer, is not significantly different for the purposes of magnetic inversion.

Choice of magnetisation direction

Since the beginning of Tertiary times both Greenland and Norway have moved northwards towards the pole by about 15° . At the time that spreading began between Greenland and Norway, the mean latitude of the incipient Norway Basin was about 52°N (Smith and Briden, 1977; Harland, 1979). By 40 Ma before present, when most of the spreading in the basin had taken place, its mean latitude had increased to about 56°N . Assuming an axial dipole magnetic field, the magnetic inclinations corresponding to these values are 68° and 71° respectively. In the models of Figures 5.2 and 5.3 an average magnetisation inclination of 70° was assumed. It should be noted that measured inclinations in the East Greenland basalt pile, which was extruded shortly before spreading began, are somewhat lower than the value inferred for the time of opening, as they range from 60° to 65° (Faller, 1975; Faller and Soper, 1979). The reason for this discrepancy is not clear.

Figure 5.4 illustrates the effect of varying the inclination of the magnetisation vector, as projected in the plane of the profile. These inversions will be discussed below.

The azimuth of the Norwegian margin does not appear to have changed significantly during the Tertiary (Smith and Briden 1977). Once again, assuming the axial dipole hypothesis is correct, the time averaged declination of magnetisation for the eastern part of the Norway Basin should be about 0° . However the western part of the Norway Basin,

outside of the central zone, has undergone an anticlockwise rotation of about 20° (Chapter 7). Hence an average declination of about -10° can be assumed for the whole Norway Basin.

Choice of filter values

As discussed in Section 5.1, the measured magnetic anomaly is filtered prior to the inversion in order to remove any long wavelength noise components such as may arise from temporal magnetic variations and inadequacies of the assumed regional field. The chosen high pass filter parameters were $f_1=0.08 \text{ km}^{-1}$ and $f_2=0.015 \text{ km}^{-1}$. Accordingly, in the inversions of Figures 5.2, 5.3b and 5.4 wavelengths longer than 125 km are completely eliminated from the magnetisation distribution, wavelengths between 125 km and 67 km are attenuated and wavelengths shorter than 67 km are unaffected.

The principle component of the quiet day magnetic temporal variation has a period of 24 hours (Matsushita 1967); of diminishing importance are components with periods of 12, 8 and 6 hours which in northern latitudes have proportional amplitudes of about 0.6, 0.25 and 0.08 respectively, relative to the diurnal variation. The projected speed along the Norway Basin profile is 12.3 km hr^{-1} . Thus these temporal components appear as anomalies with wavelengths of 300, 150, 100 and 75 km. The chosen filter eliminates the diurnal and semidiurnal components and attenuates the $1/3$ and $1/4$ diurnal components. The latter components are less important than the diurnal component, firstly because their amplitude is normally smaller, and secondly because the factor $[1-\exp(-|k|h_0)]^{-1}$ of Equation 5.3 is also relatively small. Over the range of interest this factor is proportional to wavelength for reasonable layer thicknesses.

Figure 5.3a illustrates the effect of neglecting to apply a high pass filter. Long wavelength undulations in the observed anomaly are excessively amplified in the magnetisation distribution. The overall trend, from positive in the west to negative in the east, is probably due to an

inadequacy of the regional field, as it is observed along profiles from other sources (Figure 6.5). Superimposed on this trend are the effects of quiet day variations and also more severe long wavelength disturbances such as Event 5 (Section 2.4). Comparison with the model of Figure 5.2a, for which all parameters other than f_1 and f_2 are the same, shows that the highest pass filter effectively removes these undesirable effects and produces a magnetisation direction in which the normal and reversed magnetisations, corresponding to positive and negative anomalies, are of approximately equal magnitude. Comparison of the filtered anomaly, which is shown in Figure 5.2a, and the original in Figure 5.3a suggests that very little has been lost in the way of genuine signal. However, if a more severe low cut filter were to be applied it is certain that the anomalies themselves would be affected, as they have important components with wavelengths of about 40 km.

The low pass filters f_3 and f_4 were chosen according to the criterion of Bott and Hutton (1970a) who showed that serious short wavelength instability arises if a magnetic field is downward continued to a depth greater than about $3/4$ of the shortest wavelength that is present after filtering. The mean depth to magnetic basement under the central Norway Basin is about 4 km, therefore wavelengths shorter than about 5.3 km should be filtered out. The chosen values, $f_3 = 0.15 \text{ km}^{-1}$ and $f_4 = 0.20 \text{ km}^{-1}$ correspond to wavelengths of 6.7 km and 5.0 km respectively.

Figure 5.3b shows the effect of inverting a noise series which has an rms amplitude of about 1 nT, with filter values f_1 to f_4 as described above. The input "anomaly" was a series of random numbers (from a normal distribution of mean 0 and standard deviation 1) which were rounded down to the nearest nanotesla. The resulting magnetisation distribution has an rms value of about 0.05 A m^{-1} which is quite acceptable. If all frequencies from dc to the Nyquist frequency are permitted in the magnetisation distribution, its rms amplitude increases to about 0.3 A m^{-1} .

The effect of varying the thickness of the magnetised layer

As described in the previous section, the magnetic inversions are performed under the assumption that the magnetisation of the oceanic crust does not vary vertically within a layer of constant thickness, outside of which it is negligible. It is reasonable to assume that the oceanic magnetised layer is of constant thickness as it is likely that much of the relief of the oceanic basement is caused by block faulting which occurs after most of the magnetised layer has formed (Klitgord and Mudie, 1974). If it is alternatively assumed that the magnetised layer has a flat base (for example Talwani et al., 1971) then the inferred magnetisation distribution is not appreciably altered (for reasonable topographic variations) providing that the layer thickness is at least 500 m (Klitgord et al., 1975).

The assumption that the magnetisation within the layer does not vary vertically is unlikely to be correct (Klitgord et al. 1975). However the vertical variation of magnetisation and the overall thickness of the magnetised layer are at present not well known (Johnson, 1979).

In a number of early studies in which direct inversions of magnetic anomalies were performed (Bott, 1967; Emilia and Bodvarsson, 1969; Bott and Hutton, 1970b) the magnetised layer was assumed to be about 2 km thick. Following the work of Talwani et al. (1971) and Atwater and Mudie (1973) a layer thickness of 0.5 km has been commonly used in inversions. These authors estimated the magnetisation of the uppermost crust by modelling short wavelength anomalies which could be demonstrated to arise from topographic variations. Assuming the inferred magnetisations to be representative of the magnetised layer as a whole they found that the layer need only be 500 m thick, in order to model the long wavelength sea floor spreading anomalies. The inferred magnetisations are consistent with those measured for surface dredge samples (Harrison, 1976; Lowrie, 1977).

However, deep drilling of the ocean crust (to a depth of several hundred metres) has generally shown that the average

remanent magnetisation within the upper crust is insufficient to account for the magnitude of observed anomalies unless the magnetised layer extends considerably deeper than 0.5 km (see Steiner et al. (1979) and Johnson (1979) for lists of earlier references). A number of authors (for example Harrison (1976), Hall (1976)) now consider it likely that significant magnetisations occur not only throughout oceanic layer 2, but also perhaps in the upper part of layer 3.

Shown in Figures 5.2a and b are two inversions, one assuming a layer thickness of 0.5 km and the other of 3 km. If the resulting magnetisation distributions are compared directly (as plotted in Figure 5.2) they are found to be very similar. From the scales of the plots in Figures 5.2a and 5.2b the inferred magnetisations in the 0.5 km layer can be calculated to be uniformly about 4.4 times those in the 3 km layer. The significance of the actual values of inferred magnetisation will be discussed below. The similarity in shape between the two magnetisation distributions is somewhat surprising. Ignoring topographic effects it can be seen from Equation 5.2 that the ratio of magnetisation amplitudes in the models should be dependent on wavenumber k , and be given by

$$\{1 - \exp(-3.0|k|)\} / \{1 - \exp(-0.5|k|)\}$$

This ratio is equal to 4.4 for $|k| = 2\pi / (22 \text{ km})$. Thus it seems as if the perceived shapes of the magnetisation distributions are dominated by wavelengths of about 22 km, which seems reasonable on inspection. For shorter wavelengths there are appreciable differences between the distributions. For the shortest wavelength present (5 km) the ratio given above is only 2.1, therefore the short wavelength components assume a greater relative importance in the thick layer inversion. However the short wavelength differences between these distributions are small and it is impossible to decide a priori whether the shape of the thick layer distribution is more realistic than the thin layer distribution, or vice versa. Consequently, as found by Parker and Huestis (1974) the sea floor spreading anomalies contain no information about the thickness of the magnetised layer unless one is willing to assign a bound to the intensity of magnetisation (Huestis and Parker, 1977).

Conversely it does not seem to be necessary to know the vertical variation of magnetisation in order to obtain reasonable information about its relative horizontal variation. The considerations outlined above cast doubts upon the spectral analysis presented by Harrison (1976), which purportedly shows that the oceanic magnetised layer is about 3 km thick. The apparently unique depth to the centre of the magnetised layer which is given by Harrison's technique is probably dependent on the rigid assumptions which are built into the model upon which the technique is based.

If the magnetised layer is assumed to be infinitely thick, the term $[1 - \exp(-|k|h_0)]^{-1}$ will be equal to 1 for all wavenumbers. The inferred magnetisation distribution will remain finite. In the present case, where $F1 = 0.008 \text{ km}^{-1}$, any layer thicker than 100 km will be effectively infinite, in that the term involving h_0 will deviate by less than 1% from unity. An inversion was performed for a layer thickness of 100 km (not shown) and the magnetisation distribution was found to be similar in shape to that found for a 0.5 km layer although more noisy. The magnetisations were found to be about 7.5 times less than those for the thin layer. Once again this is the ratio appropriate to wavelengths of about 22 km. Although a 100 km thick layer is unrealistic, the results of such an inversion give lower bounds on magnetisation. The significance of the magnetisation distributions derived for the Norway Basin will be further discussed below.

Effect of varying the inclination of magnetisation

Figure 5.4 shows the effect of varying the assumed inclination of magnetisation. For the sake of simplicity the magnetisation in these models was assumed to have no component parallel to the strike of the anomalies, that is the declination of magnetisation was set equal to the bearing of the projected profile, β (in this case 140°). If the declination of magnetisation is in fact D_m , then the model magnetisation, M and inclination, I_m are related to those projected into the plane of the profile (M' and I'_m) as in

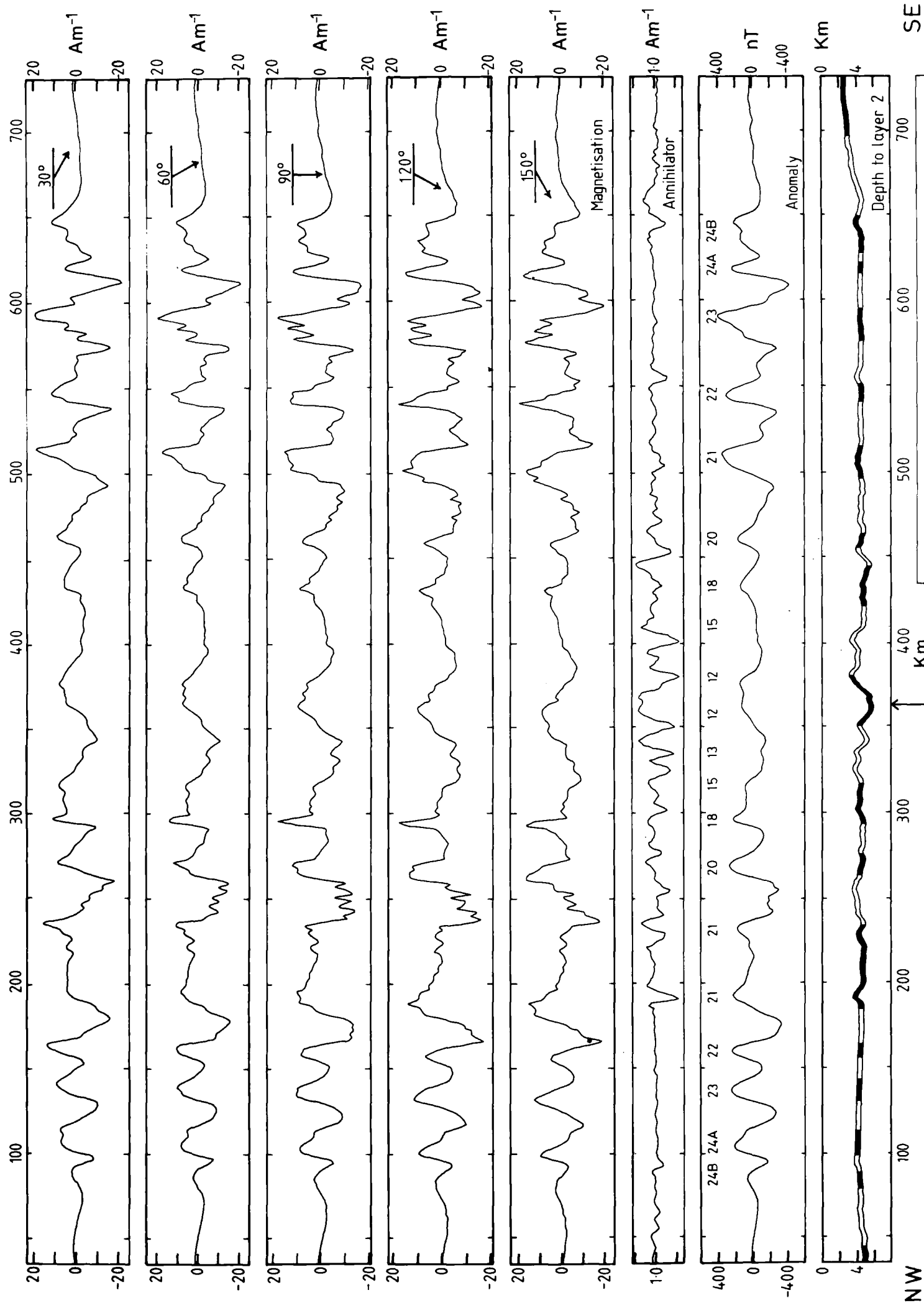


Figure 5.4
Magnetic anomaly profile across
the Norwegian Sea (689°N, 7.3°W to 639°N, 19°E)
with inversions using a range of
magnetisation inclinations.

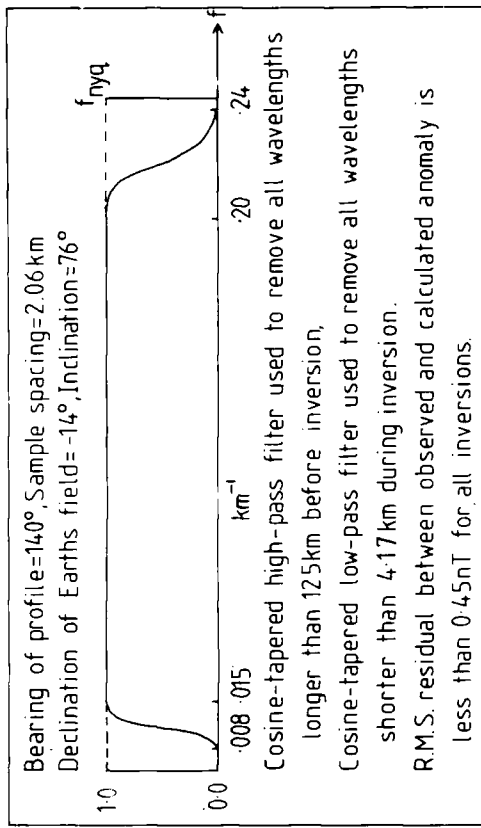


Figure 5.4) as follows

$$M = M' (1 + \tan^2(\beta - D_m) \cos^2 I'_m)^{0.5}$$

$$I_m = \arctan(\cos(\beta - D_m) \tan I'_m) \quad 0 < I_m, I'_m < 180^\circ$$

For an average value, $D_m = -10^\circ$ as assumed above then the true model magnetisations are never more than 15% greater than those shown in Figure 5.4 and the true inclinations are never more than 4° shallower than those in Figure 5.4 (although they have an opposite sense, because the bearing of the line is opposed to the assumed declination, for example $I_m = 134^\circ$ if $I'_m = 50^\circ$).

The purpose of performing inversions with different inclinations of magnetisation was to determine the inclination which gives rise to the most satisfactory looking magnetisation distributions. One would expect the average magnitude of magnetisation to remain approximately constant within each positively or negatively magnetised strip of crust; thus if the right inclination is chosen the corresponding model magnetisation should have the appearance of a square wave, provided that the various assumptions implicit in the inversion technique are realistic.

Considering a positive-negative anomaly pair on the eastern side of the basin (such as anomaly 22 and the negative anomaly to the west of it) it can be seen that for small inclinations of magnetisation (30° and 60°) the magnitudes of the positive and negative magnetisations increase eastward along the profile, such that the magnetisation distribution has a saw tooth appearance. As the assumed magnetisation vector is rotated further clockwise this effect becomes less prominent and at an angle of 120° the magnetisation distribution is quite square. With further rotation (150°), the distribution becomes skewed in the opposite sense, that is, the magnitudes decrease along the profile.

Inspection of Figure 5.4 shows that the magnetisation distributions for all anomalies east of the Aegir axis are squared up at an angle of inclination of 120° , but those to the west of the axis are squared up at an angle of about 90° . This is a curious result. The inclinations of magnetisation

within the plane of the profile which one would expect on the basis of the axial dipole hypothesis ($I_m = 70^\circ$) are 109° for the western part of the profile ($D_m = -20^\circ$) and 106° for the eastern part ($D_m = 0^\circ$); note that these do not differ in the sense inferred from Figure 5.4. Taking the evidence of Figure 5.4 at face value it appears that on both sides of the Aegir axis the magnetisation vector is rotated away from the expected direction towards the axis, by about 10° to 20° . As the increment by which inclination is changed in Figure 5.4 is coarse (30°), no claim is made for the accuracy of this value, but it was not thought to be worthwhile to use a smaller increment as the estimation of "squareness" is subjective. Nevertheless it is believed that the effect is a genuine one, for which there are at least two classes of explanation.

Firstly there could be actual deviations of the direction of in situ magnetisation. One possible cause of an effect of this sort is tectonic rotation. It is well known that faulting and tilting of oceanic crust takes place as it moves away from the accretionary axis. In general the fault scarps are inward facing and the faulted blocks tilt away from the axis. As the magnetisation of the upper part of the crust is acquired prior to faulting (Klitgord and Mudie 1974) deviations of the magnetisation vector in the sense described must surely occur. Over the Reykjanes Ridge, fault blocks have been observed to tilt away from the axis by as much as 20° to 30° , although tilts of about 10° are more common (Laughton and Searle in press). Over the mid-Atlantic ridge crest in the FAMOUS area, however, the average tilt is only 2° to 3° (Macdonald and Luyendyk, 1977). Luyendyk (1979) measured magnetisation inclinations for the northeast Atlantic sites of DSDP Leg 49 and found them to be generally shallower than expected on the basis of the site paleolatitudes, which he took to be equal to their present latitudes. Luyendyk's results were presented in terms of computed and actual paleocolatitudes, but if the inclinations are recomputed it is seen the actual inclinations are from 2° to 16° too shallow. Luyendyk suggested tectonic tilting and demagnetisation effects in massive lava flows as possible causes. There also may be a bias in the measurements due to

inadequate sampling of the secular variation. It should be noted that demagnetisation within lava flows would be expected to alter the inclination equally on both sides of the axis, so would not produce the type of deviation described for the Norway Basin profile. Previous DSDP measurements have in some instances produced similar results but on the whole there is insufficient evidence to show that deviations of average inclination are a general feature of the upper oceanic crust (Lowrie, 1977; Johnson, 1979).

An alternative explanation is that the magnetisation inclinations along the profiles are similar to those anticipated on the basis of the axial dipole hypothesis and that the deviations noted in Figure 5.4 are due to variations which are not taken into account in the modelling procedure. In particular, an implicit assumption is that the boundaries between positively and negatively magnetised blocks of crust are vertical. If this is not the case, apparent deviations of magnetisation will occur. As a simplification consider a strip of crust to be approximated by a uniformly magnetised flat topped dipping dyke, of infinite depth extent. The magnetic anomaly due to such a dyke depends on the difference between the dip of the dyke and the inclination of magnetisation (projected perpendicular to strike) but not on either of these quantities independently (Bruckshaw and Kunaratnam, 1963). Thus for example, a vertical dyke in which the magnetisation is inclined 120° gives rise to the same magnetic field as a dyke dipping 80° in which the magnetisation is inclined 110° . A possible explanation of the apparent deviation of magnetic inclination on either side of the Aegir axis is therefore that the boundaries between blocks of crust with different magnetisations dip away from the axis.

On the basis of an extensive study of ophiolite complexes Kidd (1977a) produced a model for the processes of formation of the upper oceanic crust which predicts the shape of the reversal boundaries (Figure 5.5). Within layer 2, the reversal boundaries probably follow isochrons. According to Kidd's model the isochrons dip towards the axis at about 30° within the upper part of layer 2, which is composed of

Figure 5.5. Inversion of a model magnetic anomaly profile which was computed using Kidd's (1977a) magnetic source layer configuration. Inversions for assumed layer thicknesses of 0.5 and 3.0 km, and magnetisation inclinations of 100° and 110° . The true inclination of magnetisation is 100° but it is apparently deviated to 110° (judging from the "squareness" of the magnetisation distribution) because of the effect of the lower boundaries, which dip away from the axis (compare Figure 5.4).

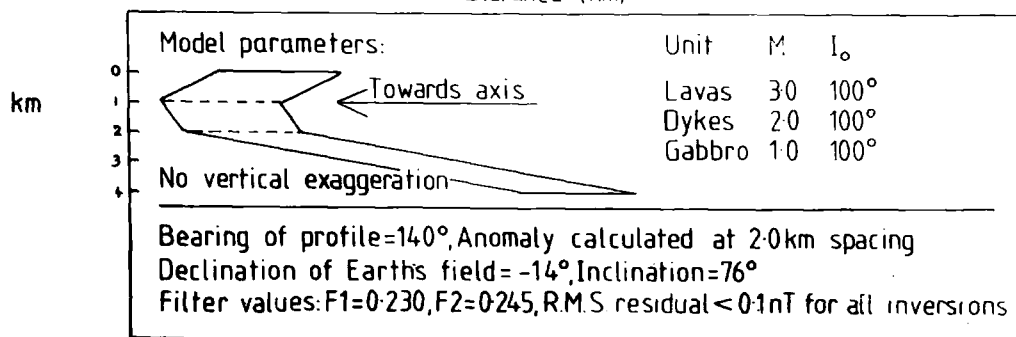
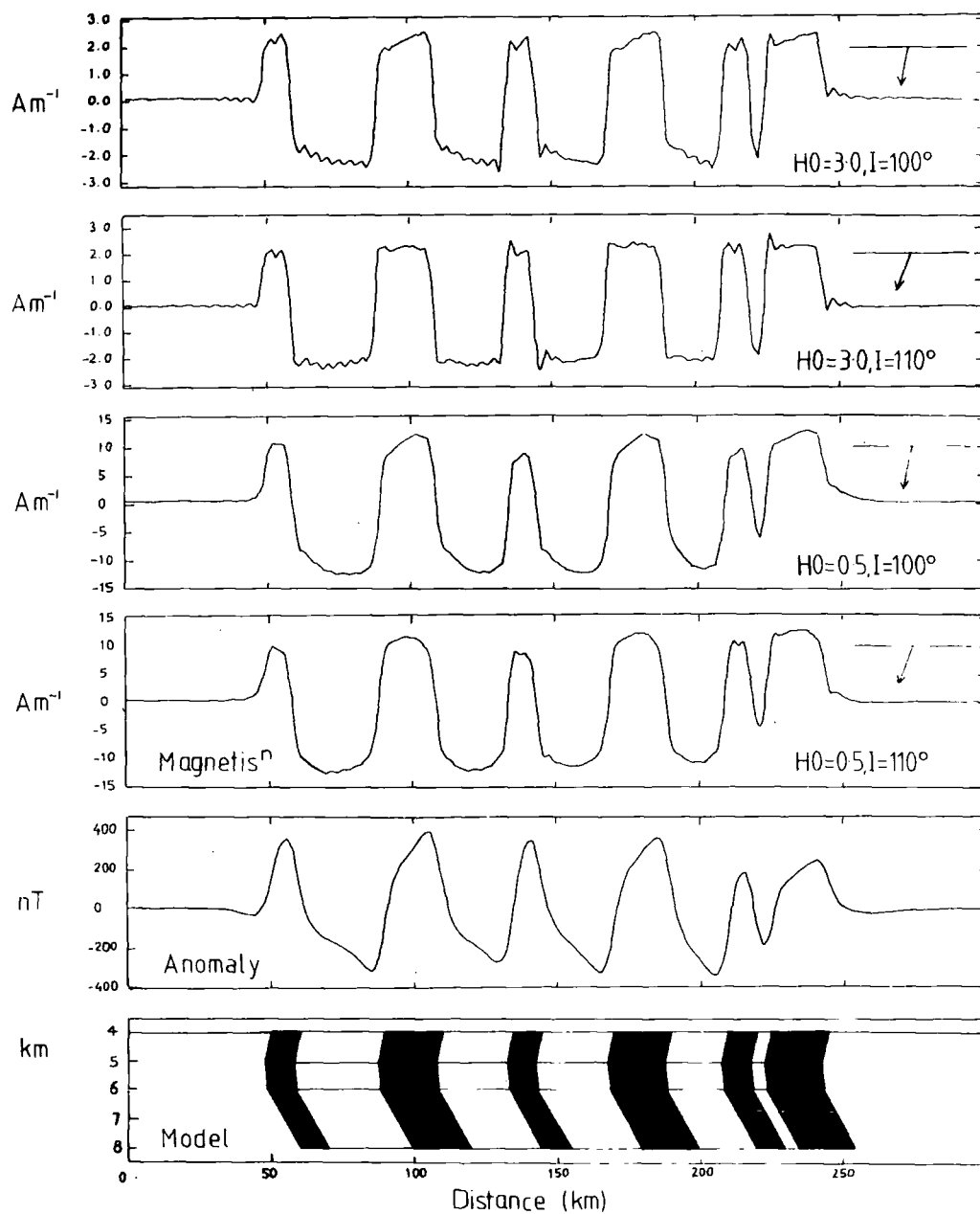


Figure 5.5

extrusive lavas and then steepen and overturn to dip away from the axis at a steep angle (80°) within the sheeted dyke complex. In the lower part of the dyke complex and in the gabbros comprising the upper part of layer 3, the reversal boundaries probably follow the Curie point isotherm, which is inferred to dip away from the axis at about 10° .

Figure 5.5 shows a model anomaly profile (roughly based on the eastern Norway Basin, anomalies 20 to 24) which was computed using the departmental program MAGN (Bott, 1975b). The reversal boundaries are simplified from Kidd's and the magnetisation intensities to the various units are loosely based on the values given by Harrison (1976) for lavas, dykes and gabbros. The inclination of magnetisation is 100° in all the layers. The model anomaly profile, which compares quite well with the observed profile, was inverted using layer thicknesses of 3.0 and 0.5 km. In both cases, skewed magnetisation distributions were obtained when the correct inclination of magnetisation was used. On the other hand squared distributions resulted when the inclination was deviated 10° towards the axis. These results show that if the reversal boundaries have the shape suggested by Kidd then significant deviations of the sort observed for the Norway Basin will arise. The effects produced by the model of Figure 5.5 do not appear to be as great as those that are observed. As the upper layer, which dips towards the axis, diminishes the effect of the deeper layers, which dip away, it is clear that a greater effect would arise if the deeper layers contributed relatively more to the anomaly. However it seems more likely that a combination of the effects of tilting in the upper layers and the dipping boundaries at depth produce the observed deviations.

Similar apparent inclination anomalies have been described and studied by previous authors, most extensively by Cande and coworkers, although this was not realised by the present author at the time that Figures 5.4 and 5.5 were constructed. Cande (1976) noted that "anomalous skewness can be explained either by modifying the assumed structure of the magnetic source layer or the assumed behaviour of the paleomagnetic field". Cande and Kent (1976) showed that a

two layer source, in which the deeper reversal boundaries are determined by the Curie point isotherm could account for the shapes of magnetic anomalies in the Pacific and southeast Indian oceans. More recently Cande (1978) has dismissed crustal tilting as a major cause of anomalous skewness and has shown that in some cases anomalous skewness may be related to long period (1 Ma) intensity variations of the paleomagnetic field. In addition Kidd (1977b) has elaborated on his model for magnetic source body shape, and has recommended that a standard shape be used in interpretation. In this context it is noted that it is possible to modify the formulae of Parker and Huestis (1974) to make the basic element approximate to the shape proposed by Kidd, using a number of dipping planes. The magnetisation could vary between layers, providing the ratios were constant horizontally, and the horizontal magnetisation distribution in any one layer could be found by inversion.

Discussion of magnetisation values

The magnetisation intensities obtained for the Norway Basin profile, using layer thicknesses of 0.5 and 3.0 km will be compared in this section with those inferred or measured directly in other regions. Firstly it should be noted that no annihilator has been added to any of the magnetisation distributions presented, so the mean magnetisation level in all cases is zero. It would be reasonable to add in a scalar multiple of the annihilator if a large component of induced or viscous remanent magnetisation was expected to be present. Assumption of the wrong mean level for the magnetisation distribution would give rise to correlations between the distribution and the annihilator. In the case of the Norway Basin profile, no noticeable correlations exist between the inverted magnetisation distribution and the annihilator if the mean magnetisation level is chosen to be zero, but on the other hand there are also no noticeable correlations if the mean level is chosen to be several $A\ m^{-1}$. This is because the fluctuations of the annihilator relative to its mean level are small (less than 20%) and the magnetisation distributions are intrinsically noisy. Thus the anomalies themselves

provide no evidence for or against the existence of a substantial mean level of magnetisation.

It is customary to assume that the remanent magnetisation dominates the induced magnetisation and this is borne out by direct measurements on DSDP cores. For example the average ratio of remanent to induced magnetisation is 21 for Leg 49 basalts (Steiner et al., 1979) and 35 for Leg 37 basalts (Ryall et al., 1977). Available data suggests that remanence also dominates induced magnetisation for deeper crustal rocks, but by a smaller factor (Hall 1976). The importance of in situ viscous remanent magnetisation (VRM) is one of the major problems concerning the magnetisation of the oceanic crust (Hall, 1976). A number of laboratory studies have shown that some submarine basalts can acquire substantial VRM components in geologically short times (Johnson, 1979). If VRM acquired in the direction of the earth's present field is additive independent of the direction of the original thermal remanent magnetisation then it will have little effect upon sea floor spreading anomalies. This is commonly assumed, but is not proven (Lowrie and Kent, 1978). However as there is no coherent alternative hypothesis, there is little choice but to assume that VRM is either negligible or uniform.

The most straightforward way of comparing the magnetisations inferred along the Norway Basin profile with results from other studies is via the work of Harrison (1976) who compiled a variety of data concerning the magnetisation of the oceanic crust. While recognising that a layer thickness of 0.5 km is probably not realistic Harrison took advantage of the fact that a substantial amount of modelling has been performed using this layer thickness in order to compare magnetisations from different areas. He also reduced magnetisations inferred using other layer thicknesses to roughly equivalent values for a 0.5 km layer. In addition, because the intensity of natural remanent magnetisation is generally proportional to the intensity of the axial dipole magnetic field at the time of crustal formation, he reduced the values to the equator by dividing by $(1+3 \sin^2 \phi)^{0.5}$, where ϕ is the paleolatitude. The mean of 12 values from

various sources, reduced in this manner is 9.3 A m^{-1} and the standard deviation is 3.5 A m^{-1} .

For the Norway Basin profile magnetisations are best estimated from the "squared" distributions of Figure 5.4 (the correction for declination is negligible for steep inclinations). The characteristic intensity of magnetisation for anomalies 21 to 23 is about $10 - 12 \text{ A m}^{-1}$, or referred to the equator, about $6 - 7 \text{ A m}^{-1}$ which falls within the range of values reported by Harrison. The magnetisation associated with the younger anomalies in the centre of the basin are significantly lower at about 7 A m^{-1} , or referred to the equator, about 4 A m^{-1} , which is at the bottom of the range reported by Harrison. Low anomaly amplitudes are characteristic of the central parts of the Norway Basin, particularly in the north (Chapter 6). In addition, the magnetisations for anomaly 24A and 24B on both sides of the basin are low, but whether or not this is significant is difficult to say, as the charts of Chapter 6 show anomaly 24 to be quite variable in amplitude. In summary, the magnetisations inferred for the Norway Basin profile are less intense than average, judging from the limited data set of Harrison (which is mostly based on surveys over crust less than 10 Ma old).

The results of DSDP Leg 49 showed that there are no significant differences in rock magnetic properties ^{between} the northeast Atlantic and the FAMOUS area further south, although differences were expected because of the presence of the Icelandic hot spot (Steiner et al., 1979). The average upper crustal stable remanent magnetisations were found to be quite low (about 5 A m^{-1}). If similarly low magnetisations exist in the Norway Basin, which seems likely, it is clear that the magnetised layer must be thicker than 0.5 km. A minimum layer thickness is about 1 km (for anomalies 21 to 23). In this case there would be no contribution to skewness from deeper boundaries and it seems more likely that magnetisation intensity decreases with depth so that the overall thickness is greater.

Transition widths

There has been considerable research into the width of the transition zone between normally and reversally magnetised crust. From near bottom studies over young crust, the transition zone has been found to be a few kilometres or less in width, but it has been suggested that it increases in width with crustal age, as a result of changes of magnetisation (Blakely and Cande, 1979). The transition widths measured from the "squared" profiles in Figure 5.4 vary between about 4 and 7 km. These values are quite close to the resolution limit imposed by the low pass filters (about 3.3 km) and because of this the possibility that the transition widths are only 2 to 3 km cannot be ruled out.

Chapter 6

CORRELATION AND IDENTIFICATION OF MAGNETIC ANOMALIES AND THE
LOCATION OF OCEAN-CONTINENT BOUNDARIES

6.1 The magnetic anomaly pattern of the northeast Atlantic

In this section a general account is given of the pattern of sea floor spreading anomalies in the northeast Atlantic. In the present context, the northeast Atlantic is defined as the oceanic region which is bounded to the northeast by the Greenland and Senja fracture zones and to the southwest by a line joining the southern tip of Greenland to the southwest margin of Rockall Plateau (Figure 1.1). It is believed that the present phase of sea-floor spreading in the northeast Atlantic began shortly before anomaly 24 time (Section 1.1). The detailed evidence for this assertion will be considered in Section 6.2. During an earlier phase of spreading, oceanic accretion occurred in the Rockall trough (Section 1.1) and also perhaps further north, adjacent to the present Norway Basin (Bott, 1975c). During the present phase of spreading, accretion occurred in the following regions (Figure 1.1)

- (i) the Lofoten and Greenland Basins (about the Mohs Ridge) in the central Norwegian-Greenland Sea
- (ii) the Norway Basin (about the Aegir axis) followed by the Iceland Plateau (about the Kolbeinsey Ridge)
- (iii) the Icelandic transverse ridge (Bott 1974), comprising the Iceland-Faeroe Ridge, the Greenland-Iceland Ridge (Denmark Straits) and Iceland itself
- (iv) the eastern and western Reykjanes basins (about the Reykjanes Ridge).

Rotation poles for the northeast Atlantic appear to have been always situated well to the north of the area (Talwani and Eldholm, 1977), so under the assumption of the rigidity of Greenland and Eurasia, the overall spreading rates in the southern Norwegian-Greenland Sea can be inferred to be similar to those in the north and the south. Accordingly,

the well understood anomaly patterns in regions (i) and (iv) can be compared directly with those in the southern Norwegian-Greenland Sea, which are less well understood. The possibility of local non-rigidity adjacent to the southern Norwegian-Greenland Sea cannot be ruled out a priori, but there does not appear to be any geological evidence to support the concept.

This section is divided into two parts. In the first, a description is given of the characteristic form of the anomaly sequence 1-24 in the northeast Atlantic. In the second, a description is given of the lineation patterns to the north and south of the southern Norwegian-Greenland Sea.

Anomalies 1-24 in the northeast Atlantic

Figure 6.1 shows a synthetic anomaly profile, computed for a source depth of 4 km for the eastern side of the northeast Atlantic at the latitude of the southern Norwegian-Greenland Sea. The profile was computed using Kidd's (1977b) model of magnetic source body shape (see the caption for details). The profile has the same azimuth (120°) as many of the tracklines across the Norway Basin in Figure 6.5; this azimuth roughly corresponds to the average of the initial and final spreading directions in the southern Norwegian-Greenland Sea (147° and 104° respectively), as measured from Talwani and Eldholm's (1977) synthetic flow lines. The half spreading rates used in the computations for Figure 6.1 are thought to be fairly realistic for this azimuth but no claim is made for their accuracy. The corresponding rates for the Heirtzler et al. (1968) timescale would be 30% lower for anomalies 7 to 24.

The anomaly shapes and relationships which are shown in Figure 6.1 accord fairly well with those which are observed in the northeast Atlantic. However identification of anomalies in the southern Norwegian-Greenland Sea was actually carried out using real anomaly profiles for comparison and the synthetic profiles are shown mainly for the purposes of discussion, as an appropriate real profile, showing all of the anomalies well developed and free of

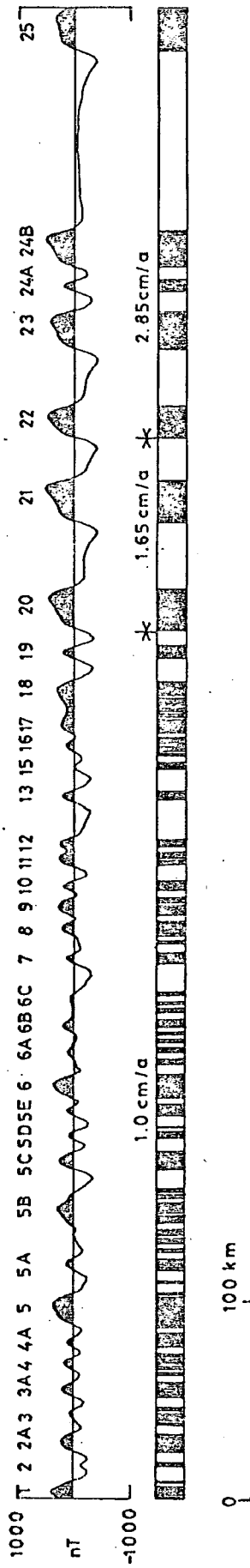


Figure 6.1. Synthetic magnetic anomaly profile for the eastern side of the northeast Atlantic at a latitude of about 67°N . Inclination of earth's field 76°N , declination -14° . Bearing of profile 120° . Magnetic source body configuration follows Kidd's (1977b) model; as in Figure 5.5 with some modifications. Lower boundary dips 20° (appropriate to a half spreading rate of 2cm/a). In the plane of the profile, the inclination of the magnetisation is 110° within the lavas (10° of tilt towards the axis) and 100° within the lower layers. Corresponding magnetisations are 5 A/m and 1 A/m . Depth to the top of the source layer is 4.0 km (unrealistically large for the younger anomalies). Based on Hailwood et al. (1979) timescale, with averaged spreading rates.

tectonic disturbance, could not be obtained.

Important features of the anomaly sequence 1 to 24 in the northeast Atlantic, as exemplified by the profile in Figure 6.1 and also by real examples are as follows.

(i) Anomalies 1 to 5 usually have high amplitudes and can be "counted out" from the axis. Anomaly 5 is particularly distinctive, and is identifiable throughout the northeast Atlantic. Over the Iceland Plateau there are complications due to shifts of axis since anomaly 5 time, but the sequence has been sorted out using closely spaced data by Vogt et al. (1980).

(ii) Anomalies 5A to 6C do not have a distinctive signature, and cannot be readily identified out of sequence. This is illustrated by the confusion which arose over the identification of anomalies over the central Iceland Plateau (Vogt et al. 1980). When anomalies 5A to 6C are seen in context the negative anomaly between 5B and 5C is recognisable and anomaly 6 is also quite prominent, although rarely as prominent as it appears on synthetic profiles.

(iii) Anomalies 7 to 12 are not well resolved. Anomalies 7 to 12 tend to coalesce to form a single low amplitude irregular positive anomaly. The negative anomaly between 6C and 7 is generally quite prominent and anomaly 12 often forms a peak, although not universally.

(iv) Anomaly 13, when developed, is distinctive, as it is a small peak in quite a broad and deep negative anomaly. As suggested by La Brecque et al. (1977), anomaly 14 has been eliminated from the timescale which was used to prepare Figure 6.1 and the resulting appearance of the synthetic anomaly 13 is more realistic than if anomaly 14 is included which can be seen by comparison of synthetic and real profiles in Vogt and Avery (1974).

(v) Anomalies 15 to 18 form a broad irregular positive band within which anomaly 18 is generally the most well developed anomaly.

(vi) Anomaly 19, like anomaly 13, is distinctive when present, but is not always developed. None of the anomalies from 7 to 19 appear to have been definitely identified in the southern Norwegian-Greenland Sea.

(vii) Anomalies 20 to 24 are characteristically well developed. The sequence as a whole is easy to identify throughout the northeast Atlantic, but in the Norway Basin the similarity of the shapes of the individual anomalies leads to the possibility of misidentification. An important feature is that the negative anomaly between anomalies 20 and 21 is significantly broader than the other negative anomalies.

Because the sequence in the Norway Basin is rather confused and the existence of anomaly 24 has been doubted (Talwani and Eldholm, 1977) it is important to establish the typical shape of anomaly 24. In the synthetic example of Figure 6.1 anomalies 24A and 24B are separately resolved, in consequence of the relatively high initial spreading rate assumed in the model. This rate is substantially higher than those assumed by previous authors, with the exception of Featherstone (1976). If a lower initial half spreading rate of about about 1.5 cm/a is assumed, anomalies 24A and 24B coalesce and are separated from anomaly 23 by a narrow weak negative anomaly (see Vogt and Avery (1974), Voppel et al. (1979) for examples). However both north and south of Iceland anomaly 23 is typically bordered by quite a pronounced negative anomaly (10-20 km wide), then a broad positive anomaly (about 30 km wide) which is irregularly developed but often double peaked (examples are given below). The simplest explanation is that the double peaked anomaly comprises anomalies 24A and 24B, more or less separately resolved. This implies quite a high spreading rate, however an equally high rate is implied by the width of the negative anomaly landward of anomaly 23. The spreading rate assumed in Figure 6.1 was averaged (from Figure 6.11) over about 9 Ma. It is considered that the half spreading rate at anomaly 24 time may have been as high as 3.5 cm/a.

The magnetic anomaly pattern in the Reykjanes basins

Magnetic anomalies in the Reykjanes basin have been extensively studied. Data collected prior to 1974 have been collated and presented by Vogt and Avery (1974). Subsequently new data have been presented by Johnson et al. (1975a), Johnson et al. (1975b), Featherstone (1976), Voppel et al. (1979) and Voppel and Rudloff (1980).

Figure 6.2 shows a portion of the detailed aeromagnetic chart of Vogt and Avery (1974), off the Rockall margin. Anomalies 20 to 24 are well developed and the negative anomaly between 20 and 21 is conspicuous. The anomaly marked 24 by Vogt and Avery is narrow and discontinuous and it is not clear whether they considered it to be 24A, or 24A and 24B combined. It is considered likely that this anomaly is in fact 24A and that the broader positive anomaly on its landward side is 24B, as assumed by Hailwood (1979).

Adjacent to the Rockall Margin, anomalies in the approximate range 5C to 18 are disturbed by closely spaced fracture zones, three of which are shown in Figure 6.2. Vogt and Avery considered that fracture zone formation was consequent on the cessation of spreading in the Labrador Sea, when the spreading direction across the Reykjanes Ridge changed from about 130° to 100° (Section 1.1). Starting from just after anomaly 18 time the strike of the magnetic lineations became more northerly, reorienting at a rate of 1° to 4° /Ma. By about anomaly 7 time reorientation was complete and the fracture zones changed from being curved to being straight. Vogt and Avery (1974) suggested that the fractures formed when the Reykjanes Ridge attained a "minimum work" configuration by attempting to orient itself normal to the new spreading direction.

By anomaly 5 time however, the fracture zones disappeared completely and the Reykjanes Ridge was spreading obliquely. The change to oblique spreading seemed to occur first in the north, and moved southward at a rate of about 2 cm/a. Vogt and Avery pointed out that this change was enigmatic if the ridge had already attained a minimum work

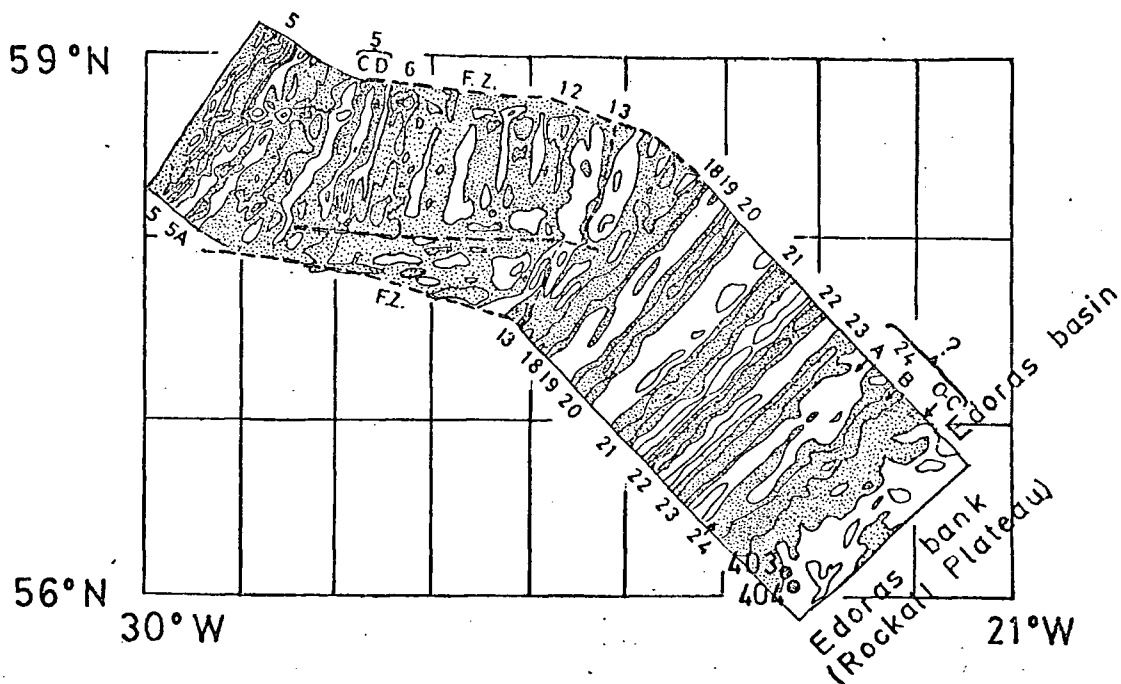


Figure 6.2. Residual magnetic field in the East Reykjanes Basin, redrawn after Vogt and Avery (1974, Figure 2a). Contour interval is 250 nT, stippled areas are positive. Fractures are dashed. Anomaly identifications follow Vogt and Avery (with some additional anomalies explicitly identified). Shown on the upper side of the strip are identifications of anomalies 24A and 24B according to the present study, apparently differing from Vogt and Avery's simple identification of a single anomaly 24 (lower side). The locations of DSDP holes 403 and 404 in the Edoras basin are shown. The tentative ocean-continent boundary (O-C) differs from that of Roberts et al. (1979), who consider it to be coincident with the anomaly which is labelled 24B.

configuration, but gave possible explanations in terms of asthenospheric flow away from Iceland. In Chapter 7 a purely geometrical reason is suggested for the formation and disappearance of the fracture zones, based on an interpretation of spreading history and geometry in the southern Norwegian-Greenland Sea.

Voppel and Rudloff (1980) have identified similar closely spaced fracture zones in the western Reykjanes basin between 56°N and 60°N and have correlated the offsets in anomaly 13 on both sides of the axis. Magnetic anomalies over the East Greenland margin have been compiled by Voppel et al. (1979) from a variety of earlier sources. South of 63.5°N anomaly 24 lies about 60 km seaward of the base of the continental slope and about 100 km from the coast. Anomaly 24 is flanked on the landward side by a narrow minimum then a zone of irregular positive anomalies. Johnson et al. (1975b) have suggested that anomaly 25 may be present at 62°N but later workers have not pursued this idea. Featherstone (1976) modelled anomaly 24 to 22 at 61.3°N and 63°N , using a half spreading rate of 2.4 cm/a and the Heirtzler et al. (1968) timescale, corresponding to a rate of 3.3 cm/a for the Hailwood et al. (1979) timescale. His synthetic anomaly 24 is resolved into two peaks which agree in position with the observed peaks, although the relative negative anomaly between 24A and 24B is not as well developed in the observed profiles as it is in the model profiles.

North of 63.5°N on the East Greenland margin, anomalies 23 and 24 cut out (Figure 1.1). Featherstone et al. (1977) inferred that the locus of spreading was initially sinuous and was straightened by a westward shift of axis before anomaly 22 time. The strip of oceanic crust formed prior to the shift is found on the opposite margin, just north of Hatton Bank, where Voppel et al. (1979) have confirmed that additional magnetic anomalies are present between 61°N and 62°N .

The magnetic anomaly pattern of the central Norwegian-Greenland Sea

Figure 6.3a, which is reproduced from Vogt et al. (1978) shows residual magnetic anomalies plotted along flight lines for the southern part of the USNRL/USNOO aeromagnetic survey, which extends from the central Norwegian-Greenland Sea into the Arctic Ocean (Phillips et al., in press). Figure 6.3b, which is from the same sources, shows the corresponding "zebra-stripe" chart. The most southerly trackline of the survey follows the western Jan Mayen Fracture Zone and continues with the same trend further east, to end at about 68.3°N , 10°E , just north of the Voring Plateau. Between this trackline and the eastern Jan Mayen Fracture Zone there are a number of published marine profiles (Gronlie and Talwani 1978) which are included on Figure 6.5 (Section 6.3).

Anomaly 24 is bordered on both sides of the Mohns Ridge by extensive smooth zones and is often double peaked, particularly on the western side of the ridge between 75°N and 76°N . The zebra stripe chart indicates that an extensive system of closely spaced fracture zones is probably not developed in the central Norwegian-Greenland Sea. For example the anomaly 15-18 band is regularly developed on both sides of the ridge and anomaly 13 does not appear to be offset. Within 100 km of the western Jan Mayen Fracture Zone, however, the anomaly pattern is disrupted and there is at least one fracture zone offsetting the Mohns Ridge, as shown in Figure 1.2 and Figure 6.6.

6.2 The age of the northeast Atlantic

In Section 1.1 it was stated that the present phase of spreading in the northeast Atlantic began shortly before anomaly 24A time. This is widely, but not universally accepted. A number of authors have suggested that spreading was taking place for a considerable length of time prior to anomaly 24 time. In this section a summary is given of the evidence for an abrupt spreading onset shortly before anomaly 24 time. The purpose of this discussion is quite specific

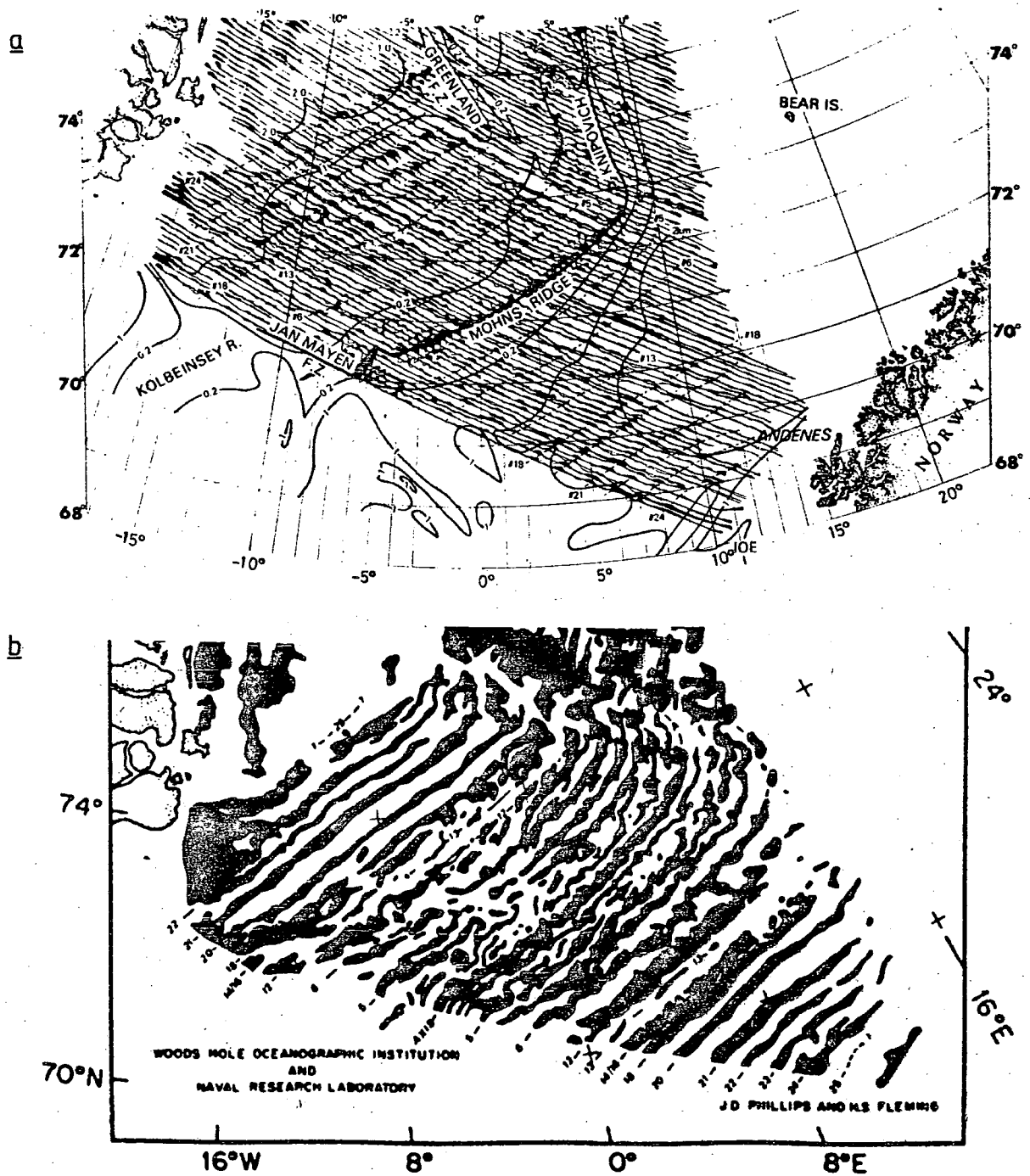


Figure 6.3 Aeromagnetic survey of the central Norwegian-Greenland Sea (Phillips et al. in press). Retouched from Vogt et al. (1978).

a Residual total field anomalies plotted along flight lines

b 'Zebra stripe' residual chart

with regard to the investigation of the southern Norwegian-Greenland Sea. At present there is not enough geophysical and drilling evidence from the southern Norwegian-Greenland Sea to infer unambiguously the age of the Norway Basin and the positions of the ocean-continent boundaries. As described in Chapter 1, the argument that the Jan Mayen Ridge is continental is strengthened if it is known that spreading began shortly before anomaly 24 time. Otherwise one must rely on the seismic reflection characteristics of Horizon 0 and the presence of a magnetic smooth zone as evidence for the continental nature of parts of the Jan Mayen block, because the seismic refraction and gravity anomaly evidence which has previously been cited does not appear to be conclusive. Two types of evidence for the age of the northeast Atlantic are considered as follows.

Evidence from magnetic anomalies considered in relation to known ocean-continent boundaries and from deep sea drilling

Throughout the northeast Atlantic anomaly 24 is the oldest anomaly which has been positively identified. However one cannot immediately conclude that older oceanic crust is not present on the grounds that the associated magnetic anomalies are not present, as these may not have been produced (Vogt et al., 1980) or may have been subsequently erased for some reason. It is therefore necessary to consider the position of anomaly 24 in relation to known ocean-continent boundaries.

Known and inferred ocean-continent boundaries in the northeast Atlantic are shown in Figure 1.1. On the western margin of the Rockall microcontinent substantial amounts of spreading prior to anomaly 24 time is ruled out by the known continental nature of Hatton Bank (Roberts, 1975). The ocean-continent boundary is most closely defined in the vicinity of DSDP sites 403 and 404 (Figure 6.2), which were drilled in the Edoras basin (Montadert and Roberts, 1979) which is a marginal basin lying at the foot of the outer Rockall Plateau. The Edoras basin contains a considerable thickness of sediment, the oldest of which could be mid-Cretaceous in age, and is presumed to be underlain by

continental crust. It is terminated on the seaward side by a basement high. The exact position of the ocean-continent boundary is unclear. Roberts et al. (1979) consider it to be some 20 km seaward of the edge of the basin, just east of the anomaly 24 of Vogt and Avery (1974). If the stronger landward anomaly is 24B, as shown in Figure 6.3, the ocean-continent boundary must be very close to the edge of the basin and virtually no spreading could have taken place prior to anomaly 24 time. The apparent oceanic basement is underlain by dipping reflectors as far west as anomaly 22.

The drilling results indicate that the transition from rifting to spreading may have taken place in less than 2 Ma (Roberts et al., 1979). The oldest sediments which were penetrated are late Paleocene deltaic deposits, which are believed to represent the top of the pre-drift sequence. The earliest Eocene sedimentary succession is dominated by the products of explosive basaltic volcanism which were deposited in shallow water. This phase lasted from 54 Ma until the beginning of anomaly 24B time at 52 Ma (Dinocyst zones Ia & Ib, NP9(top)-11, P6a-P6b). The base of the volcanoclastic sequence is marked by a thick tuff which can be correlated with a widespread ash marker in the North Sea (Harrison et al., 1979). The top of the volcanoclastic unit can be traced on seismic sections over the basement high to merge with the oldest oceanic basement. During anomaly 24 time there was rapid subsidence concomittant with deposition of largely biogenic sediments in the Edoras basin.

On the East Greenland margin, Featherstone et al. (1977) consider that the ocean-continent boundary lies immediately west of anomaly 24, where there is a transition from oceanic basement to a smooth interface at similar depth, which is underlain by coherent eastward dipping reflectors, which may represent pre-drift sediments, perhaps penetrated by igneous sheets. However the general configuration of the East Greenland margin is very similar in many ways to the configuration of the Jan Mayen Ridge and the deep horizon which Featherstone et al. discovered can be compared directly to Horizon 0. Accordingly, in the absence of drilling control, the East Greenland results do not provide

any independent evidence which can be applied to the southern Norwegian-Greenland Sea, as any uncertainty that exists concerning the Jan Mayen ocean-continent boundary exists equally for the Greenland margin.

In the central Norwegian-Greenland Sea the ocean-continent boundary has been identified only in the Voring Plateau region, where Talwani and Eldholm (1977) consider it to be coincident with the Voring Plateau escarpment (Figure 1.2) which separates the basement high of the outer Voring Plateau from the deep sedimentary basin of the inner Voring Plateau. It is clearly established (Ronnevik et al., 1979) that the inner Voring Plateau, which is a magnetically smooth zone, predates anomaly 24 considerably, as the base Tertiary volcanoclastic reflector can be traced through to the North Sea and a much deeper mid (or base?) Cretaceous reflector is present.

Talwani and Eldholm (1977) have shown that the eastern edge of anomaly 24 crosses the outer Voring Plateau and lies some 50 km off the escarpment, from which it is separated by a predominantly negative anomaly. DSDP hole 338 (Talwani, Udinstev et al., 1976) was drilled into "acoustic basement" about 15 km northwest of the escarpment. The oldest sediment is of earliest Eocene age (about 53 Ma). The nature of the underlying basalt is problematical, not only because of its conflicting radiometric age (47 Ma, Kharin et al. (1976)) and normal magnetisation (Kent and Opdyke, 1976) which suggest that it might be intrusive, but also because it is underlain by deep reflectors (Hinz and Schluter, 1979a). Possibly some of the outer Plateau is underlain at depth by rifted continental crust (Hinz and Schluter, 1979a). These results show that a maximum of 50 km of pre-anomaly 24 spreading is allowed in the Voring Plateau region. A definite maximum age cannot be given, because of the problem of apparent basement, although the data are consistent with initiation of drifting just prior to anomaly 24 time. On the other side of the Mohns Ridge, where the ocean-continent boundary is not known, anomaly 24 lies within 100 km of the coast at 73°N, but diverges further north.

In summary, the marine geophysical and drilling evidence from south of Iceland suggests that oceanic accretion began almost precisely at the beginning of anomalies 24 time (52 Ma) and was preceded by about 1.5 Ma of intensive volcanism. North of Iceland the ocean-continent boundary may lie 30-50 km landward of anomaly 24. At a high initial half spreading rate of 3.5 cm/a this width corresponds to 1-1.5 Ma, hence it is possible, but by no means proven, that spreading north of Iceland was underway during the volcanic phase which is recorded in DSDP holes 403 and 404. The onshore manifestations of this volcanism is the subject of the next section.

Evidence from onshore geology

In early Tertiary times very large volumes of tholeiitic flood basalts were erupted onto the continental crust of East Greenland (Soper et al. 1976a, 1976b, Brooks et al. 1976). The major remnant of the lava pile is exposed between Kangerdlugssuaq and Scoresby Sund, but it is possible that most of the coastal area between 66°N and 76°N was once covered by flood basalts.

Paleontological studies of the sediments which are associated with the basalts (Soper et al. 1976b) indicate that the entire pile was erupted within a few million years in latest Paleocene and earliest Eocene time. A variety of paleontological correlations establish without doubt that the basalts were erupted over the same period as the volcanogenic sediments in the Edoras basin were being deposited. The entire pile appears to be reversely magnetised (Hailwood et al., 1973 ; Faller, 1975) as is the volcanogenic sequence (Hailwood, 1979). The lava pile was therefore formed in the period preceding anomaly 24 time, 54 to 52 Ma according to Hailwood et al. (1979) immediately prior to the onset of spreading south of Iceland.

Hailwood et al. (1973) obtained radiometric ages in the range 47-52 Ma for various lavas ^{from the top of the pile} in the Scoresby Sund region. Hailwood et al. (1979) have cited the preferred age for the top of the pile (52 Ma) as independent confirmation of the

relationship between absolute and biostratigraphic age which is used in their timescale.

However there is evidence that the true age of the basalts may be somewhat greater. Fitch et al. (1978), reworking earlier data, have concluded that the basalts were erupted between 56 and 53 Ma and that the most accurate estimate of the age of the upper part of the pile is 54.5 ± 1 Ma. This is consistent with the fission track age of 54.6 ± 1.7 Ma obtained by Brooks and Gleadow (1977) for the Skaergaard intrusion, which was probably emplaced into the pile as the upper sequence was being erupted. Fitch et al. (1978) present a variety of other dating evidence which suggests that the volcanic episode in the northeast Atlantic was complete 53 Ma ago. This suggests that some, or all, of the dates based on the Hailwood et al. (1979) timescale which have been presented may be about 1 Ma too young. It is not clear how this small uncertainty arises, or how it is propagated.

The Blosseville coast basalts have been related directly (for example Soper et al., 1976b) to the Faeroese lavas (Tarling and Gale, 1968) which have a revised radiometric age (Fitch et al., 1978) of 54-53 Ma and also to the late Paleocene basalt which was drilled on the Rockall Plateau (Laughton, Berggren et al., 1972). If it is considered intrinsically likely that all of these plateau basalts were erupted from the incipient mid-ocean ridge during a rifting stage immediately prior to the onset of spreading (for example Brooks (1973), but compare Featherstone (1976)), then their dating provides independent evidence for the age of the northeast Atlantic and of anomaly 24. Otherwise it can be argued that detailed correlation with DSDP drilling results from the Edoras Basin provides the first direct evidence that the basalt effusion predated active spreading.

6.3 Preparation of magnetic anomaly charts for the southern Norwegian-Greenland Sea

In order to prepare the magnetic anomaly charts which are described in the next section, two digital data tapes

were purchased from the primary data archiving service in the USA, the National Geophysical and Solar-Terrestrial Data Centre (NGSDC). The characteristics of these tapes are listed in Appendix C. The first tape, (RQ5046), which arrived in April 1978, contains marine geophysical data from within the latitude-longitude rectangle (50°N to 85°N , 80°W to 30°E). This tape was found to contain less relevant data than had been expected and in particular magnetic data is not present for the R.V. Vema cruises V29 and V30, which together account for a substantial amount of the coverage of the Norwegian-Greenland Sea to date. It was impossible to validate or revise earlier interpretations using only the data on this tape. In July 1979 the Lamont-Doherty Geological Observatory released the V29 and V30 data, making it publically available for the first time and a second tape (MT083A) was purchased, arriving October 1979. The large amount of redundant data on the first tape had proved a handicap during processing so the areal coverage was restricted to (60°N to 75°N , 30°W to 20°E).

The data from these two tapes and from Durham cruises SHACK476 and SHACK977 was supplemented by R.V. Akademik Kurchatov data which was digitised from profiles along ship's track (Talwani et al. 1979, figures 1, 12, 19).

Two methods of plotting were used. The first, the oblique projection, attempts to give a three-dimensional impression of the anomaly field. The ship's tracks are projected onto a horizontal plane using an approximate Mercator projection in which the latitude line spacing is constant with correct scale at the mean latitude. An anomaly "fence diagram" is formed by a point moving along the track with an elevation proportional to the magnetic anomaly. This construction is viewed from infinity along a given bearing and at a given elevation angle. The second plotting method was based on the 1:4 000 000 scale Mercator projection used by Gronlie and Talwani (1978). The ship's track is again used as a baseline and the anomaly curve is plotted so that each line from a point on the track to the corresponding point on the anomaly curve has the same bearing and has a length proportional to the anomaly value. Both projection

methods allow accurate presentation of closely spaced data without loss of information near turning points. Provided that the "angle of view" is chosen to coincide with the anomaly trend, anomalies from intersecting tracks interfere very little. In order to save plotting time a new point along the ship's track was not plotted if it lay less than 1 km from the previously plotted point. If consecutive points were more than 5 km apart a gap was left.

Large scale oblique plots were produced on the Geography Department 30 inch plotter, for the Iceland Plateau, the Jan Mayen block and the Norway Basin. When coloured, to distinguish positive from negative anomalies, these plots gave a good impression of anomaly shapes and interrelationships and were used for anomaly correlation. It was not possible to reproduce these plots, but as an example Figure 6.4 is a smaller version of the Norway Basin plot in which the positive anomalies have been shaded automatically (each vertical line represents a data point). A plot of the second type (with an "angle of view" of 30°) covering the whole of the southern Norwegian-Greenland Sea (60°N to 71°N , 30°W to 10°E) was used for comparison of anomaly trends with bathymetric and geophysical features on other charts. This plot was also produced on the Geography Department plotter but was redrawn for the purposes of reproduction (Figure 6.5). Over the Iceland-Faeroe Ridge the high density of track lines and the confused nature of the anomalies precluded reproduction of all profiles. The zero and 200 nT positive contour lines from Fleischer et al. (1974), as reproduced by Voppel et al. (1979), were traced on, together with representative profiles. No profiles at all were plotted in the vicinity of the Reykjanes Ridge, where the anomaly pattern has been intensively studied.

Agreement between the anomalies along closely spaced and intersecting tracklines were generally found to be good. The few mismatches which occur may be due to navigation errors or temporal variations. As all of the cruises used satellite navigation the latter explanation is more likely. It is not certain which reference field was used in producing the NGSDC and R.V. Akademik Kurchatov residual anomalies (in the

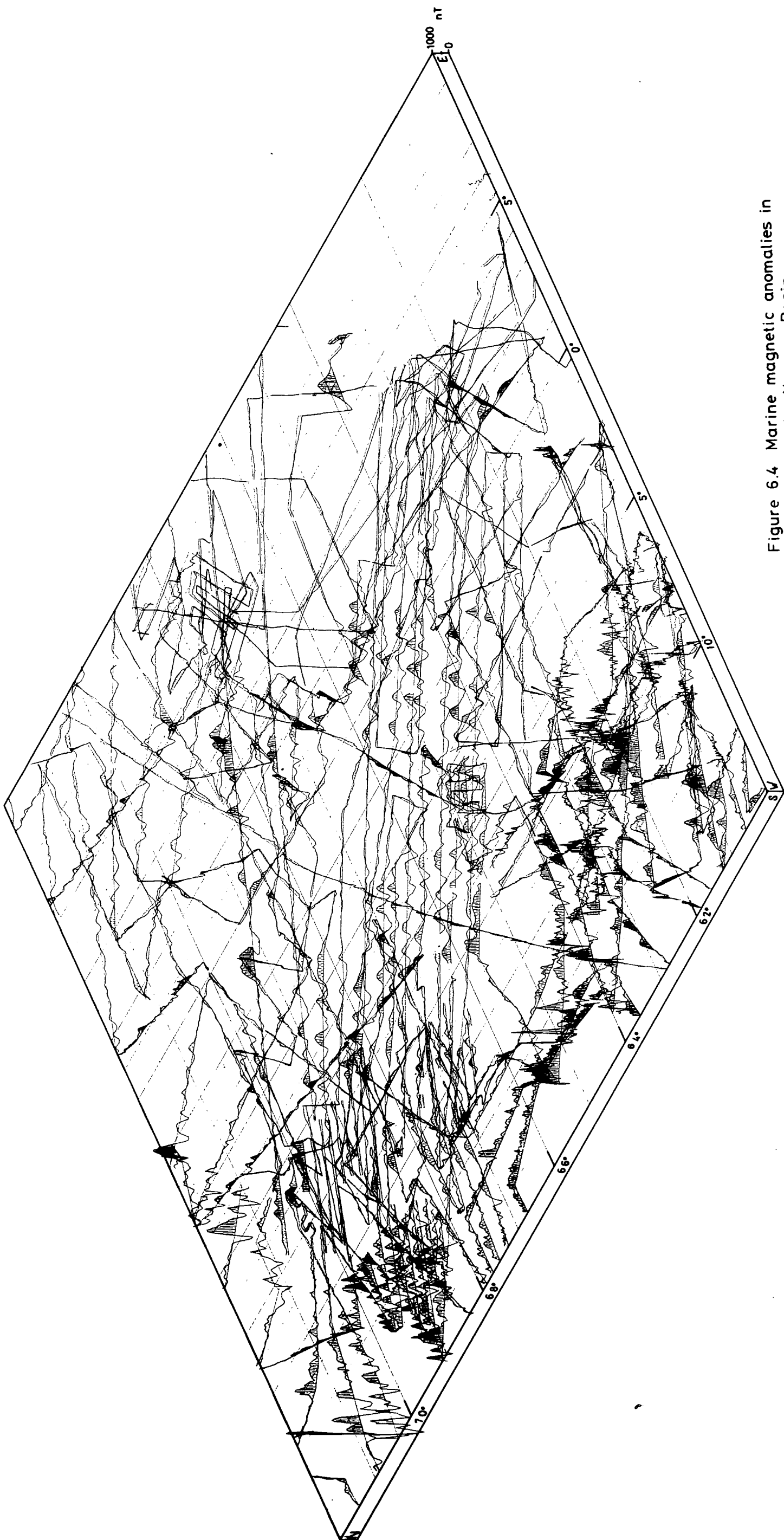


Figure 6.4 Marine magnetic anomalies in the region of the Norway Basin.

Figure 6.5. Residual total field magnetic anomalies over the southern Norwegian-Greenland Sea and surrounding areas. The anomalies are projected N30°E with the ship's track as a base line. The data is from a variety of marine surveys, as discussed in the text. Also shown for the Iceland-Faeroe Ridge are the zero (bold) and +200 nT (dashed) contours from the magnetic survey of Fleischer et al. (1974). Grid lines are omitted over regions of negative anomaly. The base map is the 1:4 000 000 Mercator projection bathymetric chart of Gronlie and Talwani (1978). The 500 and 1000 nominal fathom contours are shown dotted.

Figure 6.6. As above with anomaly identification according to the present study. Other sources which were referenced to aid anomaly identification are discussed in the text. Fracture zones are shown dashed, in bold and the Faeroe-Shetland and Voring Plateau Escarpments of Talwani and Eldholm (1977) are also shown (FSE and VPE). The axial anomaly over the present ridge system and over the Aegir axis is indicated as A.

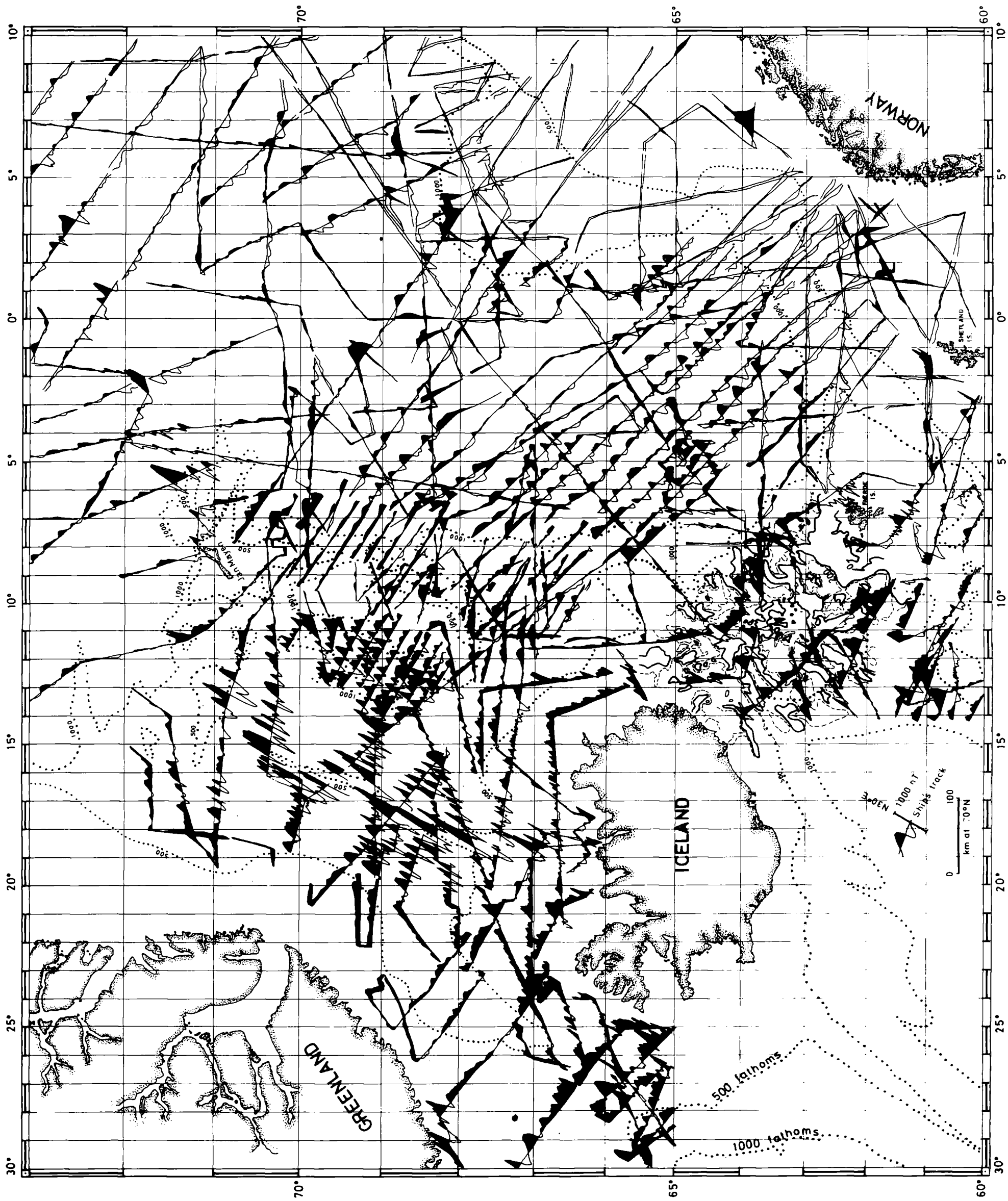


Figure 6.5

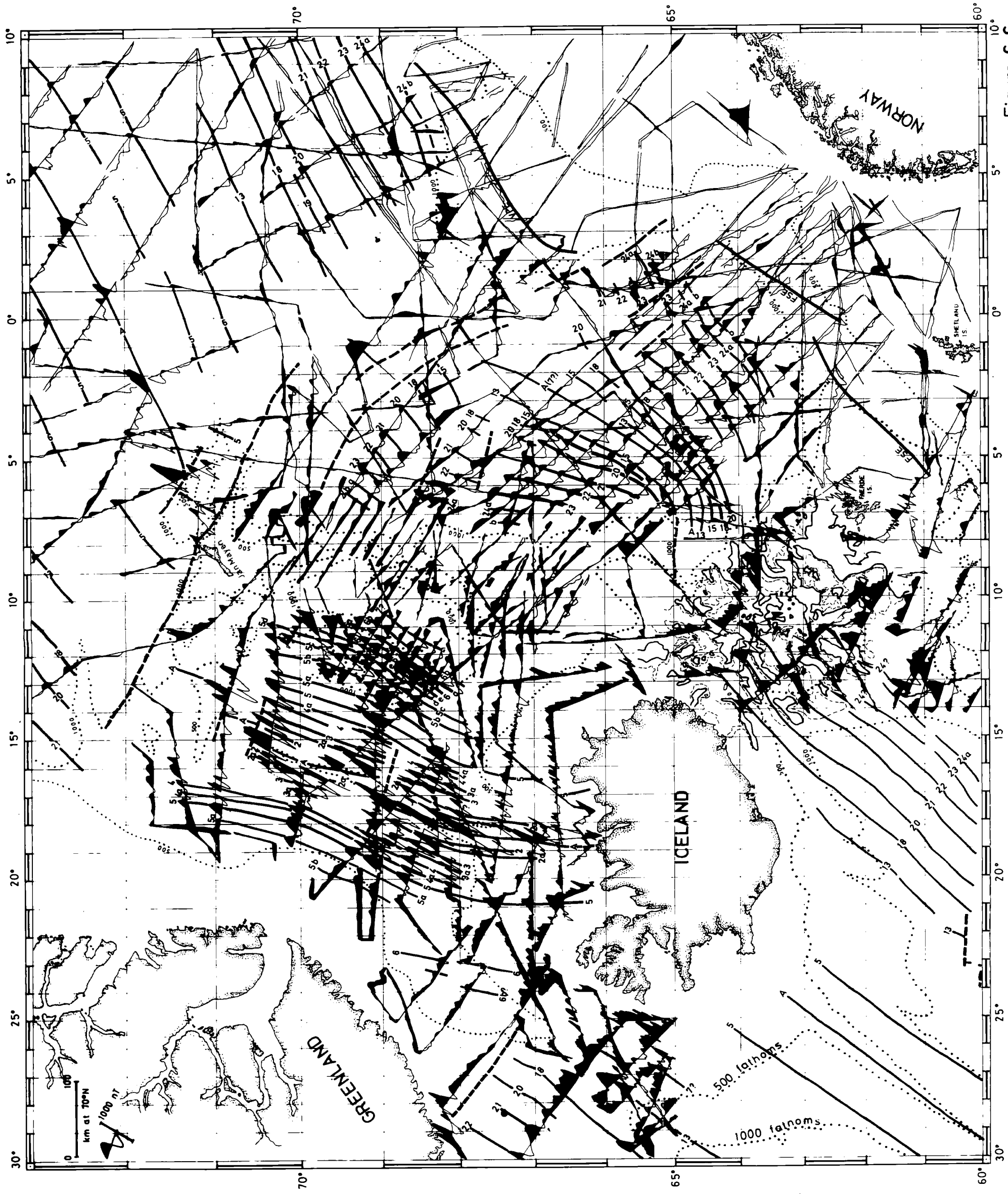


Figure 6.6

former case the information has been requested but has not yet arrived). In the vicinity of the Jan Mayen Ridge it is evident that the regional field is too small as the anomalies are consistently positive.

6.4 Anomaly correlation and identification in the southern Norwegian-Greenland Sea and surrounding areas

Figure 6.6 shows the trends of identified magnetic anomalies in the region 60°N - 70°N , 30°W - 10°E . Most of the effort in the present study was aimed at the correlation and identification of anomalies in the Norway Basin. Outside of this area the identifications have been based on other studies. The anomaly patterns in the regions outside of the Norway Basin will be briefly described before a more detailed discussion of the Norway Basin anomalies is given.

The lineations about the Reykjanes Ridge were copied from Figures 3 and 5 of Voppel et al. (1979). They consider that anomaly 24 is repeated between 61°N and 62°N (Section 6.1). The exact identification of anomaly 24 in this region appears to be in doubt.

The possible continuation of anomaly 21 over the Iceland-Faeroe Ridge is also after Voppel et al. Tentative extensions of anomalies 20 and 18 are also shown. The magnetic anomalies over the Iceland-Faeroe Ridge are very confused. All of the following factors probably contribute to the irregularity of the anomaly pattern

- (i) the shallow magnetic source depth
- (ii) the subaerial eruption (Nilsen and Kerr, 1978; Gronlie, 1979) and extensive lateral flow of lavas forming the magnetic basement (as in Iceland)
- (iii) the frequent occurrence of central volcanoes and intrusive bodies (Ingles, 1971; Fleischer et al., 1974).

There is also a possibility of disturbance caused by local shifts of spreading axis. However, in contrast to previous authors (Bott, 1974; Fleischer et al., 1974; Talwani and Eldholm, 1977), Voppel et al. (1979) consider that the Icelandic transverse ridge has, by and large, formed in a parallel fashion to the area further south. Thus they

consider that the oldest anomalies are probably duplicated west of the Faeroes Islands, but that since anomaly 21 time or slightly earlier the transverse ridge has developed more or less symmetrically.

In this context the tectonic configuration of Iceland should be mentioned. The axial rift zones (Saemundsson, 1974, 1978), where crustal accretion is presently taking place, are confined to the neovolcanic zones, which are shown in Figure 1.2. Saemundsson (1974) has proposed the following tectonic history, mainly on the basis of geological evidence. Prior to about 4 Ma before present, the spreading axis occupied the western neovolcanic zone (WVZ) and extended further north to join the Kolbeinsey Ridge without offset. At about 4 Ma, rifting in the north shifted east to the northern neovolcanic zone (NVZ), forming the Tjornes Fracture Zone (Section 1.4) and an east-west transform zone in central Iceland. At about 1.5 Ma rifting in the southern part of Iceland began in the eastern neovolcanic zone (EVZ) and a transform boundary formed in the south of Iceland, extending eastwards from the Reykjanes peninsula. This boundary is marked by the south Iceland seismic zone (Einarsson (1979); see the epicentre distribution and fault plane solution in Figure 1.2). The eastern neovolcanic zone is still active, but in effect the entire axis system has shifted eastward by about 100 km since 4.5 Ma before present. The major transform faults to the north and south of Iceland are thus quite young, although it appears that there was a sharp dextral bend in the axis, coincident with the Reykjanes peninsula, even prior to the axis shift (Saemundsson, 1974).

In view of the irregular nature of the anomaly pattern over the Iceland-Faeroe Ridge it is not possible to be sure that similar shifts of axis have not taken place between anomaly 22 and 4 Ma before present. However the hypothesis of Voppel et al. (1979) is simple and fits in well with the evolutionary scheme proposed in Chapter 7.

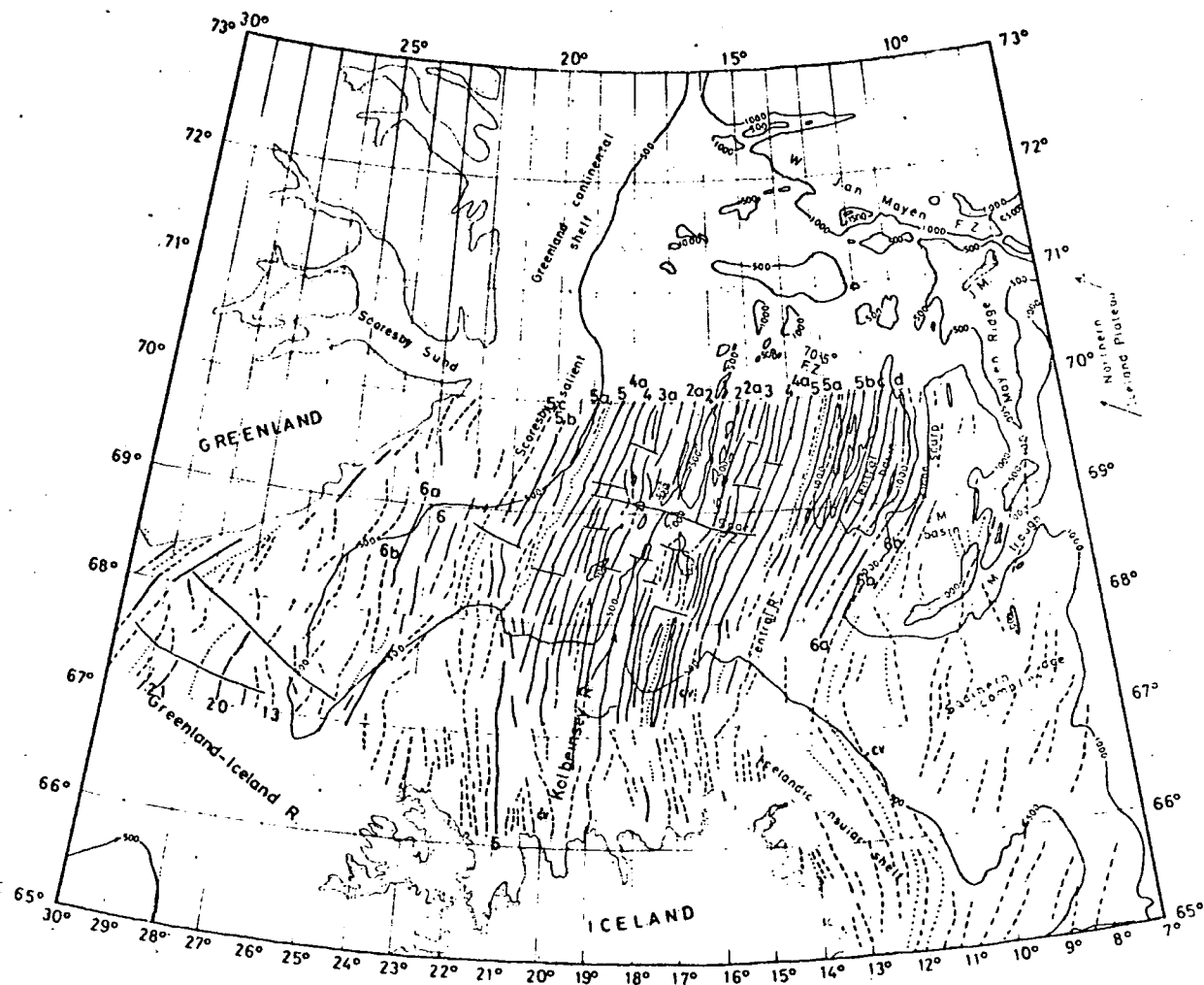
The magnetic anomaly pattern over the Greenland-Iceland Ridge is also confused (although not as confused as it appears to be in Figure 6.6). The lineations shown are based

on those of Vogt et al. (1980) who used the earlier aeromagnetic data of Johnson et al. (1975a). Although they are irregularly shaped, the anomalies over the Denmark Strait appear to be in sequence, as they are further south in the West Reykjanes Basin. The youngest anomaly identified by Voppel et al. (1979) is 13, but anomaly 7 may be recognisable in places. South of the major fracture zone just north of the Denmark Straits the detailed aeromagnetic survey of Vogt et al. (1980), (which terminates at 67°N) shows well developed anomalies from 22 through to 13 (and possibly younger). A chart showing their lineations is given in Figure 6.7. It is thought that the more southerly fracture zone which is shown in Figure 6.7 (but not in Figure 6.6) probably does not offset the anomalies appreciably. The recognition of anomaly 22 at the western edge of the aeromagnetic survey area is important as this anomaly had not previously been identified so far north. It confirms that the westward shift of axis must have taken place prior to anomaly 22 time as suggested by Featherstone et al. (1977). Magnetic anomaly profiles in this area have also been presented by Kristjansson (1976), who did not identify any magnetic lineations but noted that a number of central volcanic complexes were probably present.

The fracture zone shown in Figure 6.6, just north of the Denmark Straits was postulated by Vogt et al. (1980), on the basis that magnetic anomaly lineations terminate against it. Its position and orientation fit in well with the evolutionary hypothesis proposed in Chapter 7.

North of the fracture zone, over the Iceland Plateau, the lineations shown in Figure 6.6 are again based mainly on the results of Vogt et al. (1980). In order to augment the data shown in Figure 6.6, which is insufficient to support some of the lineations which are indicated, the marine profiles presented by Mirlin and Melikhov (1976) were redrawn at the same scale. However they were not included as their positions were doubtful in some places because only two grid lines were drawn on the original figure.

Adjacent to the Greenland margin, west of anomaly 5, the magnetic anomaly lineations are weak. The oldest anomaly



Magnetic anomaly lineations
over the Iceland Plateau after
Vogt et al. (1980)

- / Well defined positive anomaly correlation
- \\ Less certain positive anomaly correlation
- Strong negative anomaly correlation

POLAR-STEREOGRAPHIC PROJECTION
Depth in nominal fathoms after
Grönlie and Taiwani (1978)
Scale at 70°N
0 100 200 300
km

Figure 6.7

identified by Vogt et al. (1980) is 6B. Between this anomaly and the band of very short wavelength anomalies which lies close to the Blosseville Coast (not shown), there is a wedge-shaped area containing weakly lineated anomalies (Figure 6.7). It is conceivable that anomalies 7 to 18 may be present in this region, with diminished amplitudes (Chapter 7).

On the other side of axis, the identifications for anomalies older than 5 are also the same as are given by Vogt et al. Over the central Iceland Plateau the R.V. Akademik Kurchatov data had to be digitised by hand and the pattern of lineations is not as clearly revealed as it is by the aeromagnetic survey. No anomalies older than 5 have been identified south and east of about 67°N and 16°W . As shown in Figure 6.7 the lineations in this region swing around to become sub-parallel to the Icelandic shelf break.

The identification of anomalies 5 and younger about the Kolbeinsey axis is straightforward if the Vogt et al. (1980) survey is used for comparison. North of the Spar fracture zone anomaly 3A is missing because of the eastward shift of the Kolbeinsey ridge, as first noted by Meyer et al. (1972). As suggested in Chapter 1, the ridge is shown to have two dextral offsets north of 70°N . This is supported by the two lines of Mirlin and Melikhov (1976), which cross the axis at 69.85°N and 71.55°N . Unfortunately there is not enough information to identify any anomalies other than the axial anomaly over the northern Iceland Plateau zone. Mirlin and Melikhov's most northerly line, which runs parallel and 25 km south of the western Jan Mayen Fracture Zone, does not show recognisable anomalies older than about 2A.

In the southern part of the central Norwegian-Greenland Sea, the anomaly identifications are based on those in Figure 6.3 (Vogt et al., 1978; Phillips et al., in press). The anomaly trends shown in Figure 6.5 correspond precisely with those given by Talwani and Eldholm (1977) and Gronlie and Talwani (1978), with few exceptions. The dextral offset and change in strike of the Mohs Ridge about 100 km north of the western Jan Mayen Fracture Zone is an interesting feature

which has been noted before (Vogt et al., 1978). It is also interesting to note that anomaly lineations do not appear to continue south of the spur-like ridge (Section 1.10) at 72.4°N (15°W to 11°W).

Figure 6.8 shows the proposed scheme of anomaly correlation and identification in the Norway Basin, presented in the form of a "zebra stripe" chart. There does not appear to be a great deal of published marine anomaly data in excess of that which is presented in Figure 6.8. Initially it was hoped to make considerable use of the contour charts for the aeromagnetic survey of Avery et al. (1968), which cover all of the Norway Basin. However in a number of instances the contoured data, even close to flight lines, was found to be in serious disagreement with the marine profiles. This disagreement was attributed to navigation errors in the aeromagnetic survey, which was flown in 1959.

Figure 6.9 shows the anomaly identifications of Talwani and Eldholm (1977), as presented in the Geophysical Atlas by Gronlie and Talwani (1978). There are a number of major differences between their interpretation and that which is presented here, but the most important is that, for the most part, on both sides of the Aegir axis the anomaly numbers given by Talwani and Eldholm are one lower than those proposed in the present study.

The proposed anomaly identification can be justified firstly by considering the anomaly pattern in the eastern Norway Basin, south of about 65°N , where there are no problems correlating the anomalies. In Figure 6.10 the magnetic anomaly pattern in this region is compared directly with a portion of Vogt and Avery's (1974) aeromagnetic contour chart of the East Reykjanes Basin (Figure 6.2). If the anomaly identification proposed in this study is adopted there is a good correspondence between the principal features of these sequences. In particular there are five well-developed negative anomaly bands to the east of anomaly 18, of which the widest lies between anomalies 20 and 21. Anomaly 19, which is discontinuously developed in the East Reykjanes Basin is also perhaps evident in the Norway Basin

Figure 6.8. Magnetic anomalies in the Norway Basin (as in Figure 6.5) showing correlations and identifications according to the present study. Also shown are fracture zones (dashed) and the Voring Plateau and Faeroe-Shetland Escarpments (solid). NN' is the Norway Basin profile (the first 70 km of the profile, along which the magnetic anomaly is small, is not shown in this diagram).

Figure 6.9. Magnetic anomalies in the Norway Basin, as above, showing the correlations and identifications of Talwani and Eldholm (1977), traced directly from the Geophysical Atlas (Gronlie and Talwani 1978) and also from Gronlie and Talwani (1979), in the northwestern region.

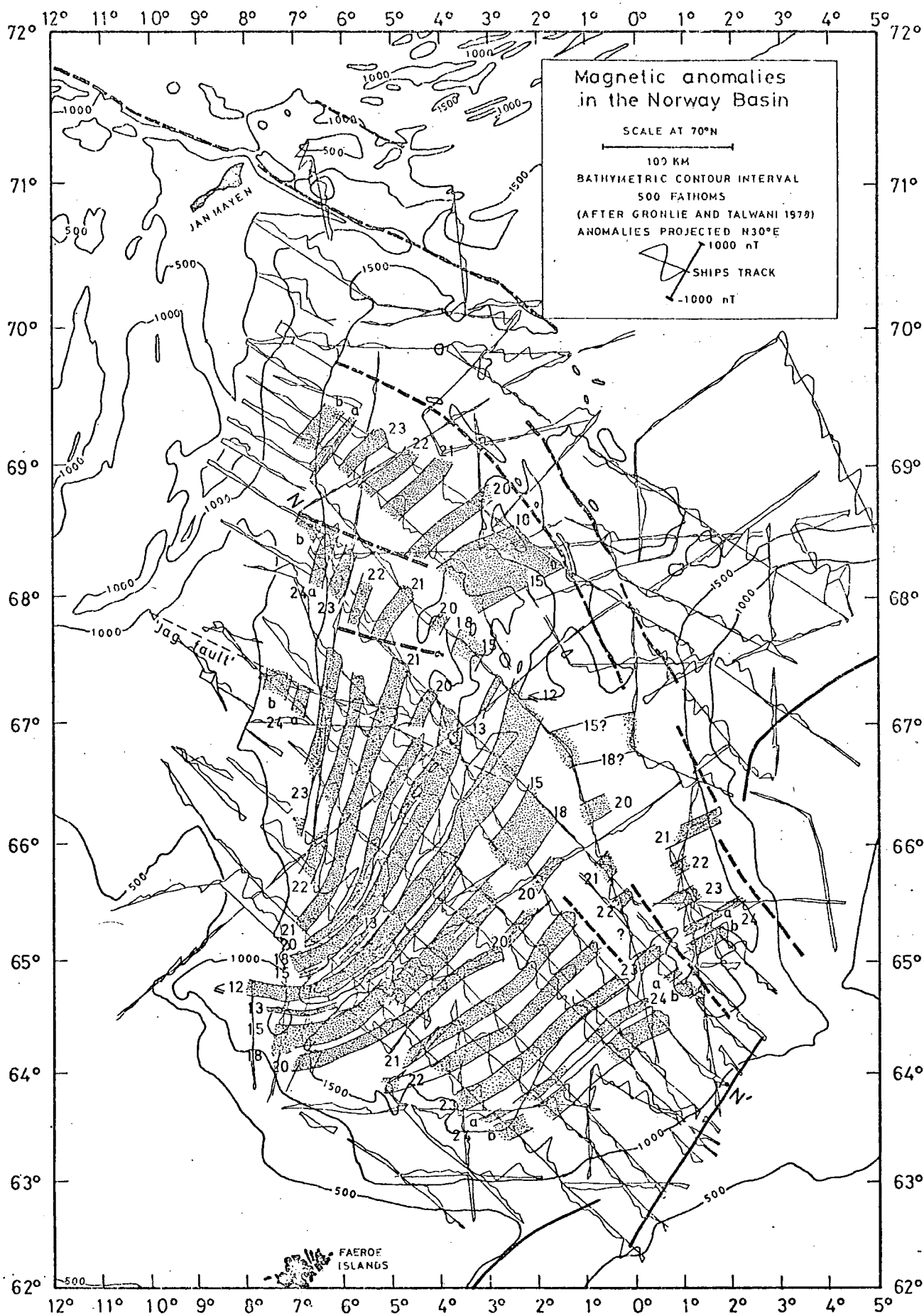


Figure 6.8

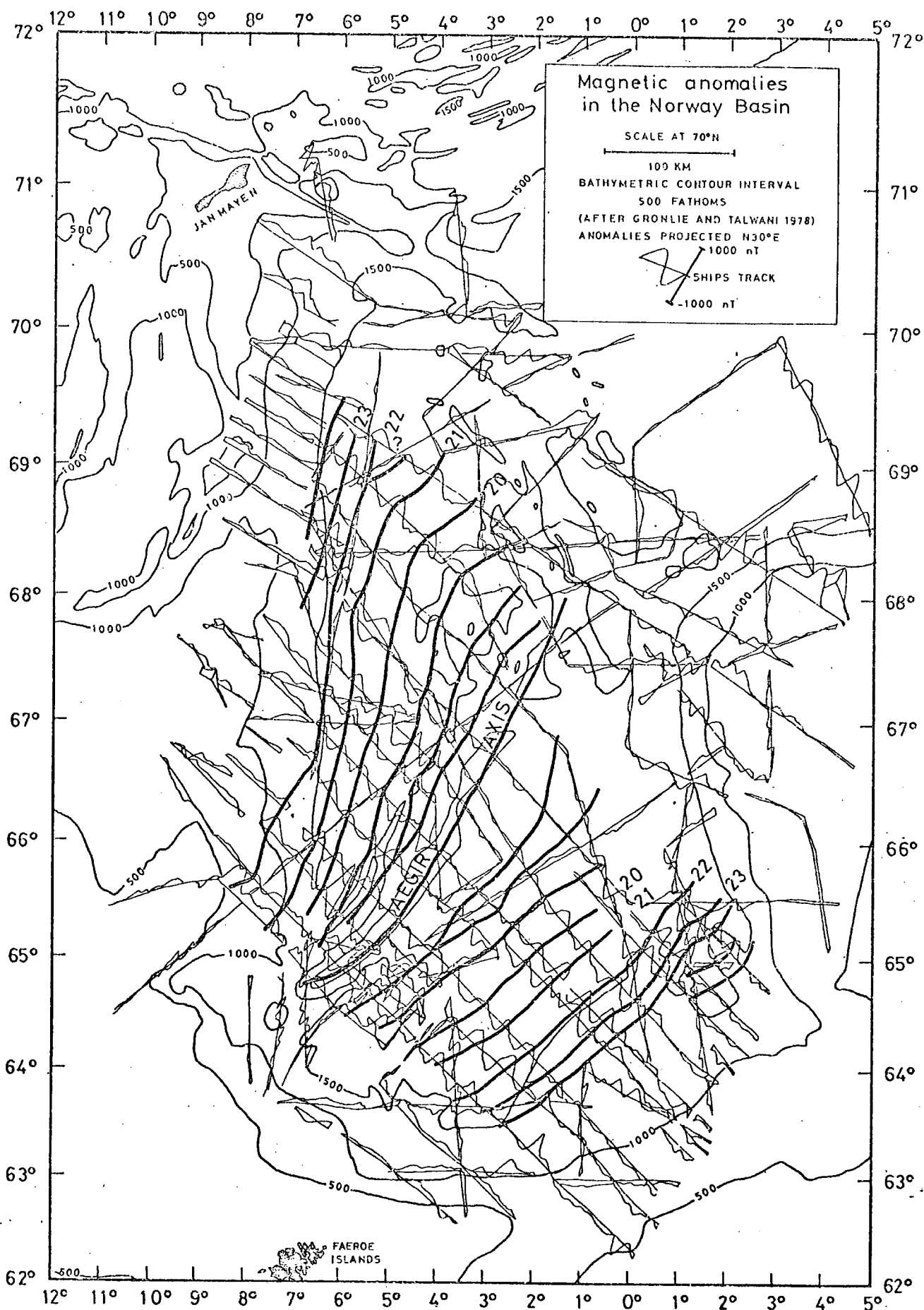


Figure 6.9

at 65.3°N . The form of anomaly 24, according to the present identification, is also similar to that observed elsewhere in the northeast Atlantic (Section 6.1). Anomaly 24A in the Norway Basin is weak and irregular while anomaly 24B is generally broader and stronger, as in the East Reykjanes Basin (with the identification of Figure 6.2). In both the eastern Norway Basin and the East Reykjanes Basin anomaly 24B is bordered to the east by a broad magnetic low which is interrupted by positive anomalies of limited extent.

If Talwani and Eldholm's identification is adopted there is a generally poorer correspondence between the two sequences as the broadest negative anomaly in the Norway Basin lies west of anomaly 20, there is no sign of anomaly 19 and most importantly, there is another well developed anomaly further west, lying within the 15-18 sequence, which has no counterpart in the East Reykjanes Basin. Also, anomaly 23 (which is anomaly 24A according to the present interpretation) is bordered closely by a further positive anomaly, which is not explicitly identified. Talwani and Eldholm consider that the ocean-continent boundary in this region is coincident with the Faeroe-Shetland Escarpment and that prior to a westward axis shift at anomaly 23 time, spreading took place between the escarpment and their anomaly 23. According to this interpretation, the linear anomaly to the east of anomaly 23 might be anomaly 24B, formed on the westward side of the shortlived axis. However there is no indication of an eastern anomaly 24 in this case. An alternative interpretation based on Talwani and Eldholm's identification of anomaly 23 is that the shortlived axis did not exist and that the anomaly to the east of 23 is anomaly 24A and 24B combined (J. Robinson personal communication). This implies a low initial spreading rate.

Figure 6.11 shows a direct comparison of the two anomaly identification schemes along the eastern part of the SHACK977 Norway Basin profile (NN' in Figure 6.8). For each identification, spreading curves have been obtained by plotting the crustal polarity profile obtained by direct inversion against the polarity timescale of Hailwood et al. (1979). The spreading curve corresponding to the proposed

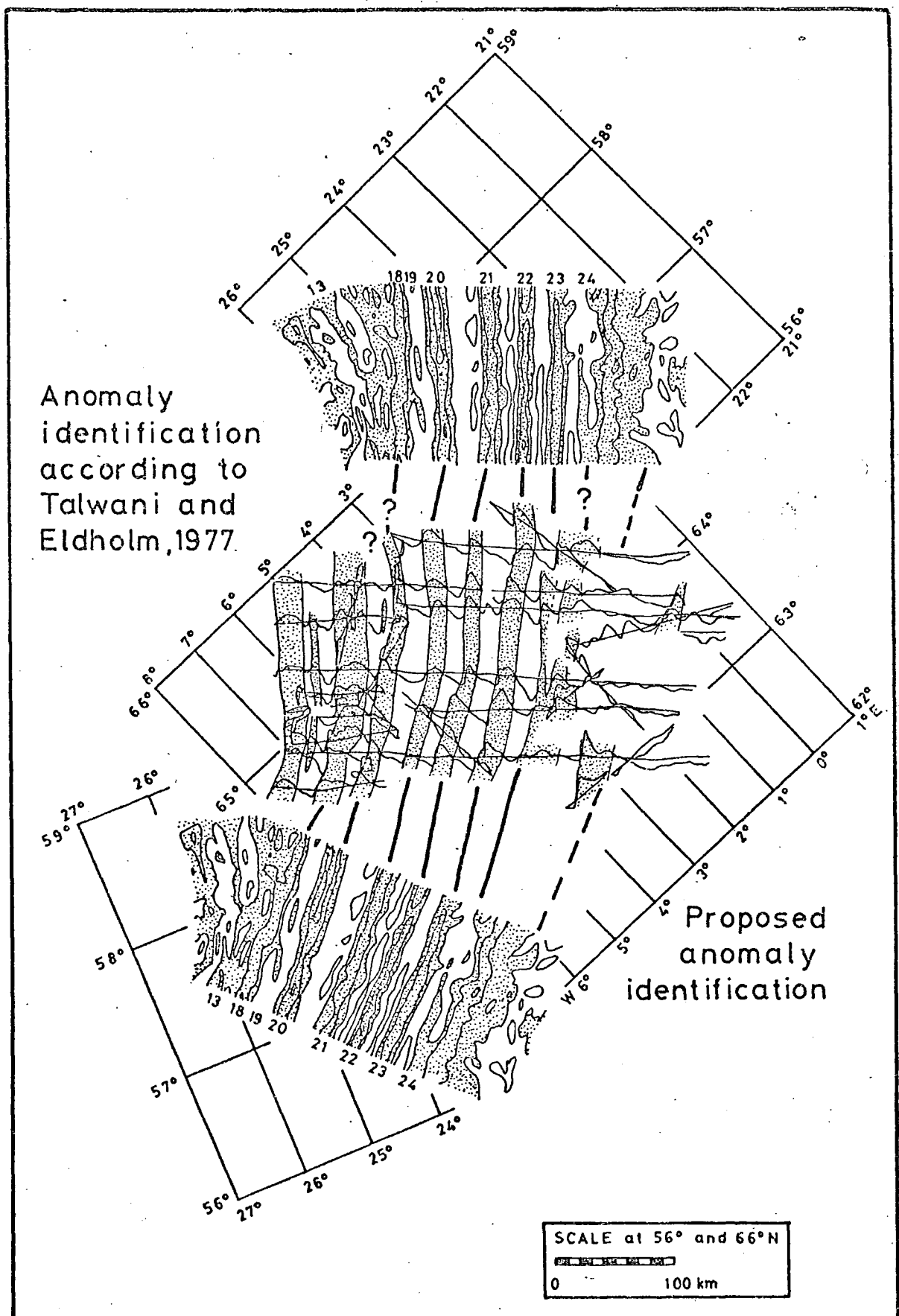
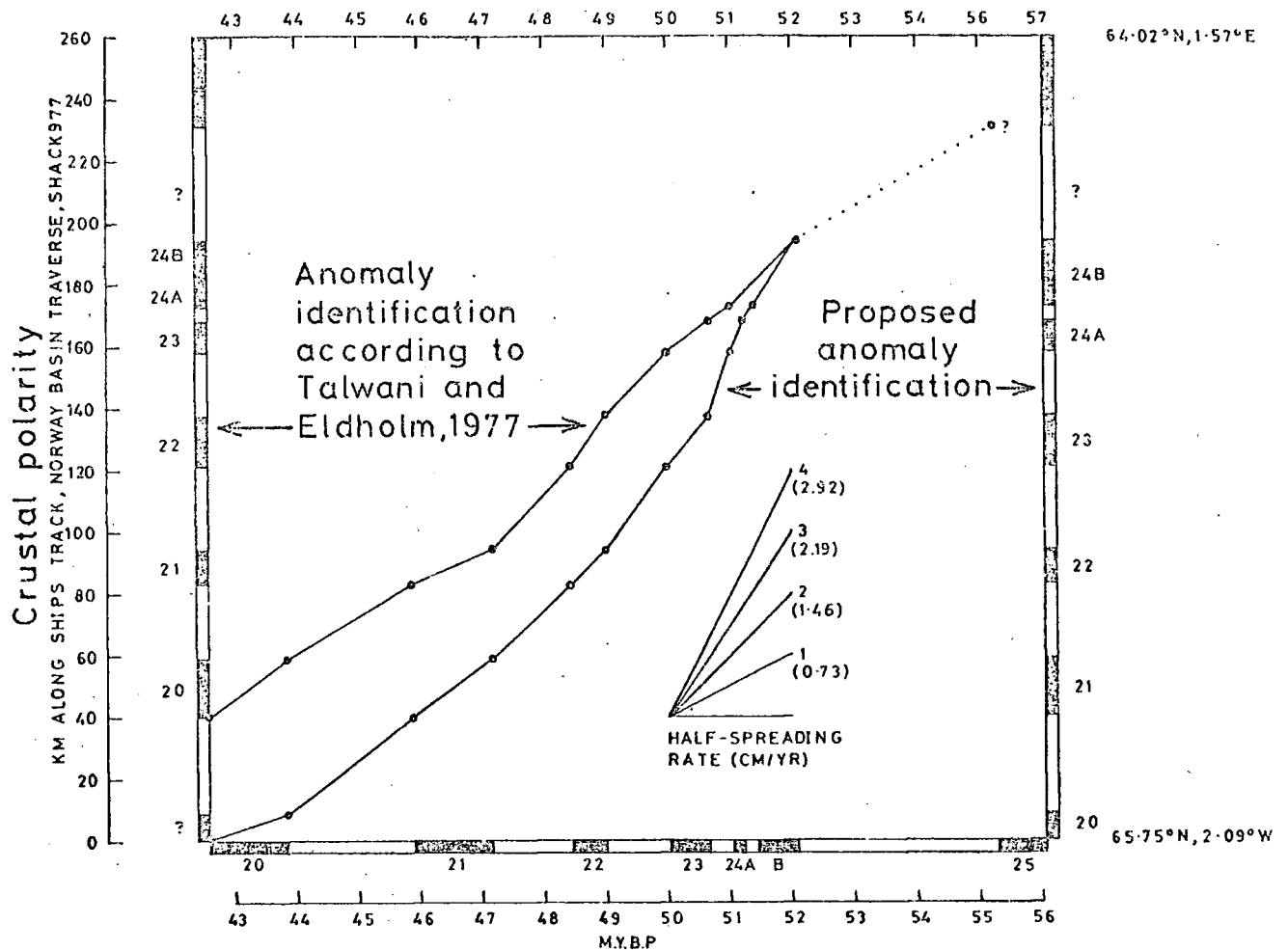


Figure 6.10. Comparison of the anomaly sequence in the eastern Norway Basin with the sequence in the East Reykjanes basin (Vogt and Avery 1974, as in Figure 6.2), illustrating the difference between the identification scheme of Talwani and Eldholm (1977) and that of the present study.



Polarity timescale of Hailwood et al., 1979

Figure 6.11. Eastern Norway Basin spreading curves for the anomaly identifications of Talwani and Eldholm (1977) and this study. The magnetic polarity along ship's track for the Norway Basin traverse (SHACK977) was derived using the inversion techniques described in Chapter 5. Inversion parameters are as in Figure 5.4 (inclination of magnetisation with respect to ship's track is 120°) and distances are relative to the western edge of anomaly 20 crust (454 km from beginning of traverse). The gradient scale shows half spreading rates relative to the Hailwood et al. (1979) polarity timescale, which for this time period is contracted by a factor of 0.73 relative to the Heirtzler et al. (1968) timescale. Corresponding half spreading rates for the latter timescale are shown in brackets.

identification shows that the initial spreading rate which is required to explain the separation of anomalies 24A and 24B is also consistent with the increased width of the negative anomaly which flanks anomaly 23. The half spreading rate decreases almost monotonically from about 3-4 cm/a at initial opening to less than 1 cm/a at anomaly 20 time. Naturally the quoted spreading rates should be treated with caution as they are based on a single line. Taking into account the different timescale used, the spreading rates deduced by Vogt and Avery (1974, figure 8) between the Azores and Iceland are similar although their initial rate is not quite so high.

The curve corresponding to Talwani and Eldholm's identification scheme has been continued under the assumption that the anomaly to the east of their anomaly 23 is comprised of anomalies 24A and 24B. The spreading curve is perhaps less smooth than the one corresponding to the proposed scheme, with half rates varying between about 1 and 3 cm/a. However it is clear that there is no a priori way of choosing between the spreading curves without a more detailed knowledge of spreading rate variations than is presently available. The direct comparison of anomaly sequences, as described above, is considered a more reliable identification technique, as the explicit consideration of spreading rates is avoided.

The proposed scheme allows a consistent numbering of the anomalies in the central Norway Basin. The central positive anomaly, over the Aegir axis, is thought to correspond to the band 7 to 12. Spreading must have ceased by the end of anomaly 7 time at least, otherwise the prominent negative anomaly between 6C and 7 would be expected to be present. On either side of the central positive anomaly there is a negative anomaly within which anomaly 13 is sporadically developed. This is flanked by the predominantly positive band, 15 to 18. Note that in the central zone lineations have been extended around into the region of the bathymetric high between the Jan Mayen block and the Iceland-Faeroe Ridge. This was done with confidence after an intensive study of the anomaly profiles and a consideration of a number of different possibilities for correlation.

Talwani and Eldholm (1977) pointed out that the central zone anomaly pattern is fan-shaped. This is supported by the anomaly correlations of the present study. The fan shaped lineation pattern is in fact more obvious in Figure 6.8 (present study) than it is in Figure 6.9 (Talwani and Eldholm). According to the present interpretation "fan-shaped spreading" began immediately after anomaly 20 time. In the next chapter a kinematic description of the "fan shaped spreading" will be given in terms of rotation about a near pole, as suggested by Talwani and Eldholm (1977).

The extension of the identification of anomalies 20 to 24 to the western side of the Aegir axis is straightforward, because of the symmetry of most profiles about the axis. South of the proposed fracture zone at 67.7°N the numbers assigned to the anomalies are again one greater than those of Talwani and Eldholm. As on the eastern side of the axis, the broadest negative anomaly lies between anomalies 20 to 21. South of 67°N , anomalies 24, 23, 22 and 21 progressively disappear as coherent anomalies. It is reasonable to infer that these anomalies were originally present but have been subsequently erased, as the corresponding anomalies are normally developed on the eastern side of the axis.

The thin dashed line in Figure 6.7 shows the trend of the "JAG" transform fault proposed by Gairaud et al. (1978). In the vicinity of the proposed fault there is a close network of profiles, across which the correlations are unambiguous, precluding the possibility of a 40 km sinistral offset in the anomaly pattern. The magnetic anomaly data therefore support the conclusion (Sections 1.3 and 1.8) that the JAG fault does not exist.

About 70 km further north a small fracture zone offsetting the anomalies sinistrally is proposed on the basis of the anomaly evidence. The anomaly identifications immediately north of the proposed fracture zone are based on the recognition of anomaly 24, the broad negative anomaly between anomalies 20 and 21 and the anomaly 18 to 15 band. There is a small but definite sinistral offset which previous

authors have accommodated by a bend in the lineation pattern. That the offset occurs relatively sharply is suggested by the "stretching" of anomaly 21 along the Norway Basin profile, which crosses it at an acute angle at 67.5°N . When the long wavelength components are removed from the anomaly recorded along the Norway Basin profile (as in Figure 6.12) anomaly 21 resolves into two separate peaks. It is therefore considered likely that anomaly 21 is actually duplicated along the Norway Basin profile, by sinistral offset along the proposed fracture zone. Figure 6.12 shows that the following changes occur across the proposed fracture zone, passing from northwest to southwest;

- (i) the basement becomes conspicuously more rugged
- (ii) the average sediment thickness decreases
- (iii) the average free air gravity anomaly increases.

While these points support the idea of a crustal discontinuity, the sediment thickness and gravity anomaly changes in a contrary manner to that which would be expected, as it is proposed that the crust to the south of the fault is older than it would be if the fault were absent. However if sediment transport was in a southerly direction these changes may be explained by damming of sediment against a basement high which was related to the proposed fracture zone.

Although many other possibilities were explored no better explanation could be found for the anomaly pattern in this region. It was initially thought that this fracture zone might correspond to the JAG fault of Gairaud et al. (1978). However if it is continued to the Jan Mayen block with any reasonable trend it crosses profile CP105 (Figures 1.8 and 1.10) at a latitude where there is no evidence of faulting.

Another fracture zone is proposed at about 68.5°N . North of this fracture zone the anomalies correlate in a northeasterly direction. There can be no doubt about the anomaly identifications in this region, which are the same as those of Gronlie et al. (1978), who have presented some profiles which are not shown in Figure 6.8. The stippled zones corresponding to anomalies 24A and 24B are in

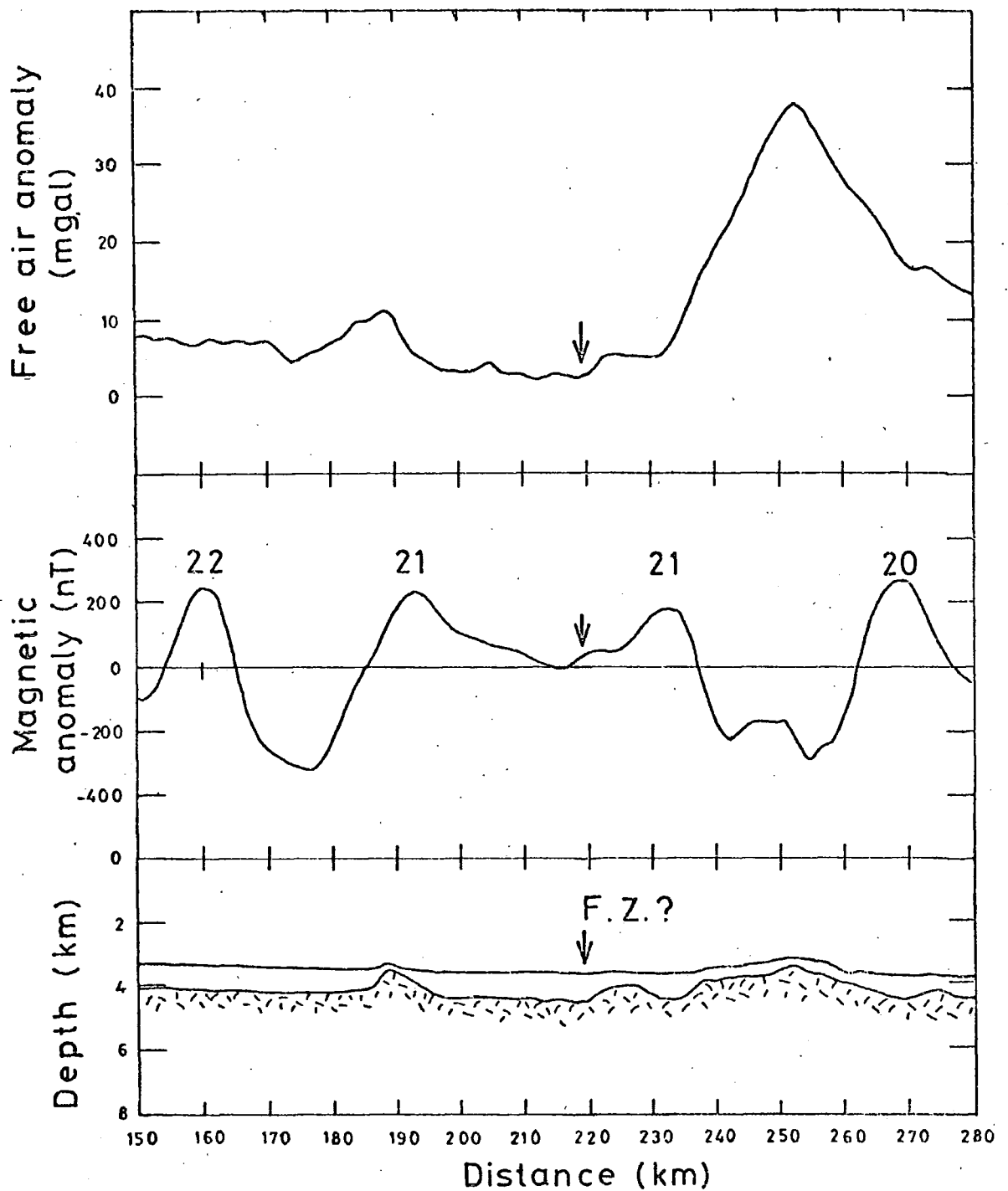


Figure 6.12. Gravity and magnetic anomalies and depths to the sea bed and acoustic basement along the portion of the Norway Basin profile which crosses the proposed fracture zone at 67.5°N (Figure 6.8). Note that the magnetic anomaly is filtered to remove wavelengths longer than 125 km as in the magnetic inversions of Section 5.2. The anomalies in the magnetic charts and Appendix A are not filtered.

contradiction with the most northerly profile which they cross in Figure 6.8, but are in agreement with a virtually coincident profile illustrated by Gronlie et al. (RK10, figures 11 and 12). The existence of a fracture zone with a substantial sinistral offset is demanded if the anomaly identifications in Figure 6.8 are accepted, but the precise position and trend of the fracture zone is less certain. If the fracture zone is positioned as shown, it crosses the Norway Basin profile (which is not shown in its entirety in Figure 6.8) at about 70 km along track. As noted in Chapter 3, there is no evidence for a fracture zone in this vicinity so the situation may be somewhat different.

On the eastern side of the Norway Basin two fracture zones are also proposed south of the eastern Jan Mayen Fracture Zone. The anomalies in this vicinity are difficult to identify but it is believed that anomaly 24 identifications are correct. The more northerly fracture zone crosses disturbed anomalies and is approximately coincident with a number of bathymetric features of similar trend (Figure 1.4).

6.5 Ocean-continent boundaries in the southern Norwegian-Greenland Sea

This section is mainly concerned with the ocean-continent boundaries of the Jan Mayen block. Figure 6.13 is a compilation of seismic (Chapters 1 and 3), gravity (Chapter 4) and magnetic anomaly data (this Chapter) bearing on the location of the ocean-continent boundaries of the Jan Mayen block.

The striped lines represent the estimated ocean-continent boundaries based on the seismic data of Figures 1.9, 1.15, 3.1 and 3.2. In general the extent of continental crust was conservatively estimated in the instances where doubt exists, under the assumption that there are not large areas of continental crust hidden beneath opaque acoustic basement in the western and eastern margins of the block. Thus the western boundary was placed immediately outside the most westerly line of transparent

peaks which penetrate the opaque layer of the Jan Mayen basin (Section 1.9). This line diverges markedly from the anomaly 6B lineation, going towards the south.

On the eastern margin the positions of anomalies 24A and 24B (from Figure 6.8) are indicated as "zebra stripes". Also shown (in stipple) are the transition zones across which the Moho appears to deepen by about 4 km passing from the Norway Basin to the Jan Mayen block (modelled directly for profiles 1, 2 and the Norway Basin profile, and inferred for profiles 6 and 8; Sections 4.4 and 4.5).

All of these estimates of the eastern ocean-continent boundary are reasonably concordant. Anomaly 24B always lies to the east of the acoustic basement transition and with the exception of the Norway Basin profile so do the Moho transition zones. The summarised data in Figure 6.13 indicate that the Moho transition zone is probably underlain by thickened oceanic crust. They are also consistent with an onset of spreading shortly before anomaly 24B time. The apparent offsets in the acoustic basement boundary agree quite well in position with the proposed fracture zones in the Norway Basin.

Figure 6.14 compares the acoustic basement transition boundaries of Figure 6.13 with previous estimates of the ocean-continent boundaries of the Jan Mayen block. Talwani and Eldholm (1977) subdivided the Jan Mayen block into three regions (1, 2 and 3, outlined in bold in Figure 6.14). They considered regions 1 and 2 to be underlain by continental crust while they suggested that region 3 (comprising ridges 1 to 3) was the zone of compensatory fan shaped spreading. However the results of Gairaud et al. (1978) shows fairly conclusively that ridge 1 is underlain at depth by Horizon 0. Moreover the general tectonic style within Talwani and Eldholm's region 3 is similar to that within region 2 and is not typical of oceanic regions. Further north, Talwani and Eldholm's eastern boundary is in general agreement with that suggested in the present study. Their western boundary however is parallel to the magnetic lineations over the central Iceland Plateau and diverges from the proposed

Figure 6.13. Definition of the western and eastern boundaries of the Jan Mayen block.

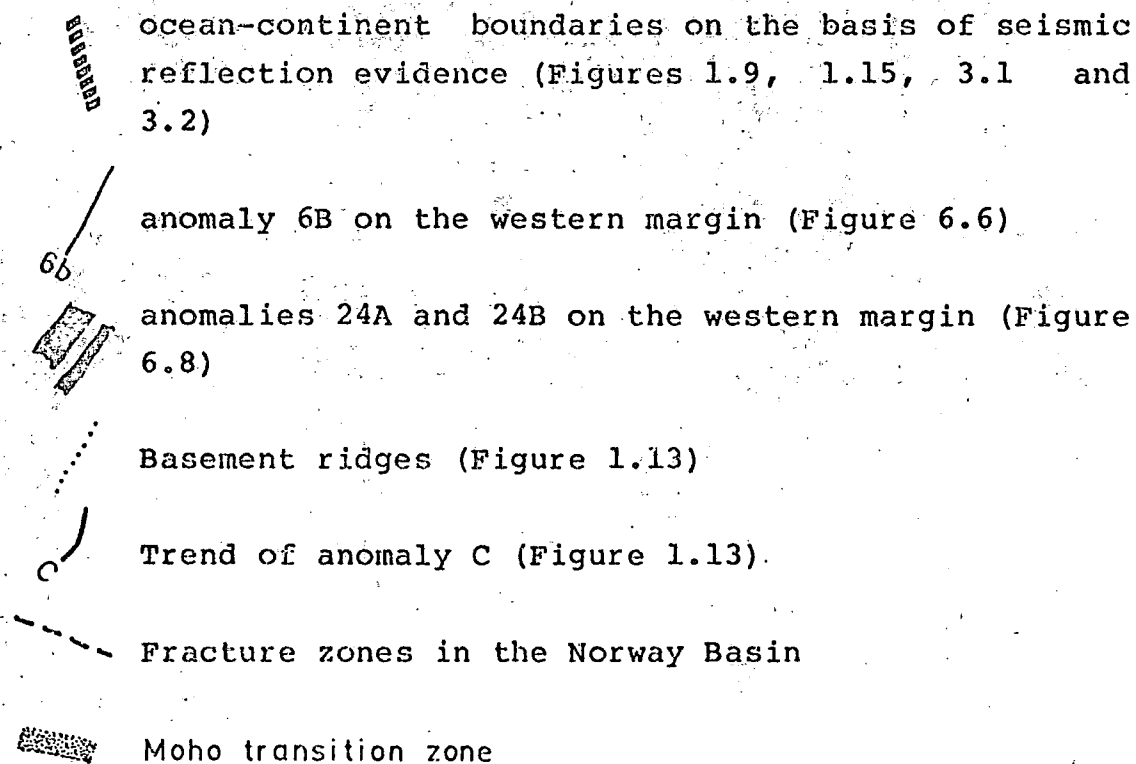
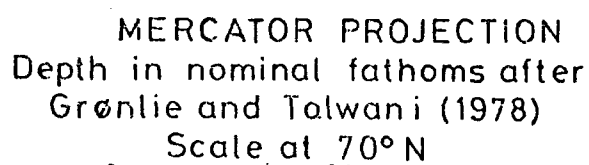
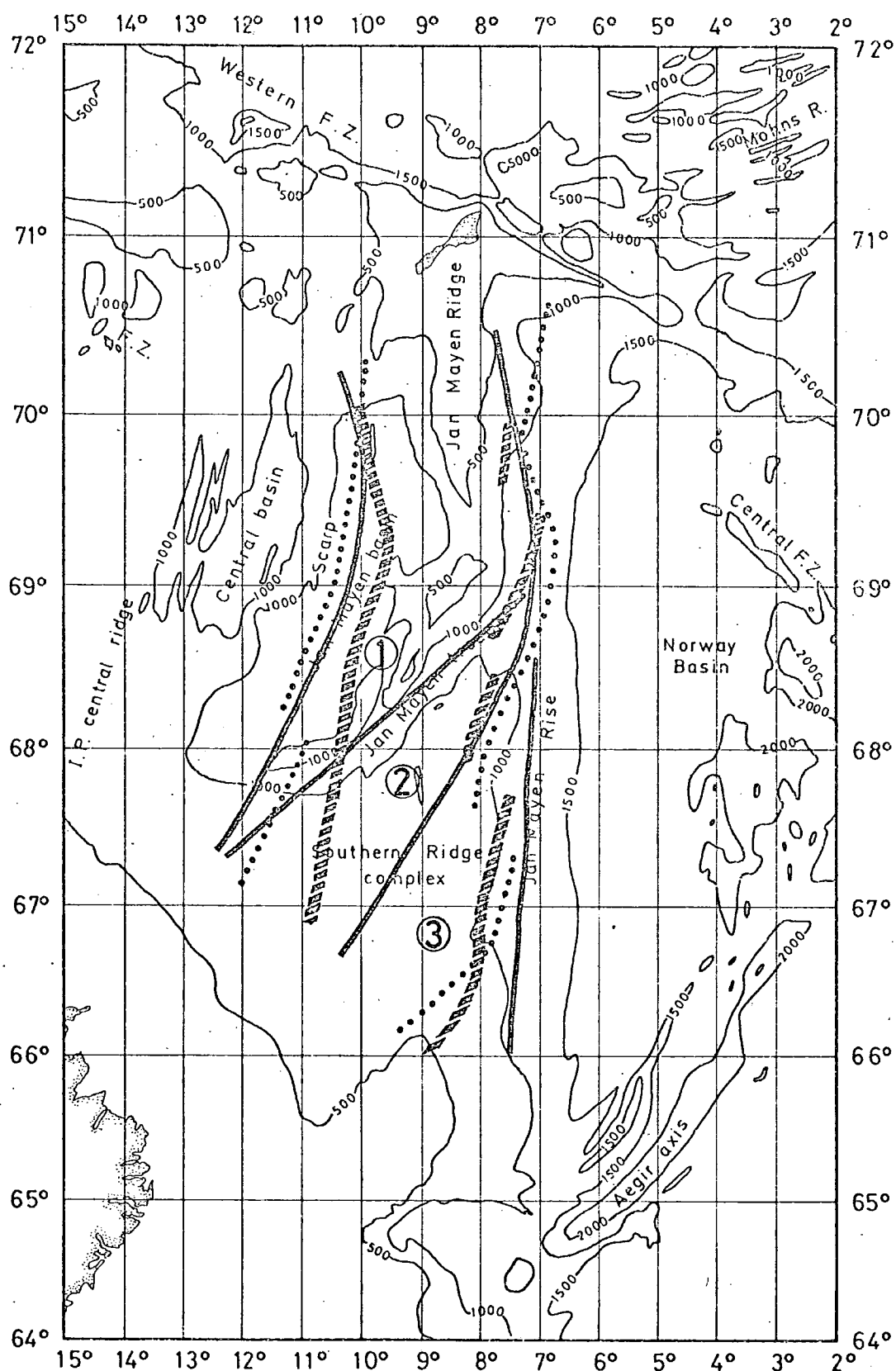


Figure 6.14. Comparison of proposed ocean-continent boundaries (Figure 6.13) with the ocean-continent boundaries of Gairaud et al. (1978), which are dotted and the crustal regions of Talwani and Eldholm (1977), which are outlined in bold. Talwani and Eldholm consider regions 1 and 2 to be continental and region 3 to be oceanic.



Scale bar for 75 N, showing 0, 100, and 200 km.



MERCATOR PROJECTION
 Depth in nominal fathoms after
 Grønlie and Talwani (1978)
 Scale at 70°N

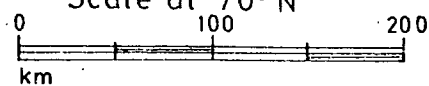


Figure 6.14

boundary. The western boundary of Gairaud et al. (1978) also follows the western margin of the Jan Mayen basin, rather than the more easterly line of "transparent" peaks. The proposed eastern boundary is fairly close to that of Gairaud et al., which is not surprising as a similar approach and much the same data were used, although in general they have estimated the extent of continental crust more liberally.

Judging from the data presented in this study, the eastern ocean-continent boundary of the Norway Basin is located just east of anomaly 24B on the Norway margin (Figure 6.8). This position is consistent with the change in reflection characteristics along the Norway Basin profile (Figure 3.2) and the position of the Moho transition zone in the gravity model. The same conclusion was previously reached by Robinson (1980) on the basis of his detailed study of the SHACK476 profiles. This ocean-continent boundary is located nearly 100 km west of the Faeroe-Shetland escarpment, which Talwani and Eldholm (1972) considered to be the ocean-continent boundary, but which Robinson (1980) believes to represent the edge of a basaltic sheet.

The East Greenland continental margin is treated briefly in Section 7.1.

Chapter 7

PLATE TECTONIC EVOLUTION OF THE SOUTHERN NORWEGIAN-GREENLAND
SEA

In the first part of this chapter a plate tectonic synthesis for the evolution of the southern Norwegian-Greenland Sea is presented, based mainly on work presented in Chapters 1 and 6. The implications of this synthesis with respect to the structure and evolution are then discussed. Finally, a possible explanation of the evolution of this area, in terms of the evolution of the North Atlantic and Arctic Oceans is proposed.

7.1 Plate tectonic evolution of the southern
Norwegian-Greenland Sea and surrounding areas

Firstly the plate tectonic evolution of the Norway Basin will be discussed and then it will be shown how this provides the key to understanding the evolution of the surrounding areas.

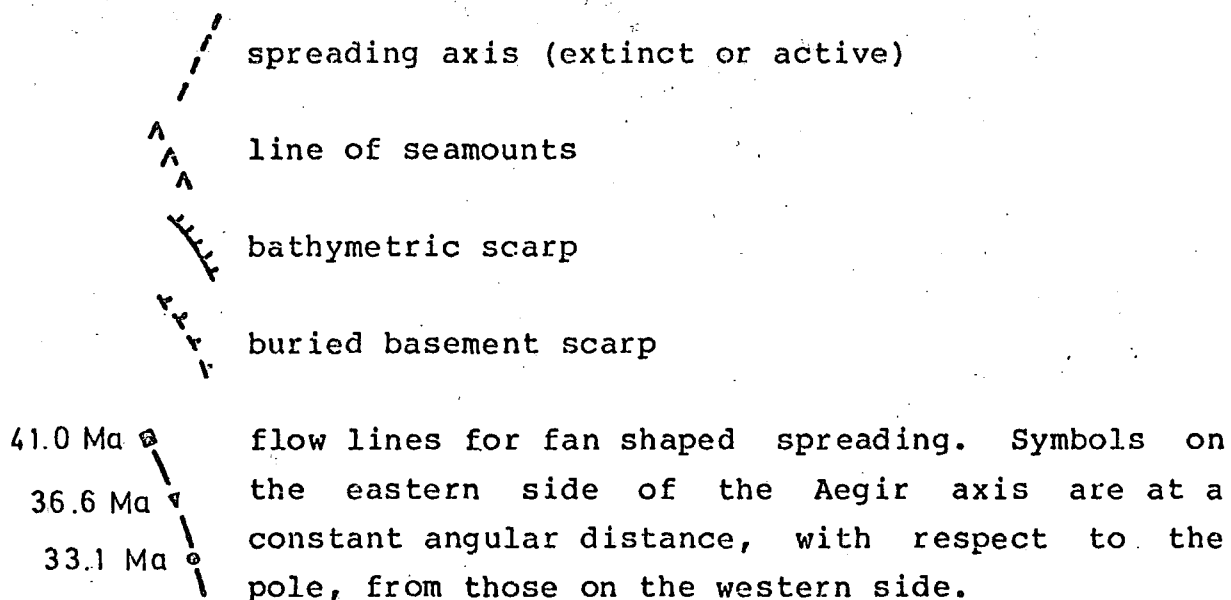
Plate tectonic evolution of the Norway Basin - the fan shaped
spreading hypothesis confirmed and extended

Figure 7.1 shows the magnetic lineations in the Norway Basin redrawn from Figure 6.8 as "zebra-stripes" on a polar stereographic projection. The central zone anomaly pattern (younger than 20) is fan shaped, as originally pointed out by Talwani and Eldholm (1977), who suggested that spreading about a near pole may have been responsible for the convergent pattern, but did not locate a pole other than by specifying that it would need to be near the southern end of the Aegir axis. In the present study a single rotation pole was found (shown in Figure 7.1) which not only accounts for the pattern of the central anomalies, but also for a number of important bathymetric features in the Norway Basin.

Before describing how this pole was obtained it is

Figure 7.1. Polar stereographic projection of the Norway Basin and environs, showing magnetic anomaly lineation, physiographic features and flow lines for fan shaped spreading. Bathymetry in nominal fathoms after Gronlie and Talwani (1978).

Magnetic anomaly lineations after Figure 6.3, 6.6 and 6.8. From the centre of the basin outwards, the heavily stippled stripes correspond to the anomaly 7?-12, 13 and 15-18 bands, while the lightly stippled stripes correspond to anomalies 20, 21, 22, 23, 24A and 24B. The lineations immediately north of the bathymetric bight (65°N , 8°W) which are not identified, are from Vogt et al. (1980).



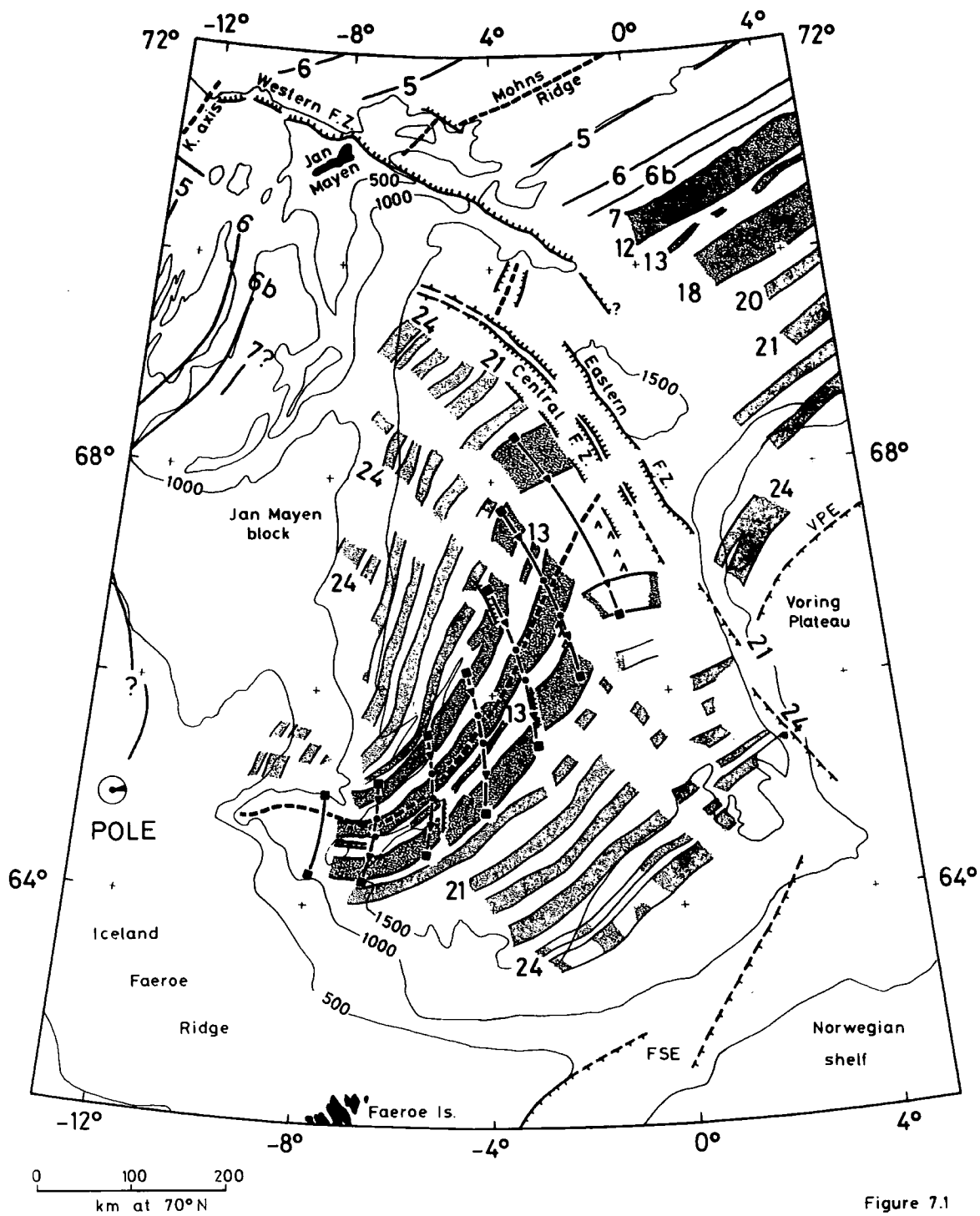


Figure 7.1

necessary to deal briefly with some of the properties of the polar stereographic projection. A small circle about a given pole projects stereographically as a circle whose centre is offset from the projected pole position along a radial line from the axis of the projection (Phillips 1972). However, over a sufficiently small area about the pole the projection preserves distances and angles to a desired degree of accuracy. Thus small circles become nearly concentric about the projected pole and plane angles measured on the projection with the pole as centre are nearly the same as the equivalent angular distances around small circles. It was verified directly that over the small area of the globe represented by Figure 7.1 angles and distances are preserved to well within the accuracy demanded for plate reconstructions. The same is not true for the Mercator projection of the same area.

On account of this property there is a simple graphical construction to find the pole position, P and the amount of finite rotation to match a lineation A to a lineation B, given that the pole is local. Two points X and Y, such that the estimated lines PX and PY are approximately perpendicular and of equal length, are marked on the base chart which shows the lineations A and B. Using an overlay, lineation A is traced off and then the overlay is shifted to give a best fit between the traced outline and lineation B. The points X and Y are then marked on the overlay as X' and Y' and the overlay is restored to its original position. The intersection of the perpendicular bisectors of the lines XX' and YY' then gives P and the angle XPX' (for example) gives the required rotation angle. Flow lines from A to B are circles about P.

The pole position in Figure 7.1 (64.87°N , 12.3°W) was found in this manner by matching the western edge of anomaly 15-18 band on the northwestern side of the axis with the corresponding lineation on the southeastern side. The curvature of the lineations constrains the pole position quite severely. The rotation is 22.2° . The goodness of fit is indicated by the correspondence of the southeastern lineation with the square symbols, which are uniformly 22.2° (with respect to the pole) from their counterparts which are

aligned along the northwestern lineation.

Using the same pole, rotation angles were determined to match the inner edges of the anomaly 15-18 band and the outer edges of the central band (13.0° and 4.2° respectively). The goodness of fit is illustrated as described above, with triangular and circular symbols. It is clear from Figure 7.1 that the same pole can satisfactorily account for the entire fan shaped spreading phase. This is not necessarily an expected result.

There are a number of bathymetric features which provide additional evidence which is consistent with the pole position shown in Figure 7.1. Firstly, the ridges and troughs of the central Jan Mayen Fracture Zone, which was described in Section 1.10, are remarkably parallel to small circles about the pole. The central fracture zone has not been previously described as a distinct entity. It is suggested that the curved fracture zone came into existence at the time that rotation about a near pole began, and that it replaced the Eastern Jan Mayen fracture zone as the locus of transform motion between the Aegir axis and the Mohs Ridge (see below for a more detailed discussion of the region immediately north of the central fracture zone.)

Within the central zone of the Norway Basin there are a number of features (scarps and lines of seamounts) which are indicated schematically in Figure 7.1 and which are parallel to small circles about the proposed pole. Examination of the detailed bathymetry in Figure 1.4 will show that the lineations represented on Figure 7.1 form the major subset of the obvious lineations in the central zone (with the exception of those which are parallel to the Aegir axis). These lineations provide strong corroboration of the implied oblique spreading regime.

A further piece of bathymetric evidence is that closure back to about anomaly 18 time also causes almost complete closure of the prominent bathymetric high which forms a western extension of the Aegir axis, as indicated by the square symbols in Figure 7.1, which are coincident with the

1000 fathom isobath on both sides of the bight. This bight does not appear to have been previously explained. According to the present interpretation (see below), it represents a pre-existing fracture zone which was opened when fan shaped spreading began.

Taken together, the various lines of bathymetric evidence provide close control on the position of the pole, independently of the magnetic anomaly lineations. The central Jan Mayen Fracture Zone provides a strong constraint on the position of the pole with respect to variation in a NW-SE direction, but a less strong constraint in the perpendicular "radial" direction. However the pole must be further southwest than about 66.5°N , 8.5°W , otherwise compression across the southern Aegir axis is implied, during opening. The pole could not lie much further southwest than its proposed position and still give a realistic closure of the bathymetric bight. The lineations in the central zone of the basin constrain the pole to lie within a narrow WSW-ENE trending region of limited extent. This region intersects the line of constraint imposed by the central fracture zone at a position very close to the proposed position based on the matching of the edge of anomaly 18. During the preparation of Figure 7.1, numerous trial poles were considered, all of which gave roughly equivalent overall solutions, with varying emphasis on the relative importance of magnetic lineations, fracture zones et cetera. Nearly all of these poles lie within an oval area about 40 km long and 30 km wide, stretching in a southwesterly direction away from the pole shown in Figure 7.1.

Figure 7.2 is a plot of angular separation versus absolute age. The angular separations obtained from Figure 7.1 (with the addition of a value for the end of anomaly 20 time, see below) fall neatly on a straight line of slope $2.3^{\circ}/\text{Ma}$. Extrapolated straight to zero this curve implies that spreading ceased at anomaly 10 time (32 Ma), but the spreading rate may have slowed down at the end. Accretion beyond anomaly 7 time appears to be ruled out by the non-appearance of the following negative anomaly, even in the northern part of basins where the spreading rates are similar

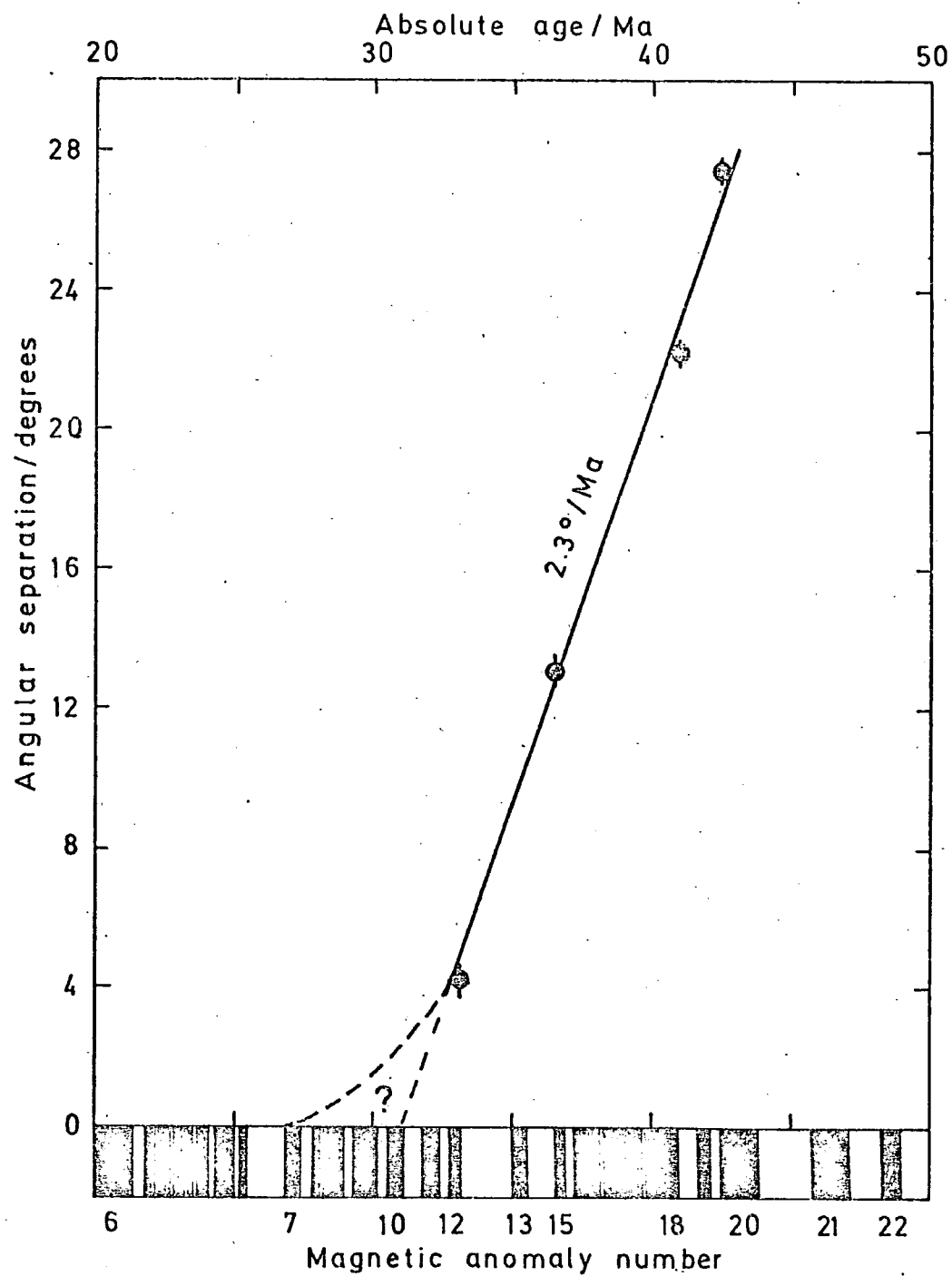


Figure 7.2. Spreading curve for the Norway basin during the period of fan shaped spreading - angular separation versus absolute age.

to those about the Mohns Ridge.

Immediately north of the central Jan Mayen Fracture Zone there is bathymetric evidence of what may be a rift structure representing the extinct axis in this region (also suggested by Larsen 1980; note added during final revision). The position of this feature in relation to the anomalies in the Lofoten Basin, taken at face value, suggests that spreading ceased at about anomaly 6B time (Larsen, 1980). However as was noted in Section 1.10 (see Figures 1.4 and 7.1), there appears to be a curved southeastern extension of the western Jan Mayen Fracture Zone which is parallel to the central fracture zone, so it is possible that the axis was offset dextrally in this region and that spreading ceased prior to anomaly 7 time. Gronlie et al. (1979) have identified anomalies 23 to 21 in the region between the central and western Jan Mayen Fracture Zones. On detailed examination of their data however, there does not appear to be sufficient evidence to support this or any other anomaly identification.

The reconstruction of the Norway Basin at the end of anomaly 20 time (42.7 Ma) which is shown in Figure 7.3 was achieved by closing back to the beginning of anomaly 18 time, as indicated in Figure 7.1, then closing by a further 5.2° about the same pole. A small amount of overlap at the southwest occurs if complete closure is obtained in the north. Anomalies 20 to 24 are parallel on either side of the axis and the inferred anomaly offsets (Section 6.4) match up well, although a better reconstruction would be obtained if the Jan Mayen block were situated slightly further north.

In the south, the bathymetric bight is completely closed. The inferred position of anomaly 20 age crust over the Iceland-Faeroe Ridge is shown (Section 6.4). It is suggested that a 200 km transform fault (offsetting the axis dextrally) existed in the vicinity of the closed bathymetric bight.

In Figure 7.3 it can be seen that the northern part of the Jan Mayen block is cut across by the eastern Jan Mayen Fracture Zone. Upon reconstruction beyond anomaly 20 time,

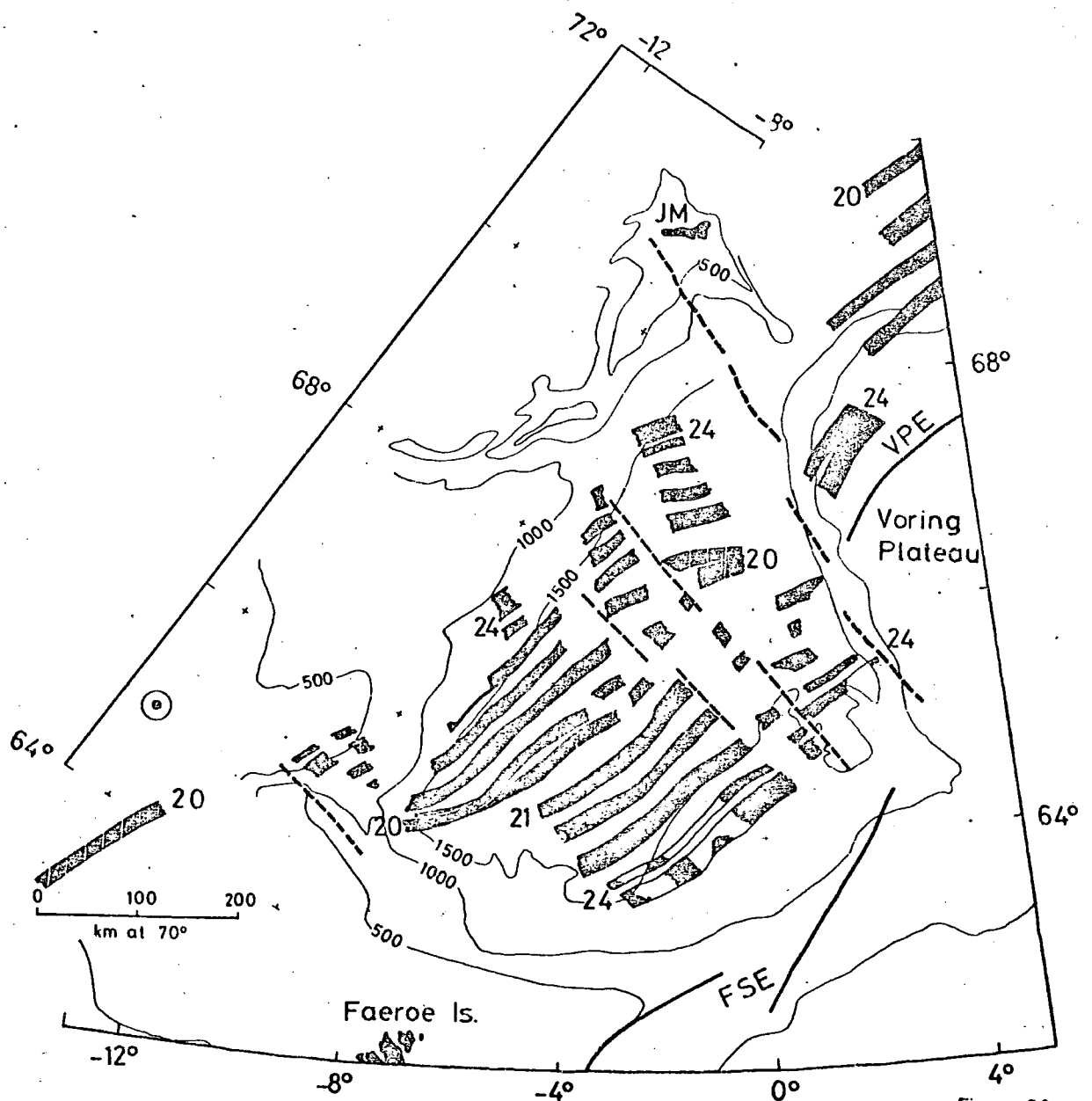


Figure 7.3

Figure 7.3. Reconstruction of the Norway Basin to the end of anomaly 20 time (42.7 Ma), showing proposed fracture zones.

the Jan Mayen bank cannot be accommodated south of the eastern Jan Mayen Fracture Zone. In section 1.7 it was suggested that the Jan Mayen bank might be of fundamentally different composition to the Jan Mayen Ridge further south as it is underlain by highly magnetic crust. It is probable that the bank is underlain by oceanic crust (also suggested by Larsen 1980; note added during final revision). A straightforward interpretation of Figure 7.3 suggests that the bank is underlain by crust of greater age than anomaly 20, but there is another possibility which is discussed below.

It is evident from the reconstruction that the Norway Basin was enclosed prior to about anomaly 21 time, at which time the Iceland-Faeroe Ridge, the Voring Plateau and parts of the Jan Mayen Ridge were probably emergent (Gronlie 1979).

The complementary axis; fan shaped spreading about the Kolbeinsey axis

As inferred by Talwani and Eldholm (1977), simultaneous fan shaped spreading, convergent to the north, must have occurred in some region west of the Aegir axis, during the period that fan shaped spreading was taking place in the Norway Basin. In this study, Talwani and Eldholm's suggestion that this spreading took place within the eastern part of the southern ridge complex is rejected, because it is believed that their region 3 is underlain partly by continental crust and partly by oceanic crust (anomaly 24 age and slightly younger) generated about the Aegir axis (Section 6.5). In addition this region is not large enough to account for all the fan shaped spreading which must have taken place.

Now that the existence of the intermediate Iceland Plateau axis has been disproved (Vogt et al., 1980), the simplest hypothesis is that the complementary spreading took place about the newly formed Kolbeinsey axis. As pointed out in Sections 6.4 and 6.5, there are wedge shaped regions of apparent oceanic crust, divergent to the south, on both sides of the Iceland Plateau.

Figure 7.4 shows a reconstruction of the Iceland

Plateau to anomaly 6B time. This was achieved by rotating the Jan Mayen block by -5.5° (clockwise) about a pole situated at 69°N , 130°E . This pole position represents a rounded average of total opening poles determined for anomaly 5 and 13 times by Talwani and Eldholm (1977) and for anomaly 6 time by Phillips et al. (in press). The amount of finite rotation was determined concordantly from the results of Talwani and Eldholm and of Phillips et al. by interpolating the curves of total angular separation versus time. This is a valid procedure because, from the vantage point of the southern Norwegian-Greenland Sea, which is distant from the total opening poles, they have not shifted very much since anomaly 20 time. In order to construct Figure 7.4 the new coordinates of two widely spaced points from within the Jan Mayen block were computed and located on a tracing of the Greenland margin. The shifted points were then positioned over the original points on a chart of the Jan Mayen bathymetry, which was traced. This is an accurate method because of the conformal properties of the stereographic projection.

Figure 7.4 shows a satisfactory correspondence between the position of anomaly 6B on the Greenland margin and the reconstructed position of the eastern anomaly 6B (Vogt et al., 1980). Under the assumption that spreading in the Norway Basin ceased at about anomaly 7 time (and neglecting the small difference in age between anomaly 7 and anomaly 6B), Figure 7.4 is believed to represent the configuration about the Kolbeinsey axis at the time that fan shaped spreading ceased.

In a qualitative way, it is clear from Figure 7.4 that complementary fan shaped spreading did take place about the Kolbeinsey axis. In order to rigorise this concept it is necessary to find a finite closure pole for Jan Mayen with respect to Greenland, which is consistent with the relative poles for Eurasia-Jan Mayen and Eurasia-Greenland. In addition this rotation should bring the northern scarp of the Norway Basin bathymetric high into line with the Denmark Straits fracture zone and the Jan Mayen ocean-continent boundary to a reasonable position seaward of the Greenland

Figure 7.4. Reconstruction of the Jan Mayen block relative to Greenland, at anomaly 6B time (24 Ma). Bathymetry in nominal fathoms after Gronlie and Talwani (1978). Bathymetric contours, anomaly 6B lineation and ocean-continent boundary (OCB) shown dashed for the Jan Mayen block, solid for Greenland. Showing the proposed finite rotation pole (GR-JM) to close JM against GR, at 42.7 Ma. XX' is the line along which this pole must lie to bring the bathymetric high into line with the Denmark Strait Fracture Zone (DSFZ).

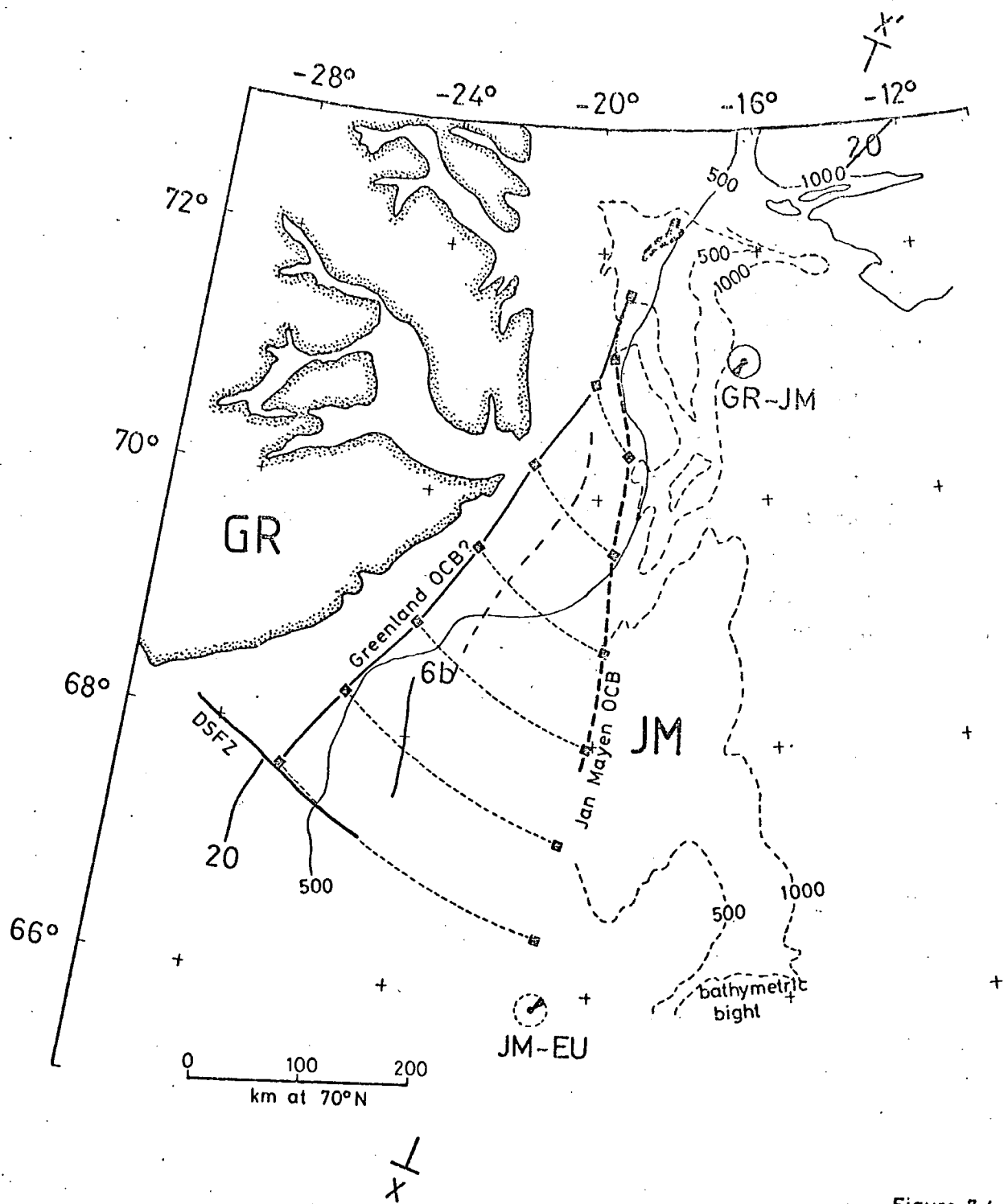


Figure 7.4

coast.

Again under the assumption of near constancy of pole position (from the vantage point of the southern Norwegian-Greenland Sea) the angular separation of Eurasia and Greenland between the end of anomaly 20 time and anomaly 7 time can be estimated from the results of Phillips et al. (in press), as 3.0° with a probable error of at least $\pm 0.2^{\circ}$. In the reference frame of Greenland the mean finite difference pole in this period appears to have been located in the vicinity of 55°N , 130°E (Phillips et al., in press).

Using the quaternion method of Le Pichon et al. (1976) a possible finite rotation for the closure of Jan Mayen against Greenland was determined by following the rotation describing the opening of Eurasia with respect to Jan Mayen (in its shifted position as shown in the figure, the pole is at 65.9°N , 21.0°W and the rotation is -27.4° (clockwise)) by the closing rotation for Eurasia relative to Greenland, as described above (55°N , 130°E , -3.0°). The composite pole is 70.88°N , 19.50°W , with a rotation of -29.14° .

This pole is not satisfactory, even though it brings the Jan Mayen ocean-continent boundary to a reasonable position, because it swings the northern scarp of the bathymetric bight about 70 km south of the Denmark Strait fracture zone.

As the total opening rotation pole for Eurasia relative to Greenland is not precisely determined, it was considered reasonable to adjust the position of the pole for Jan Mayen-Greenland to obtain a more satisfactory closure. It is clear that, in the polar stereographic projection of Figure 7.4, the closure pole should lie roughly along the perpendicular bisector of the line joining the eastern and western ends of the Denmark Straits fracture zone and the scarp of the bathymetric bight, respectively, in order to bring these features into line. This line is indicated as XX' in Figure 7.4. The pole cannot lie very far southward along this line otherwise the northern part of the Jan Mayen block moves away from Greenland during closure. Alternatively, if the pole lies much further north than about

72°N the amount of clockwise rotation that can be accommodated without causing an unrealistic overlap of the Jan Mayen bank is significantly smaller than the amount of rotation which took place in the Norway Basin. The composition of the Greenland-Jan Mayen rotation which is implied by such a pole position with the Jan Mayen-Eurasia rotation then gives rise to a total spreading azimuth which is too far south of east and to a total opening pole which lies to the south of the area instead of to the north.

The chosen pole position is at 71.12°N and 16.36°W in the reference frame of Greenland (Figure 7.4). It lies to the south of the Jan Mayen block and is only 110 km, in an easterly direction, from the derived pole which was discussed above. In Figure 7.4 the western Jan Mayen ocean-continent boundary has been extended southward from its mapped position (Figure 6.13) and indicated by the line of square symbols. If the Jan Mayen block is rotated by -29.0° about the chosen pole this line is shifted to a satisfactory position off the Greenland coast and lies just seaward of the anomaly 20 lineation south of the Denmark Strait fracture zone, consistent with a commencement of spreading at the end of anomaly 20 time.

Composition of the Jan Mayen-Greenland opening about this pole and the Eurasia-Jan Mayen opening gives a total opening of 3.12° about a pole situated at 118.3°E, 48.1°N. This pole position lies within the area of scatter of determined finite difference poles for anomaly 7 to anomaly 20 time. For example Phillips et al. (in press) position the anomaly 13-18 pole at 39.6°N, 116.2°E, the anomaly 12 to 13 pole at 55.0°N, 130.4°E and the anomaly 6 to 12 pole at 55.4°N, 157.9°E. The amount of rotation (3.1°) is also consistent with the results of Phillips et al. It is considered that these results provide strong support for the proposed complementary spreading regime, particularly as it was found (by trial and error) that the composite pole position was sensitive to the position of the Greenland-Jan Mayen pole.

The rotated Jan Mayen ocean-continent boundary implies a

position for the Greenland ocean-continent boundary which is consistent with recently acquired knowledge of the Greenland margin. It lies within the area of the Scoresby Salient, but Brooks (1979) and Vogt et al. (1980) have suggested that this feature was formed by very rapid progradation in late Tertiary times. Larsen (1980) has presented a schematic cross section of the Greenland shelf between 68°N and 70°N , based on the multichannel seismic data of Hinz and Schluter (1979a,b), which is reproduced as Figure 7.5. This is consistent with an ocean-continent boundary very close to the coast, immediately seaward of the plunging plateau basalts. Larsen's proposed ocean-continent boundary is very similar in position to that shown in Figure 7.4 (which is based on results obtained prior to the publication of his paper), but his evolutionary scheme is not the same as that which is presented here.

If the opening pole was in the position shown, then the northern part of the Jan Mayen block must have moved southward relative to Greenland, by about 60 km during the period of fan shaped opening. As it can be assumed that crust of anomaly 20 age and older in the Greenland basin was rigidly attached to Greenland during this period it is clear that new crust must have been created to the north of the continental Jan Mayen block during this period. Extrapolating the proposed Greenland ocean-continent boundary northward it can be seen from the position of anomaly 20 in the Greenland basin that there must have been a dextral offset between the newly formed line of separation and the Mohns axis at the end of anomaly 20 time. It is suggested that a short easterly oriented spreading axis (or alternatively a very leaky transform fault) was formed to the north of the continental Jan Mayen block and that as the block moved south, part or all of the Jan Mayen bank formed by accretion about this axis.

Closure of the southern Norwegian-Greenland Sea

Figure 7.6 shows a closure of the southern Norwegian-Greenland Sea to the beginning of anomaly 24B time (52 Ma before present). In order to construct this diagram

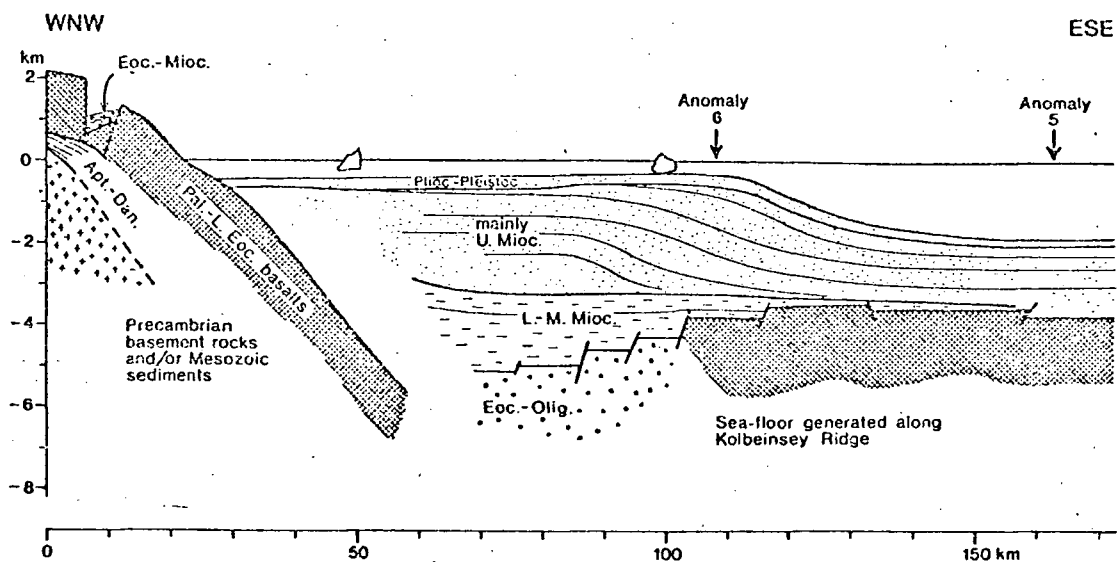


Fig. 8. Schematic cross-section of the shelf between 68° and 70°N (sub-area B) showing the assumed deep structure. Basal Miocene unconformity on shelf contemporaneous with unconformity in onshore Eocene - Miocene sediments. Seismic data from Flinz and Schlüter (1978, 1979).

Figure 7.5. The East Greenland margin between 68° and 70°N, from Larsen (1980). According to the present interpretation the Eocene-Oligocene basement is oceanic crust generated by fan shaped spreading about the Kolbeinsey axis from 43 to about 27 Ma.

Jan Mayen was closed against Greenland, as indicated in Figure 7.4 and the Norway Basin was closed to the end of anomaly 20 time (43 Ma) as shown in Figure 7.3. Spreading prior to 43 Ma was assumed to be governed by Talwani and Eldholm's (1977) early spreading pole (5.74°S , 124.9°E in the reference frame of Eurasia). The average width of the Norway Basin at 43 Ma, from the outer edge of anomaly 24B on the western side to the same lineation on the eastern side, was determined from Figure 7.3 as about 700 km. Talwani and Eldholm's pole lies about 12300 km from the Norway Basin, accordingly 3.3° of rotation is required to close from 43 to 52 Ma (this compares with 3.5° extrapolated from the total opening rotation of Phillips et al. in press, for anomalies 20, 23 and 24). Using this pole and rotation angle, Greenland and Jan Mayen were closed with Eurasia using the method which was described above (for the closure of the Iceland Plateau to anomaly 6B time).

Figure 7.6 shows a good correspondence between the lineations, except in the north where there is a 20 km overlap. It is noted that closure about this pole did not follow exactly the fracture zones which are shown in Figure 7.3; possibly one or more of these features is wrongly positioned. The Jan Mayen block (south of the Jan Mayen bank) fits well between the Eastern Jan Mayen Fracture Zone and the Faeroe Islands. The bathymetric high is not shown in Figure 7.6, because it formed within oceanic crust and hence is reconstructed further east than the 52 Ma isochron. Note that because the ocean-continent boundary of East Greenland in the Denmark Straits area lies at least 100 km west of anomaly 20 (Vogt et al., 1980), so must the line of initial rifting, south of the rotated Denmark Straits fracture zone, lie at least 100 km to the west of the Greenland-Jan Mayen suture. The reconstruction of Figure 7.6 therefore supports the contention that the Faeroes block is continental (Bott et al., 1971; Bott et al., 1974) rather than oceanic (Talwani and Eldholm, 1977), assuming that oceanic accretion began at 52 Ma. Even allowing for a further 70 km of spreading prior to anomaly 24B time (1 Ma at 3.5 cm/a half rate) one must conclude that the Faeroes platform is continental.

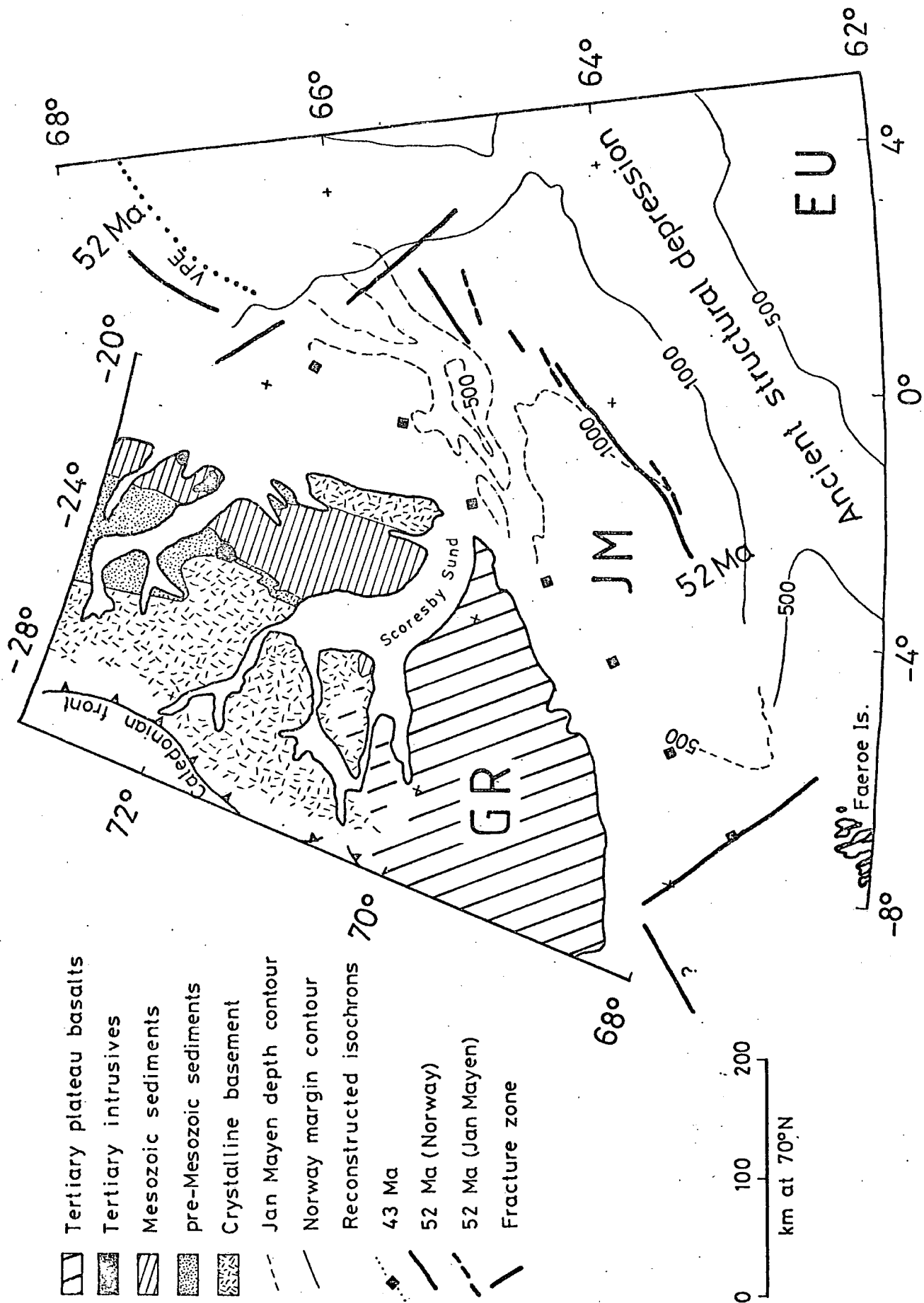


Figure 7.6 Closure of the southern Norwegian-Greenland Sea

7.2 Implications with respect to the structure and evolution of the Jan Mayen block

In this section the implications of the proposed plate tectonic scheme are discussed with respect to the Jan Mayen block. Firstly, the various rotations which were composed to give the reconstruction in Figure 7.6 are consistent with the relative motions of Greenland and Eurasia, as determined by Talwani and Eldholm (1977) and Phillips et al. (in press) using anomaly lineations to the north and south of the area, where no microcontinental fragments have been involved in the spreading since anomaly 24 time. The reconstruction of Figure 7.6 therefore supports the hypothesis that the Jan Mayen block is partly underlain by continental crust. However a rigorous quantitative analysis of all the spreading data from the northeast Atlantic would be necessary in order to assess the width of the Jan Mayen microcontinental fragment independently of the data presented in Figure 6.13, which indicate that the fragment is about 90 km wide.

With reference to Figure 7.6, the possibility that the Jan Mayen block is underlain by oceanic crust formed during an earlier spreading phase (Section 1.6) can be discussed more fully. Bott (1975c) suggested that the early spreading in the Rockall trough continued further north to separate the Faeroes block from Europe across the Faeroe-Shetland Channel and that a certain amount of spreading could have taken place even further north. The possibility that a transform fault existed to the north of the Faeroes block, offsetting this axis to the line of the Jan Mayen block, cannot be dismissed a priori. However Robinson (1980) has found evidence for a linear basement depression, underlain by relatively thin crust, lying between the present 500 and 1000 fathom isobaths on the Norwegian margin, and has suggested that this feature represents the zone of spreading or at least crustal thinning, which was associated with the creation of the Rockall trough.

The reconstruction of Figure 7.6 provides some constraints on the likely pre-Tertiary geology of the Jan Mayen block. In the diagram, the geology of East Greenland

is shown schematically, simplified and extrapolated over areas of ice cover after Escher and Watt (1976). The reconstruction suggests two possible sources for the East Greenland flood basalts, which extend inland from the Blosseville Coast. Firstly, they may all have been erupted from the rift south of the Denmark Straits fracture zone and from the fracture zone itself, that is, from the "hot spot" of the incipient Icelandic transverse ridge. Alternatively, some of the lavas, at least, may have spread laterally from the site of the Aegir axis. In the first case it is difficult, in the second impossible, to avoid the conclusion that most if not all of the southern part of the Jan Mayen continental fragment must also have been covered by great thicknesses of flood basalt. It is also reasonable to suppose that this material has persisted. Thus it is suggested that Horizon 0, in the region of the southern ridge complex, may represent plateau basalt. This basalt was not drilled at DSDP site 350, as the material from this site was dated at 41 Ma (Section 1.8).

The likelihood that Horizon 0 also represents plateau basalts under the Jan Mayen Ridge sensu stricto is difficult to evaluate. It is possible that the adjacent land area north of Scoresby Sund was once covered by basalts (Soper et al. 1976b) and if this were the case, remnants of the lava pile are to be expected under the Jan Mayen Ridge. The fact that the magnetic field over the Jan Mayen Ridge is quiet south of 70°N , does not preclude the possibility that the rocks of Horizon 0 are significantly magnetised, providing that the magnetisation is fairly uniform. It was hoped to analyse the magnetic field over the ridge, in the vicinity of faults and cut-offs in Horizon 0, in order to determine its likely magnetisation, but time did not allow this exercise. However, Navrestad and Jorgensen (1979), were able to determine reliable depths to basement from their aeromagnetic data, which suggests that Horizon 0 is comprised of significantly magnetised rocks.

In addition to plateau basalts, there are a wide variety of rock types present in East Greenland (Escher and Watt, 1976), which may be represented at depth beneath the Jan

Mayen block. Hinz and Schluter (1979b) have suggested that under the Jan Mayen Ridge, Horizon 0 is comprised of crystalline basement, overlain on its dip slope by a relatively thin Mesozoic sequence. This is a plausible model in view of the geology of Liverpool and Jamieson Lands and is also consistent with the seismic reflection configuration of the basement. The reflection characteristics of continental plateau basalts do not appear to be well known.

According to the proposed scheme, the initial stage of separation of the Jan Mayen block from Greenland lasted from 43 to 27 Ma before present. During this period, the southern part of the Jan Mayen block moved more or less at right angles away from Greenland, while the northern part moved transcurrently down the margin. The 43 Ma age for the initiation of the split agrees very well with the age of the basalt drilled in hole 350, on the crest of ridge 1 (41 Ma, Section 1.8). The tectonic disruption of the southern ridge complex post-dated the initial split. It is conjectured that the configuration of en echelon rifts (as suggested in Section 1.8) may have resulted from the shear imparted to the Jan Mayen block in consequence of its rotational motion with respect to Greenland. The final phase of effusive volcanism, during which the topography of the southern ridge complex was flooded by the basalt which now forms the flat lying opaque layer may have taken place after fan shaped spreading had ceased. If extensive lateral flow of magma was involved, as seems likely, this may have been partially responsible for the obliteration of magnetic anomalies over the western part of the Jan Mayen basin.

A 43 Ma age for the initial separation of the Jan Mayen block casts doubts upon the hypothesis that the Horizon A unconformity was formed in consequence of the separation (Johnson and Heezen, 1967). The unconformity may have been formed any time after 40 Ma, possibly as late as 27 Ma, corresponding to a possible age for the oldest sediments of sequence 1 (Section 1.6). It is tentatively suggested that Horizon A has nothing to do with the separation of Jan Mayen but was instead caused by the major mid-Oligocene drop in sea level at about 30 Ma before present (Vail et al., 1977) when

the sea level dropped about 250 m below its present level.

7.3 An hypothesis for the plate tectonic evolution of the northeast Atlantic since anomaly 20 time

Figure 7.7a which was constructed from the polar stereographic chart of Figure 1.1, shows a reconstruction of the North Atlantic and Arctic Oceans at the end of anomaly 20 time. The details in the vicinity of the southern Norwegian-Greenland Sea are very similar to those shown in Figure 7.3. At the time, the Aegir axis was offset to the east relative to the main trend of the mid-Atlantic ridge. The nearly enclosed Norway Basin was bounded to the north by the Eastern Jan Mayen Fracture Zone and to the south by the proposed fracture zone on the northern side of the Icelandic transverse ridge.

At about the end of anomaly 20 time, spreading ceased in the Labrador Sea which meant that the spreading direction in the northeast Atlantic reoriented in an anticlockwise fashion, as indicated in Figure 7.7a. The new direction of spreading was not parallel to the old transform faults. If only sinistral offset in the axis had been present, the corresponding transform faults would have become "leaky" and spreading would have continued much as before except in an oblique direction. However, the dextral offset in the axis south of the southern Norwegian-Greenland Sea constituted an impediment which prevented this from taking place as the new direction of spreading caused compression across the transform fault corresponding to the offset.

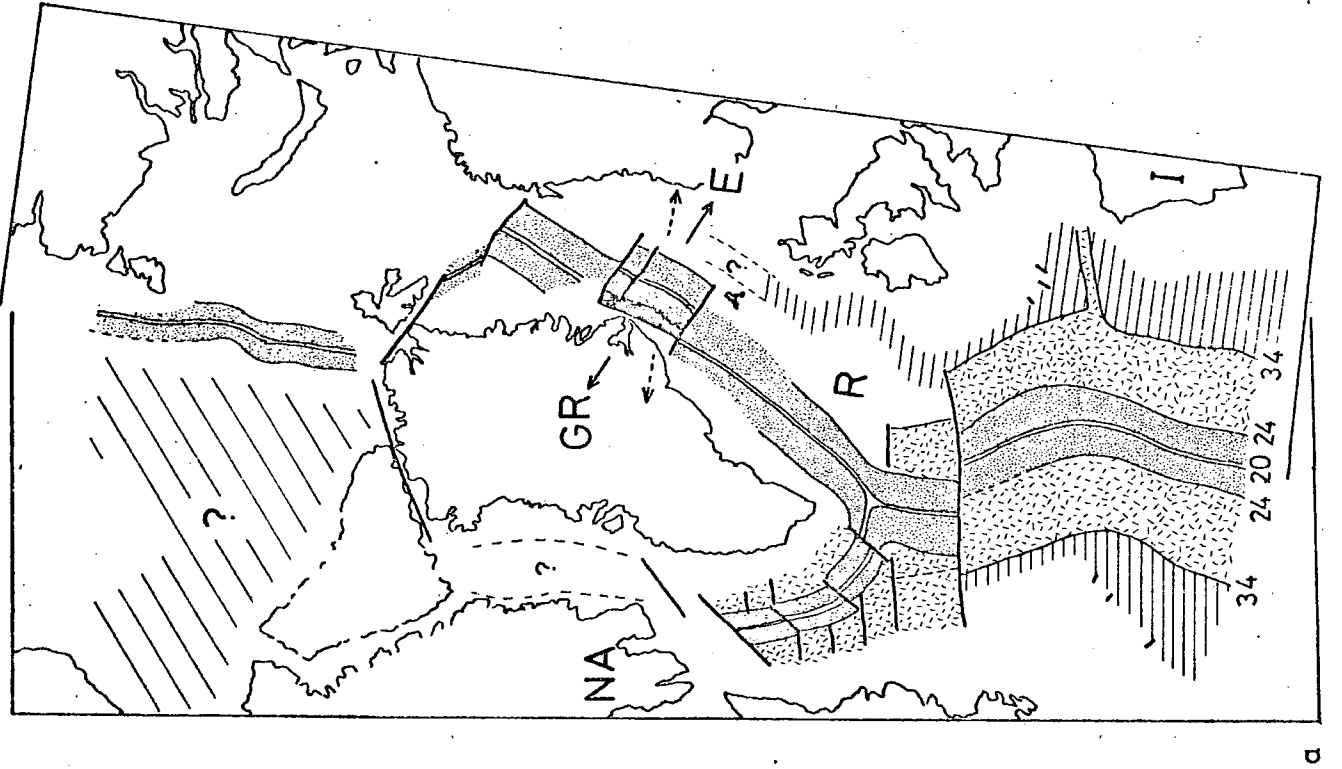
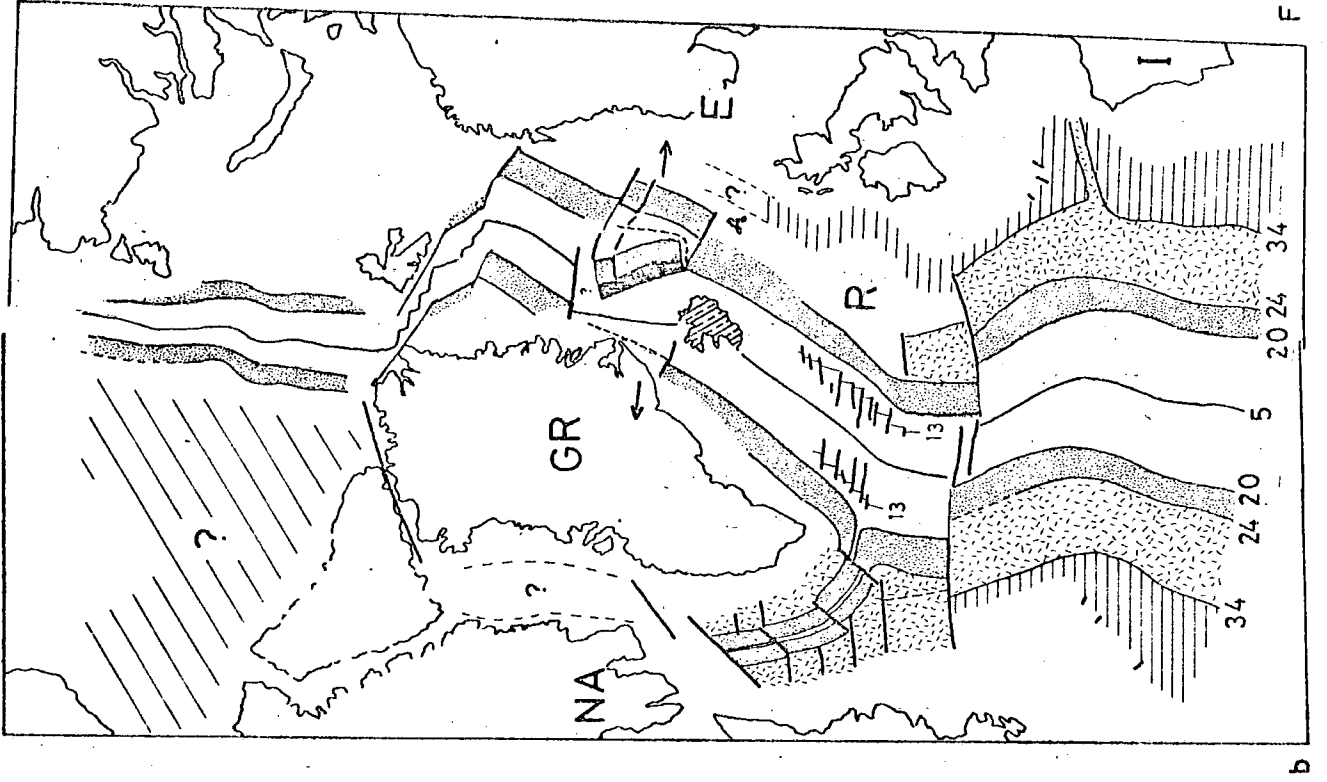
It is hypothesised that the resulting stresses caused the detachment of the Jan Mayen block. This was accompanied by the propagation of a new rift northwards along the line of the Kolbeinsey axis. Continued spreading in the new direction caused the rotation of the Jan Mayen block. This continued, with simultaneous fan shaped spreading on both sides of the block, until the Jan Mayen microcontinent had reoriented sufficiently and free movement was able to take place. Spreading in the Norway Basin then ceased and all of the spreading took place about the Kolbeinsey axis.

Figure 7.7. Plate tectonic evolution of the northeast Atlantic from 43 Ma to 10 Ma. Based on the polar stereographic charts and magnetic lineations of Figure 1.1. NA-North America, GR-Greenland, E-Eurasia, R-Rockall microcontinent, I-Iberia. The Jan Mayen microcontinental fragment is shown black.

Figure 7.7a shows the configuration at the end of anomaly 20 time (43 Ma), when spreading ceased in the Labrador Sea and spreading between Greenland and Eurasia reoriented from the direction shown by the solid arrows to that shown by the dashed arrows.

This caused compression across the fracture zone north of the Faeroes and the Jan Mayen block was split off and rotated, as shown in Figure 7.7b, which shows the configuration at anomaly 5 time, after spreading had ceased in the Norway basin. The shape of Iceland at that time (striped) is based on the present coastline.

Figure 7.7



The reconstruction of Figure 7.7b illustrates the situation at anomaly 5 time, a while after the Aegir axis became inactive. It is hoped that a comparison of Figures 7.7a and 7.7b will suggest the details of the proposed mechanism rather more fully than they have been described above. It seems plausible that the stresses originating at the southern transform fault after anomaly 20 time would have propagated southward. It is suggested that these stresses also caused the formation of closely spaced fracture zones about the Reykjanes Ridge (Vogt and Avery, 1974). Because of the decoupling across the Jan Mayen Fracture Zone the stresses were not transmitted northward, so fracture zones were not formed about the Mohns Ridge. When all stresses opposing the new spreading direction had been dissipated by the reorientations within the southern Norwegian-Greenland Sea the fracture zones south of Iceland disappeared and an oblique spreading regime was established.

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Appendix A

RAW AND REDUCED DATA FROM THE 1977 CRUISE

In addition to the monitor records which were used in this study, 150 seismic field tapes (numbers 151-300) containing data in multiplexed form (Figure 2.4) are stored in the department.

Annotated analogue depth, gravity and magnetic records and ship's logs are also stored. Reduced data profiles, drawn on the IBM1130 system (Figure 2.5) are reproduced on the following pages. Additional track charts and plots of data along ship's track, which are not reproduced here, are also available.

The original Merged-merged Format 3 (Talwani et al. 1972, also see Appendix C) data tape containing 2 minute values (Figure 2.5) for all of the cruise is stored at Newcastle, within the NUMAC system. The tape name is GPO401 (owner GPO4) with characteristics as follows; 1/2 inch, 9 track, 800 BPI, NRZ, IBM compatible. Each logical record (80 characters) is separated by an inter-block gap. Commands for the recovery of data to a temporary file "-file" are as follows

```
$E -file
$MOUNT GPO401 *T* NV
$copy *T* -file
```

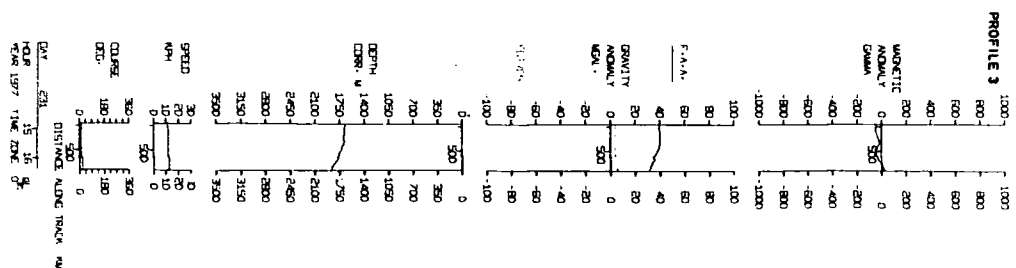
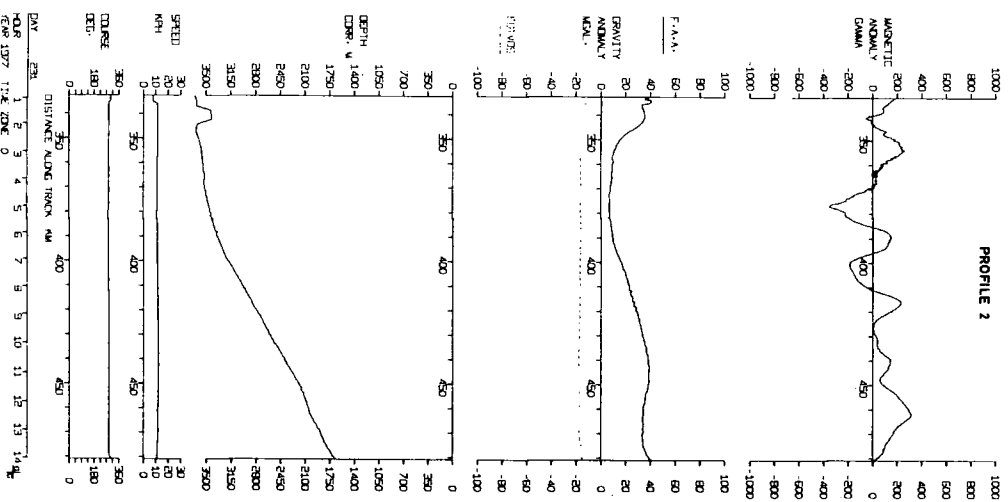
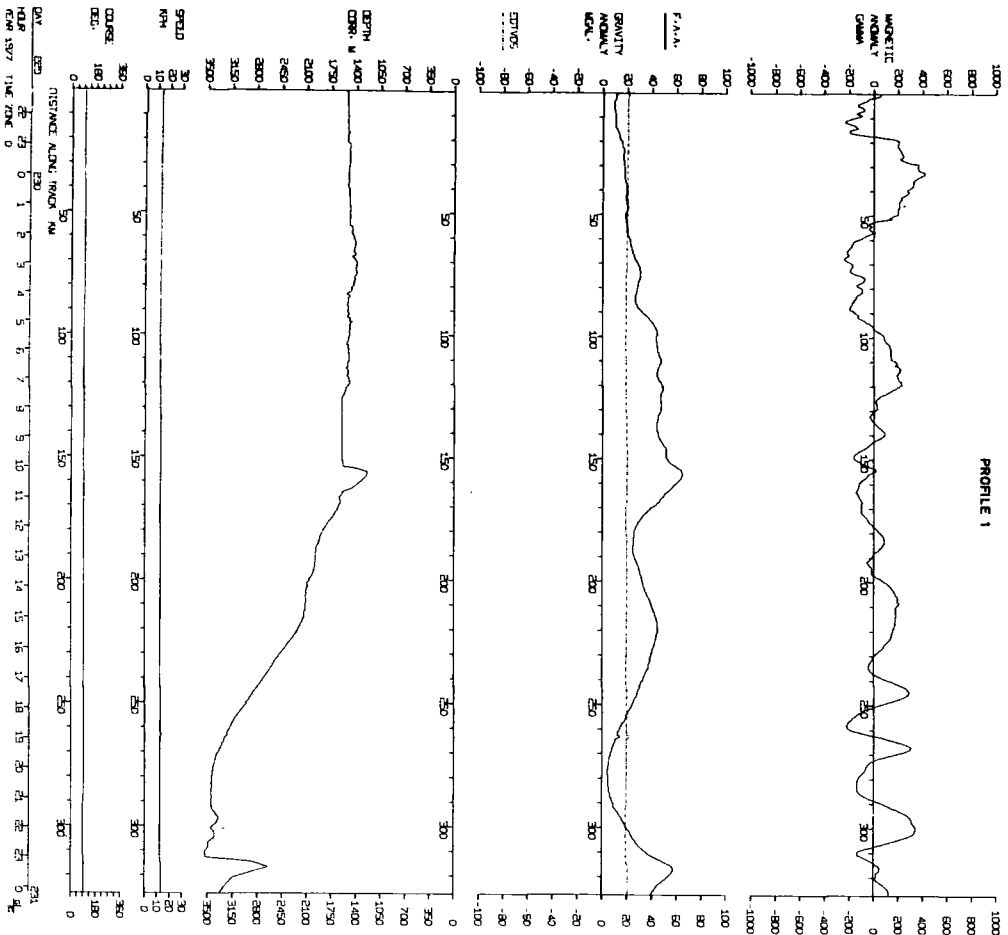
The gravity data on this tape and on the reduced data plots which follow are 14.8 mgal too high (Section 2.4).

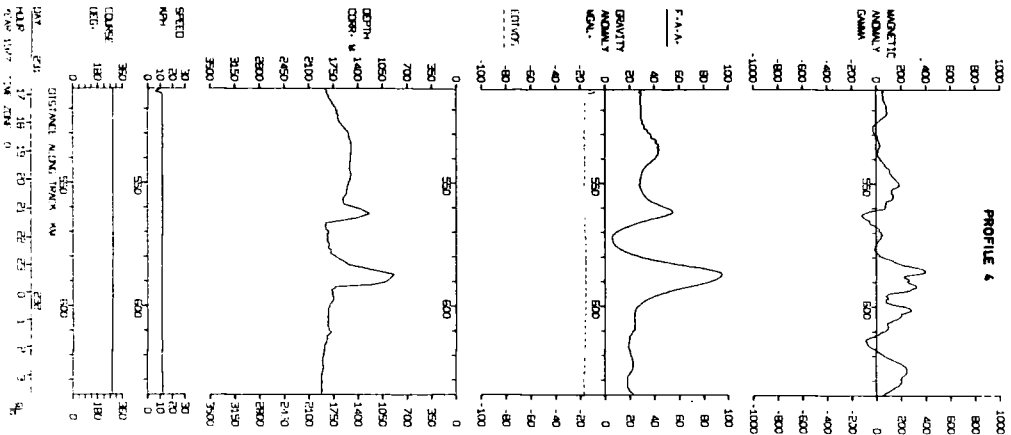
As described in Section 2.5, a modified Merged-merged Format 3 data file (with corrected gravity values) was produced for the second part of the cruise, with the ship identification in columns 1 to 8 replaced by profile and seismic tape indicators and with distance information in columns 43 to 52. This file has been recorded on the 1600 BPI tape GPO4FS (owner GPO4) which is also stored at Newcastle. For ease of data recovery the tape was written using the MTS file save program *FS (Greenwood, F.T., 1975. NUMAC programming note 73, *FS). The file was stored in documented form using the DSAVE command. The commands to mount the tape and restore the file (SHACK977.2) to a temporary file "-file" and to list the documentation are as follows

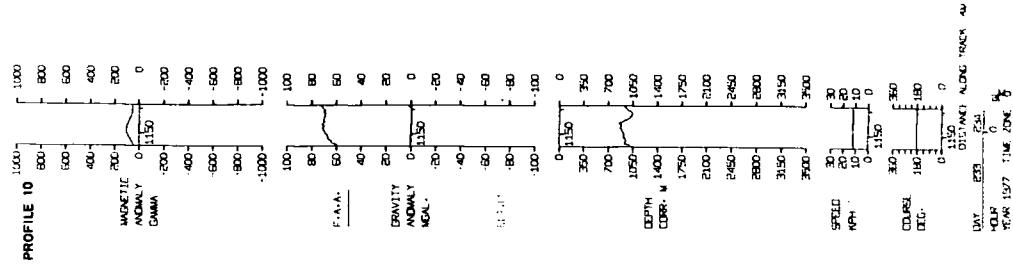
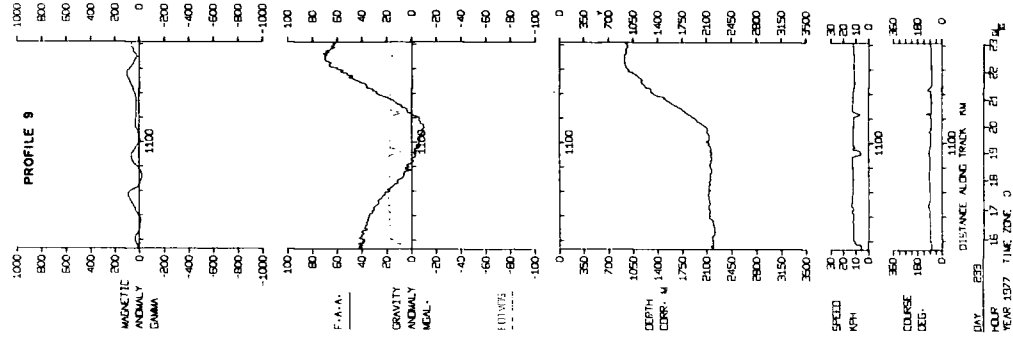
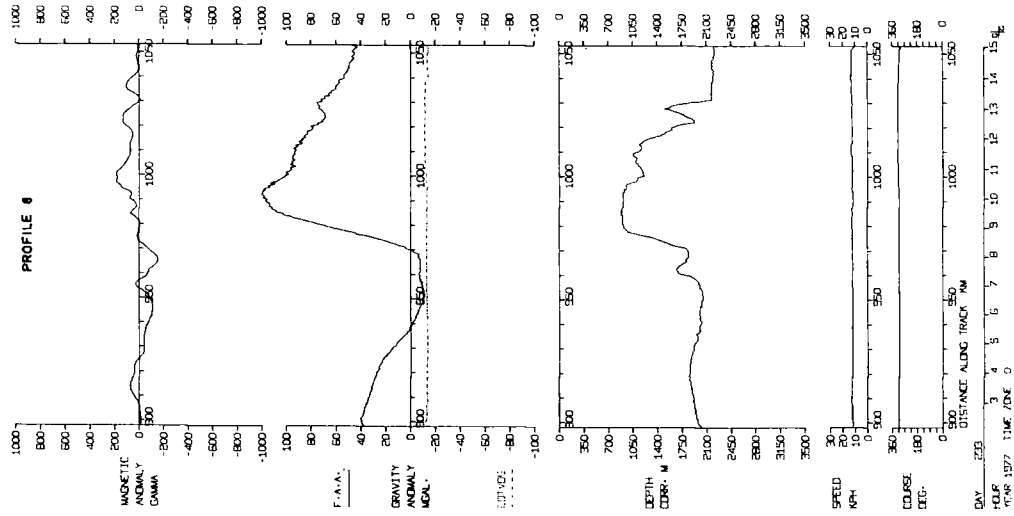
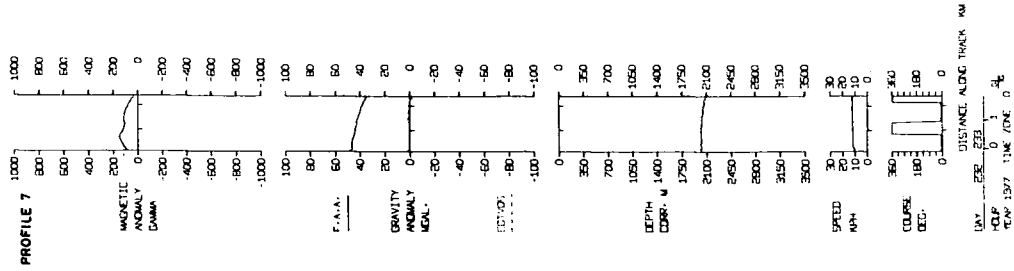
```
$E -file
$MOUNT GPO4FS *T* NV
$RUN *FS 0=*T*
RESTORE SHACK977.2 -file
DLIST SHACK977.2 ON *PRINT*
STOP
```

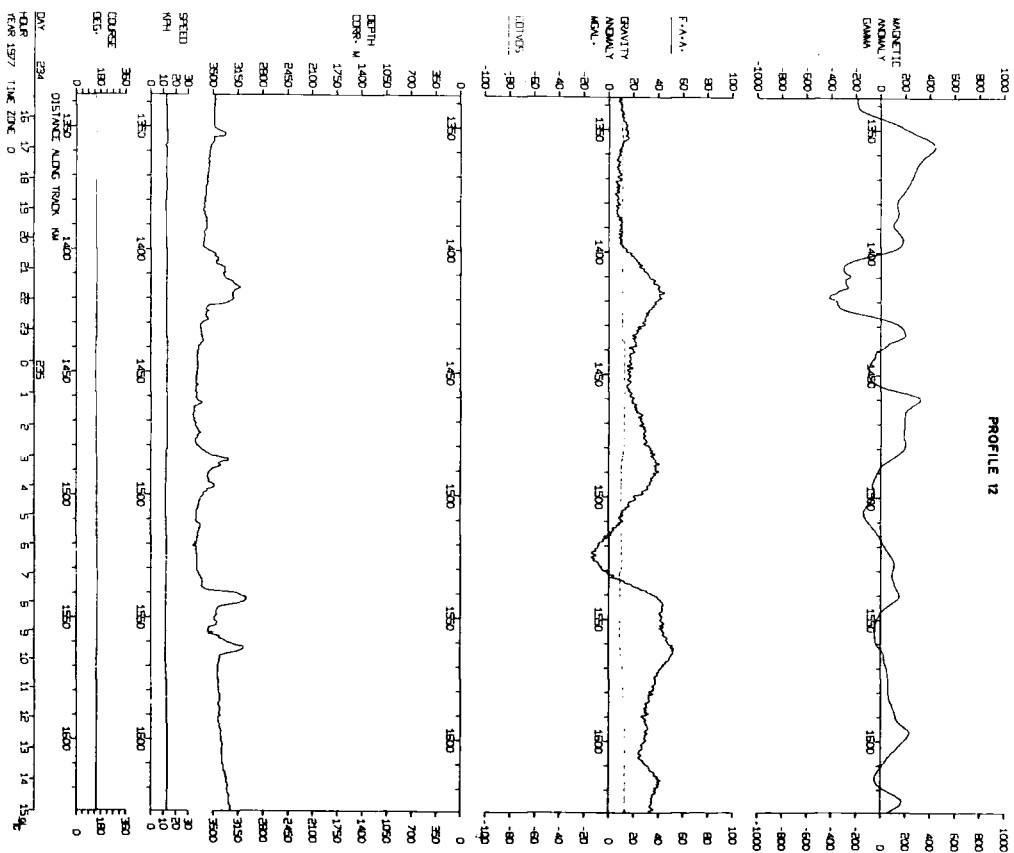
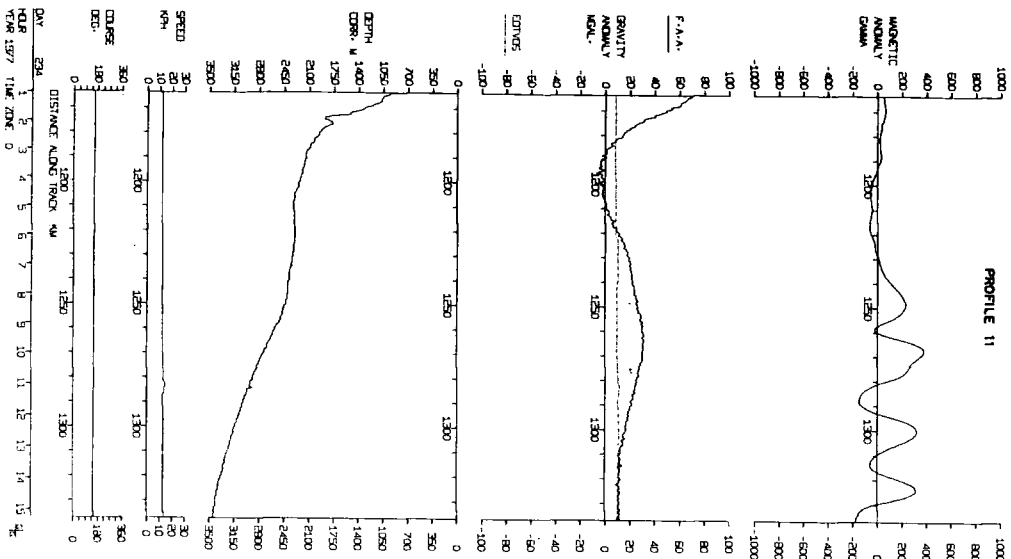
The documentation contains full details of data format. The programs described in Appendix B.1 are also recorded (with documentation) as *FS files DIST, TIMDEP, SPLINE, GRVINV, MAGINV, FURPAK. In addition the files TIM1, TIM2, TIM4,

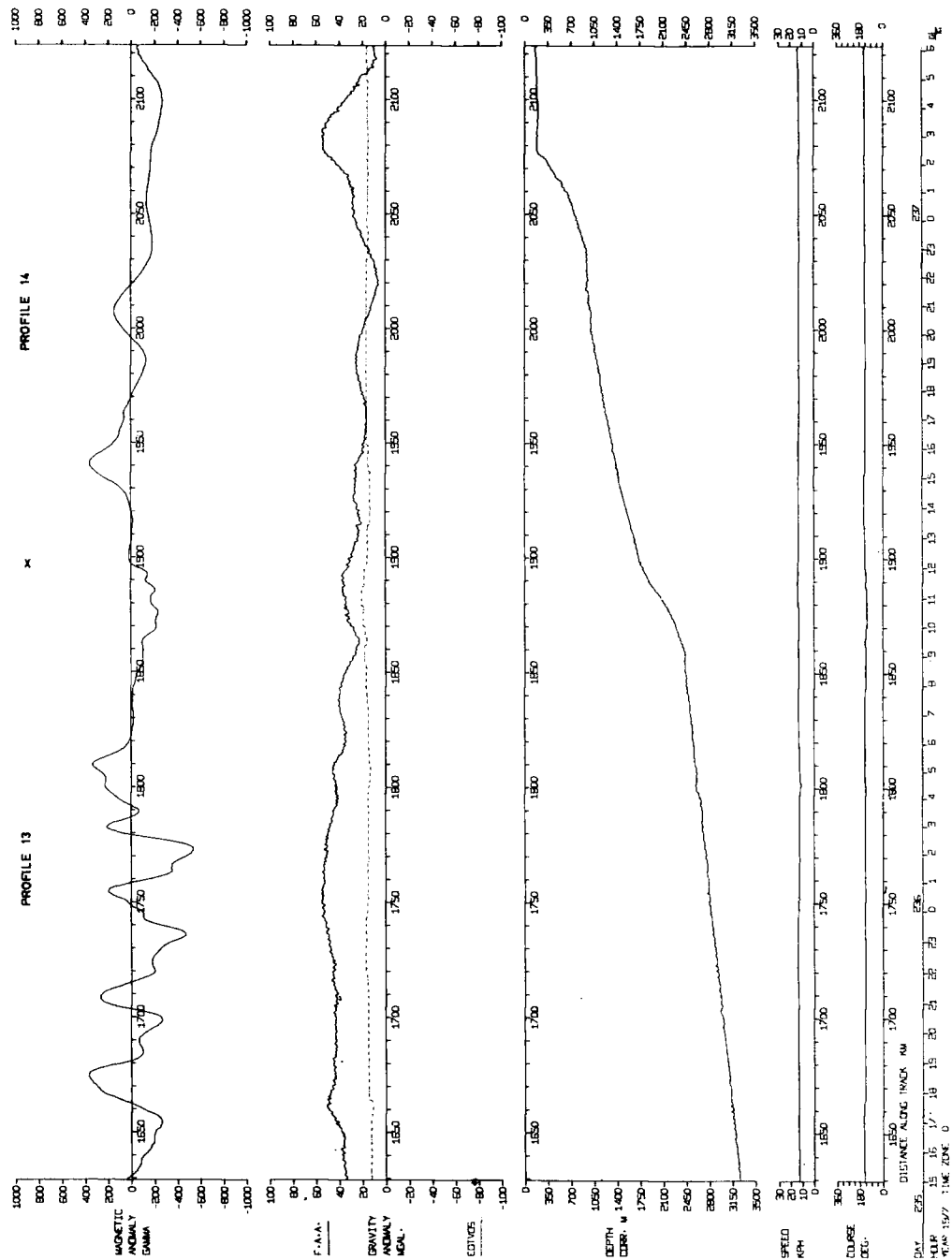
TIM6, TIM8, TIMN contain the two way times between the sea bed and acoustic basement (Section 4.2) for profiles 1 to 8 and the Norway Basin profile. The data format is as required for input to TIMDEP (Appendix B). Any of these files can be recovered in the manner indicated above. The *FS facility is MTS dependent, but will be supported for the next 8 years at least (N.F.Hall, personal communication).

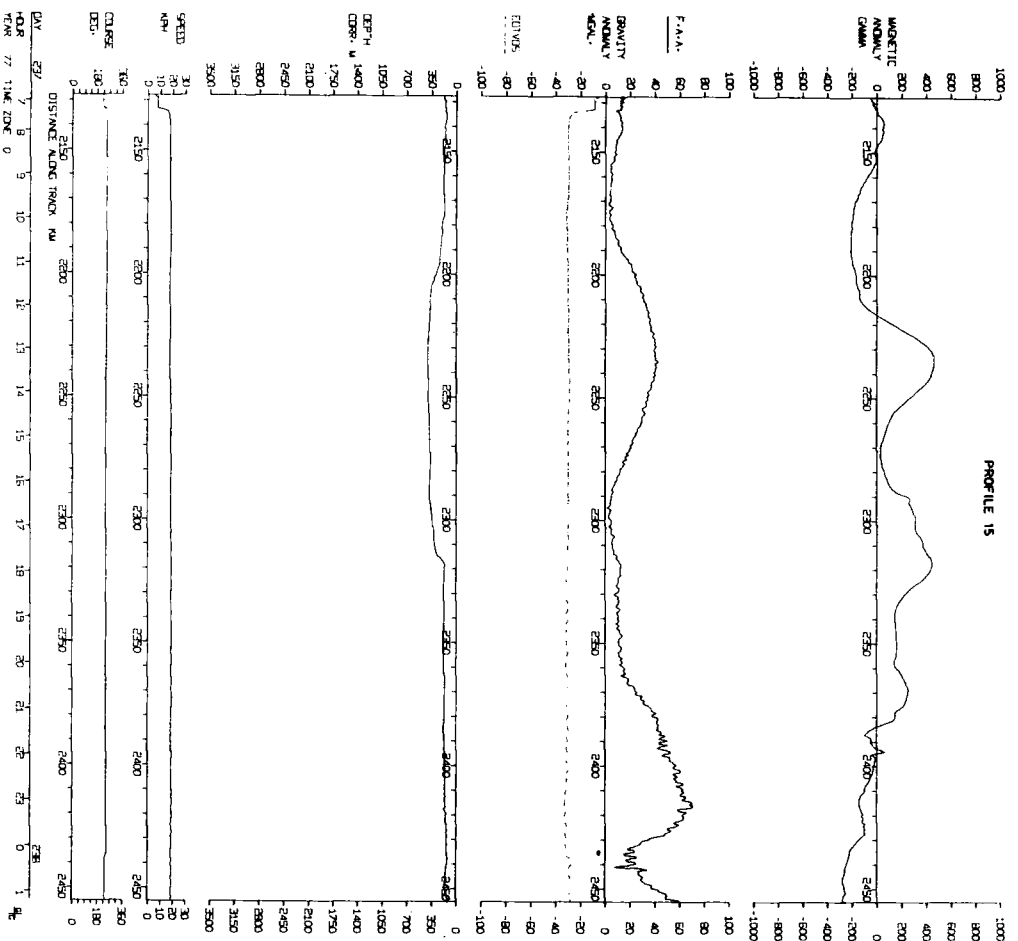












Appendix B

DESCRIPTIONS AND LISTINGS OF COMPUTER PROGRAMS

B.1 Programs for interpreting the 1977 cruise data

Descriptions

DIST

Purpose

To compute distance along ship's track as described in Section 2.5.

Run command

\$RUN OBJ.DIST 4=infile 7=outfile1 8=outfile2

Description of input files

infile is a line file containing geophysical information in Merged-merged Format 3 (Talwani et al. 1972, see Appendix C) or in any modification which has latitude and longitude (in decimal degrees) in the format (F8.4,F9.4) beginning at col. 26. If an initial distance (in km) is to be specified this should be in format (F10.2) in cols. (43-52) of the first line.

Description of output files

outfile1 contains the same information as infile in cols (1-42) and (53-80). Cols (43-52) contain distances along track (in km) in format (F10.2) with a starting distance as in infile.

outfile2 contains auxiliary information (total lengths etc) in self explanatory format.

TIMDEP

Purpose

To produce output suitable for input to GRVINV and MAGINV as described in Section 4.2.

Run command

\$RUN OBJ.TIMDEP+OBJ.FURPAK 1=infile1 2=infile2 7=outfile

Description of input files

infile1 contains, in addition to header information, "10 minute" values of two way interval times through a number of sedimentary layers (that is "thicknesses" in seconds).

Header block formats

Line 1 (I2,2I5)

INTFCE - number of sedimentary layers

ITIM0 - time in minutes from the beginning of the profile, of the first set of interval times. Should be multiple of 2 minutes.

L - number of points along the profile for which the interval times are given.

Line 2 (10F6.2)

V - array of interval velocities for the layers

Main block formats

Each line (I4,10F6.2)

ITIM - time in minutes from the beginning of the profile. Should be a multiple of 2 and increase sequentially by 10.

TZ - array of two way interval times for each of the layers. infile2 contains "2 minute values" of geophysical information in a modified Merged-merged format (such as produced by DIST). It must contain the following information

Main block formats

Each line (25X,F8.4,F9.4,F10.2,5X,F5.3,8X,F5.0,F5.1)

RLAT - latitude in decimal degrees

RLON - longitude in decimal degrees

DIST - distance along track in km

DEP - corrected depth in km

RMAG - residual total field anomaly, in nT

GRAV - free air gravity anomaly, in mgal

Description of output files

outfile contains information in the main block input format lir MAGINV and GRVINV (or SPLINE, if desired) including 10 minute values of latitude, longitude, distance along track, filtered magnetic and gravity anomaly and depths below the sea surface of a number of interfaces (including the sea bottom).

Note

In the case that INTFCE=1, the program is set up to compute thickness as a function of interval time according to the equations of page 67, if V(1) is set to zero.

SPLINE

Purpose

To "spline up" output from TIMDEP to produce data equally spaced in distance.

Run command

\$RUN OBJ.SPLINE+*NAG l=infile 7=outfile

Description of input files

infile is an output file from TIMDEP, containing data equally spaced in time.

Description of output files

outfile has the same number of lines as infile, but the data have been splined to be equally spaced in distance.

Note

This uses the *NAG subroutine E01ADF
(NAGLIB:1673/1489:Mk7:Dec78)

GRVINV, MAGINV

Purpose

To perform direct inversions of gravity and magnetic data, as described in Chapters 4 and 5.

Run command

\$RUN OBJ.GRVINV+OBJ.FURPAK+*GHOST l=infile 9=plotfile

Description of input files

infile is the same for GRVINV and MAGINV; different parts of the header block are dedicated to input for each program.

Header block formats

Line 1 (8A4)

TITLE - title required for plot

Line 2 (3I5)

L - number of points in the profile

INDIR - indicator of profile direction; =1 if profile is to be plotted (L to R) in the same direction that it is read in, =-1 if profile is to be reversed.

ILAYER - for MAGINV; =1 if sea bottom is assumed to be the magnetic basement, =2 if the next interface down is assumed to be magnetic basement.

Line 3 (2F10.4)

BETA1 - average bearing of profile in degrees.

BETA2 - bearing of line perpendicular to structure, onto which data is to be projected. Note: both angles should be for the reversed line if INDIR=-1.

Line 3 (4F10.0) for MAGINV

RI0 - inclination of earth's magnetic field, positive downwards, in degrees.

D0 - declination of earth's field, clockwise from north in degrees.

RIM - inclination of magnetisation, as above

DM - declination of magnetisation, as above

Line 5 (6F10.0) for MAGINV

FV - array of filter values (km^{-1}) corresponding to f_1 , f_2 , f_3 and f_4 (see page 110)

H0 - thickness of magnetic source layer (km)

AV - average value required for magnetisation (A m^{-1})

Line 6 (2I5) for GRVINV

NLYR - number of input interfaces

NINV - number of inverted layers; =1 or 2

Line 7 (I1,4X,7F10.4) for GRVINV

IZ = 0 if inverted layer is upper layer, =1 if inverted layer is lower layer

F3G, F4G - filter values (km^{-1}) corresponding to f_3 , f_4 (see page 81)

ROINV = density contrast (top-bottom) across inverted layer
(Mg m^{-3})

ZB - baselevel for inversion (km)
Line 8 (11,4X,7F10.4) for GRVINV
parameters for second inverted layer, as for line 7. If
NINV=1, leave blank
Line 9 (10F10.4) for GRVINV
DELRO - array of density contrasts (Mg m^{-3}) across known
interfaces. Leave blank if NLIR=0

Main block formats

Each line (F8.4,F9.4,F7.2,F6.0,F6.1,10F6.3)
RLAT - latitude in decimal degrees
RLON - longitude in decimal degrees
DIST - distance along track (km)
RMAG - total field magnetic anomaly (nT)
GOBS - free air gravity anomaly (mgal)
ZZ - array of depths (km) to known interfaces

Description of output files

plotfile suitable for post-processing by, for example,
*DURPLOT. See Figures in Chapters 4 and 5.

Note

THETA1 and THETA2 in GRVINV are the equivalents of BETA1 and
BETA2 in MAGINV. Plotting software is based on the *GHOST
graphical system (Harrison, A.M. (ed), 1977. NUMAC document
U52)

FURPAK

Purpose

FURPAK contains a set of subroutines that are used to perform
simple array computations, Fast Fourier transforms and
frequency domain manipulations. These subroutines are called
by TIMDEP, GRVINV and MAGINV.

Listings of all these programs follow.

C DIST A.G.NUNNS 1978

DIST 1

C
C THE PROGRAMME CALCULATES DISTANCE ALONG SHIPS TRACK USING
C SUBROUTINES PROJ AND FIT
C 4=INPUT FILE CONTAINING TRACK INFORMATION
C 7=OUTPUT FILE CONTAINING TRACK INFORMATION PLUS DISTANCES
C 8=OUTPUT FILE CONTAINING LENGTH OF TRACK AND HEADING
C
C AN INITIAL DISTANCE D0 IS READ FROM THE FIRST LINE OF THE INPUT
C FILE AND ALL DISTANCES ARE ADDED ONTO THIS DISTANCE.
C

```
      IMPLICIT REAL*8(A-H,O-Z)
      DIMENSION X(5000),Y(5000),DIST(5000),A(11),B(7)
      READ(4,400) Y(1),X(1),D0
      DO 5 I5=2,5000
      READ(4,410,END=7) Y(I5),X(I5)
      L=I5
5  CONTINUE
7  CALL PROJ(L,X,Y,BETA1,BETA2)
   CALL FIT(L,X,Y,DIST,C,D)
   CALL BSRF(4,L)
   DO 10 I=1,L
   READ(4,420) A,B
   DISTI=DIST(I)+D0
   WRITE(7,700) A,DISTI,B
10  CONTINUE
   DTOT=DIST(L)-DIST0
   WRITE(8,800) L,DTOT,BETA1,BETA2
   STOP
400 FORMAT(25X,F8.4,F9.4,F10.2)
410 FORMAT(25X,F8.4,F9.4)
420 FORMAT(11A4,8X,7A4)
700 FORMAT(10A4,A2,1X,F9.2,7A4)
800 FORMAT(1X,'NUMBER OF POINTS=',I6,'/',1X,'LENGTH=',F9.2,'/
1,1X,'INITIAL HEADING=',F8.2,'/',1X,'FINAL HEADING=',F9.2)
END
      SUBROUTINE PROJ(L,RLON,RLAT,BETA1,BETA2)
```

C
C A.G.NUNNS 1978
C
C THE SUBROUTINE TAKES AS INPUT TWO ARRAYS RLON AND RLAT WHICH
C CONTAIN VALUES OF LONGITUDE AND LATITUDE FOR A SET OF L POINTS FORMING
C A PROFILE ON THE EARTHS SURFACE.THE PROGRAMME FITS A GREAT CIRCLE
C BETWEEN THE FIRST AND LAST POINTS ON THE PROFILE AND THEN PROJECTS
C EACH POINT ALONG A PERPENDICULAR GREAT CIRCLE ONTO THIS GREAT CIRCLE.
C THE ARRAY RLON IS RETURNED CONTAINING THE DISTANCES IN KILOMETRES
C ALONG THE GREAT CIRCLE FROM THE FIRST POINT TO EACH PROJECTED POINT.
C THE ARRAY RLAT CONTAINS THE DISTANCE IN KILOMETRES BETWEEN THE
C ORIGINAL AND PROJECTED POSITIONS OF EACH POINT. THE SUBROUTINE ALSO
C RETURNS BETA1,THE INITIAL BEARING OF THE GREAT CIRCLE LYING ALONG
C THE PROFILE AND BETA2 THE FINAL BEARING.CORRECTIONS FOR THE CONVERSION
C OF GEOGRAPHIC TO GEOCENTRIC LATITUDE AND FOR THE SURFACE ELLIPTICITY
C ARE MADE.

```
      IMPLICIT REAL*8(A-H,O-Z)
      DIMENSION RLON(L),RLAT(L)
      REAL*8 LAT1,LATL,LATL,LON1,LONL,LONL1,LONI
      DATA A,EPS/6378.16D0,3.3529D-3/
C  CONVERT FROM DEGREES TO RADIANs AND CONVERT FROM GEOGRAPHIC TO
C  GEOCENTRIC LATITUDE.
      DEL=-EPS*DSIN(0.017453D0*(RLAT(1)+RLAT(L)))
      DO 5 I=1,L
      RLON(I)=0.017453D0*RLON(I)
      RLAT(I)=0.017453D0*RLAT(I)+DEL
5  CONTINUE
      LON1=RLON(1)
      LONL=RLON(L)
      LONL1=LONL-LON1
      LAT1=RLAT(1)
      LATL=RLAT(L)
      CLAT1=DCOS(LAT1)
      SLAT1=DSIN(LAT1)
      CLATL=DCOS(LATL)
      SLATL=DSIN(LATL)
      CLONL1=DCOS(LONL1)
      SLONL1=DSIN(LONL1)
      CALPH=CLAT1*CLATL*CLONL1+SLAT1*SLATL
      SALPH=DSQRT(1.D0-CALPH*CALPH)
      C1=SLAT1*CLATL
      C2=CLAT1*CLATL*SLONL1
      C3=CLAT1*SLATL
      RLON(1)=0.D0
      RLAT(1)=0.D0
      R=A*(1.D0-EPS*(DSIN(0.5D0*(LAT1+LATL))**2))
      DO 10 I=2,L
      LATI=RLAT(I)
      LONI=RLON(I)
```

```

      CLATI=DCOS (LATI)
      SLATI=DSIN (LATI)
      DSINY=(C1*CLATI*DSIN (LONI-LONL)+C2*SLATI-C3*CLATI*DSIN (LONI-LONL)
1) /SALPH
      RLON (I)=R*DARCOS ((CLATI*CLATI*DCOS (LONI-LONL)+SLATI*SLATI)/
1DSQRT (1.D0-DSINY*DSINY))
      RLAT (I)=R*DARSIN (DSINY)
10 CONTINUE
      SBETA1=CLATL*SLONL1/SALPH
      SBETA2=CLATL*SLONL1/SALPH
      CBETA1=(SLATL-SLAT1*CALPH)/CLAT1/SALPH
      CBETA2=(SLATL*CALPH-SLAT1)/CLATL/SALPH
      BETA1=57.296*DATAN2 (SBETA1,CBETA1)
      BETA2=57.296*DATAN2 (SBETA2,CBETA2)
      RETURN
      END
      SUBROUTINE FIT (L,X,Y,D,A,B)
C
C A.G.NUNNS
C
C THE SUBROUTINE TAKES AS INPUT TWO ARRAYS X AND Y WHICH CONTAIN THE
C CO-ORDINATES OF L POINTS ALONG A PROFILE.A BEST FIT STRAIGHT LINE TO
C THE PROFILE IS OBTAINED BY MINIMISING THE SUM OF THE SQUARED NORMAL
C DISTANCES OF THE POINTS X(I),Y(I) FROM THE LINE (MAXIMUM LIKELIHOOD
C ESTIMATE).THE SUBROUTINE RETURNS A AND B WHERE Y=AX+B REPRESENTS THE
C LINE.THE ARRAYS X AND Y ARE RETURNED CONTAINING THE CO-ORDINATES OF
C THE POINTS OBTAINED BY PROJECTING THE ORIGINAL POINTS PERPENDICULARLY
C ONTO THE LINE.ALSO RETURNED IS AN ARRAY D WHICH CONTAINS THE DISTANCES
C ALONG THE LINE OF THE PROJECTED POINTS FROM THE PROJECTION OF X(1) AND
C Y(1).
      IMPLICIT REAL*8 (A-H,O-Z)
      DIMENSION X(L),Y(L),D(L)
      SX=0.0D0
      SY=0.0D0
      SX2=0.0D0
      SY2=0.0D0
      SXY=0.0D0
      N=L/100+1
      J=0
      X1=X(1)
      Y1=Y(1)
C NOT ALL POINTS ARE USED.TO MINIMISE ROUNDING ERRORS THE ARRAYS X AND
C Y ARE SAMPLED AT EVERY NTH POINT SO AS TO USE LESS THAN 100 POINTS.
      DO 10 I=1,L,N
      J=J+1
      DELX=X(I)-X1
      DELY=Y(I)-Y1
      SX=SX+DELX
      SY=SY+DELY
      SX2=SX2+DELX*DELX
      SY2=SY2+DELY*DELY
      SXY=SXY+DELX*DELY
10 CONTINUE
      FJ=FLOAT (J)
      E=(SXY-SX*SY/FJ)
      IS=1
      IF (E.LT.0.D0) IS=-1
      F=(SY2-SY*SY/FJ)-SX2+SX*SX/FJ
      G=F/E/2.0D0
      A1=G*G+1
      AINT=IS*DSQRT (A1)
      A2=G*AJNT
      A=G+AJNT
      B=(SY-SX*A)/FJ
      DN=2*(A1+A2)
      DENX=DSQRT (DN)
      DENY=IS*DSQRT (2*(A1-A2))
      DO 20 I=1,L
      XI=X(I) -X1
      YI=Y(I) -Y1
      D(I)=YI/DENY+XI/DENX
      X(I)=(XI+A*(YI-B))/DN+X1
      Y(I)=(A*(XI+A*YI)+B)/DN+Y1
20 CONTINUE
      B=B+Y1-A*X1
      RETURN
      END

```

DIMENSION V(10),TZ(10,1024),RLAT(1024),RLON(1024),DIST(1024),
1DEP(8192),RMAG(8192),GRAV(8192),S(8192),F(8192)

TIMDEP 1

C
C INTFCE IS NO. OF INTERFACES (EXCLUDING WATER-SEDIMENT INTERFACE)
C ITIMO IS THE TIME OF THE FIRST THICKNESS VALUES (IN MINUTES FROM
C THE BEGINNING OF THE PROFILE)
C L IS THE NUMBER OF THICKNESS VALUES GIVEN/INTERFACE AND CORRESPONDS
C TO THE FINAL NUMBER OF RECORDS TO BE PRODUCED IN DEPTH FILE.
C

READ(1,1000) INTFCE,ITIMO,L
1000 FORMAT(I2,2I5)

C CHECK FOR ERRORS

IF(INTFCE.LT.0) WRITE(6,6000)
IF(ITIMO.LT.0) WRITE(6,6100)
IF(L.GT.1024) WRITE(6,6200)

6000 FORMAT(1X,'WARNING INTFCE IS NEGATIVE')
6100 FORMAT(1X,'WARNING ITIMO IS NEGATIVE')
6200 FORMAT(1X,'WARNING L IS GREATER THAN 1024')

C
IF(INTFCE.EQ.0) GOTO 15

C READ IN INTERVAL VELOCITIES

DO 5 J=1,INTFCE
READ(1,1100) V(J)

1100 FORMAT(10F6.2)

C HALVE INTERVAL VELOCITY TO USE WITH TWO WAY TIMES

V(J)=0.5*V(J)

5 CONTINUE

C

IF(V(1).GT.0) GOTO 7
WRITE(6,600)

600 FORMAT(' TYPE IN 1 FOR VARYING K,OTHERWISE RETURN')

READ(5,500) IND

500 FORMAT(I1)

C READ IN TWO-WAY INTERVAL TIMES FOR LAYERS

C

7 ITIM1=ITIMO
DO 10 I=1,L
READ(1,1200) ITIM2,(TZ(J,I),J=1,INTFCE)

1200 FORMAT(I4,10F6.2)

C CHECK THAT TIMES ARE INCREASING BY 10

IF(ITIM2.NE.ITIM1) WRITE(6,6300) ITIM2,ITIM1

6300 FORMAT(1X,'WARNING TIME ERROR;TIME=',I4,' SHOULD BE',I4)

ITIM1=ITIM1+10

10 CONTINUE

C READ IN VALUES FROM MAIN DATA FILE

C SPACE FORWARD TO FIRST RECORD FOR WHICH THERE ARE TWO-WAY TIMES

C AND THEN READ OFF EVERY VALUE OF DEPTH,MAGNETICS AND GRAVITY

C AND EVERY FIFTH VALUE OF LATITUDE,LONGITUDE AND DISTANCE.

C

15 L5=0
CALL FSRF(2,ITIMO/2)

DO 30 I=1,L

L5=L5+1

READ(2,2000) RLAT(I),RLON(I),DIST(I),DEP(L5),RMAG(L5),GRAV(L5)

2000 FORMAT(25X,F8.4,F9.4,F10.2,5X,F5.3,8X,F5.0,F5.1)

DO 20 K=2,5

L5=L5+1

READ(2,2100) DEP(L5),RMAG(L5),GRAV(L5)

2100 FORMAT(57X,F5.3,8X,F5.0,F5.1)

20 CONTINUE

30 CONTINUE

C FIND L2I, INTEGER POWER OF TWO, GREATER THAN OR EQUAL TO 1.25*L5

L2I=1

L5P=L5*5/4

DO 40 K=1,1000

L2I=2*L2I

IF(L2I.GT.L5P) GOTO 50

40 CONTINUE

50 IF(L2I.GT.8192) WRITE(6,6400)

6400 FORMAT(1X,'WARNING L2I GREATER THAN 8192')

C

C NOW RESAMPLE EVERY FIFTH POINT OF DEP,RMAG AND GRAV AFTER FIRST

C APPLYING ANTI-ALIASING FILTER IN THE FREQUENCY DOMAIN.

C

C GENERATE SINE TABLE S, AND ANTI-ALIASING FILTER F

CALL SINTAB(L2I,S)

CALL BDPASS(L2I,1.0,0.0,0.0,0.09,0.10,F,IFAIL)

IF(IFAIL.NE.0) WRITE(6,6500)

6500 FORMAT(1X,'WARNING ERROR IN GENERATING FILTER')

CALL RESAMP(L5,L2I,DEP,S,F)

CALL RESAMP(L5,L2I,RMAG,S,F)

CALL RESAMP(L5,L2I,GRAV,S,F)

C

C NOW COMPUTE DEPTHS TO INTERFACES AND WRITE OUT INTO DEPTH FILE

IF(INTFCE.EQ.0) GOTO 80

DO 70 I=1,L

```

      IF(V(1).LE.0.005) GOTO 52
      TZ(1,I)=DEP(I)+V(1)*TZ(1,I)
      GOTO 55
52    CONTINUE
C     RK=0.6+EXP(1.0-(RLON(I)+9.0))/(1.0+EXP(1.0-8.0*(RLON(I)+9.0)))
      RK=1.0
      TZ(1,I)=DEP(I)+TZ(1,I)*(0.5+0.25*RK*TZ(1,I))
55    IF(INTFCE.EQ.1) GOTO 65
      DO 60 J=2,INTFCE
      TZ(J,I)=TZ(J-1,I)+V(J)*TZ(J,I)
60    CONTINUE
65    WRITE(7,7000) RLAT(I),RLON(I),DIST(I),RMAG(I),GRAV(I),DEP(I),
      1(TZ(J,I),J=1,INTFCE)
7000  FORMAT(F8.4,F9.4,F7.2,F6.0,F6.1,10F6.3)
70    CONTINUE
      STOP
80    DO 90 I=1,L
      WRITE(7,7000) RLAT(I),RLON(I),DIST(I),RMAG(I),GRAV(I),DEP(I)
90    CONTINUE
      STOP
      END
      SUBROUTINE RESAMP(L5,L2I,X,S,F)
C THE SUBROUTINE TAKES AN ARRAY X OF L5 POINTS,APPLIES
C AN ANTI-ALIASING FILTER IN THE TIME DOMAIN AND RESAMPLES
C THE ARRAY AT EVERY FIFTH POINT,RETURNING THE RESAMPLED
C VALUES IN THE SAME ARRAY
      DIMENSION X(L2I),ZERO(8192),S(L2I),F(L2I)
C PAD OUT TO L2I POINTS WITH COSINE JOIN.
      CALL TAPJN(L2I,L5,X)
C FILL ZERO WITH 0.0
      CALL CONST(L2I,ZERO,0.0)
C TRANSFORM TO FREQUENCY DOMAIN
      CALL FURRY(L2I,X,ZERO,1.0,S)
C MULTIPLY REAL AND IMAGINARY PARTS OF TRANSFORM BY FILTER F
      CALL MULT(L2I,X,X,F)
      CALL MULT(L2I,ZERO,ZERO,F)
C TAKE INVERSE TRANSFORM
      CALL FURRY(L2I,X,ZERO,-1.0,S)
C RESAMPLE EVERY FIFTH POINT
      N=0
      DO 10 I=1,L5,5
      N=N+1
      X(N)=X(I)
10    CONTINUE
      RETURN
      END

```

TIMDEP 2

```

      IMPLICIT REAL*8 (A-H,O-Z)
      DIMENSION RLAT(1024),RLON(1024),DIST(1024),DISTS(1024),      SPLINE
      1RMAG(1024),GRAV(1024),Z0(1024),Z1(1024)
      DO 10 I=1,1025
      READ(1,1000,END=20) RLAT(I),RLON(I),DIST(I),
      1RMAG(I),GRAV(I),Z0(I),Z1(I)
1000  FORMAT(F8.4,F9.4,F7.2,F6.0,F6.1,2F6.3)
10    CONTINUE
20    L=I-1
      DX=(DIST(L)-DIST(1))/(L-1)
      DISTS(1)=DIST(1)
      DO 30 I=2,L
      DISTS(I)=DISTS(I-1)+DX
30    CONTINUE
      CALL SPLINE(L,DIST,DISTS,RLAT)
      CALL SPLINE(L,DIST,DISTS,RLON)
      CALL SPLINE(L,DIST,DISTS,RMAG)
      CALL SPLINE(L,DIST,DISTS,GRAV)
      CALL SPLINE(L,DIST,DISTS,Z0)
      CALL SPLINE(L,DIST,DISTS,Z1)
      WRITE(7,1000) (RLAT(I),RLON(I),DISTS(I),RMAG(I),GRAV(I),Z0(I),
      1Z1(I),I=1,L)
      STOP
      END
      SUBROUTINE SPHINE(L,DIST,DISTS,VAL)
      IMPLICIT REAL*8 (A-H,O-Z)
      DIMENSION DIST(L),DISTS(L),VAL(L),
      1W(1024),D(1024),DUM(1024)
      L1=L-1
      DO 10 I=1,L
      CALL EOLADF(L1,DISTS(I),DIST,VAL,W,D,L,DUM(I))
10    CONTINUE
      DO 20 I=1,L
      VAL(I)=DUM(I)
20    CONTINUE
      RETURN
      END

```


GRVINV 1

```

      DIMENSION GJ(1024),GINV(1024),
      1ZB(2),XL1(2),XL2(2),IZ(2)
      COMMON /PLOT/ RLAT(1024),RLON(1024),DIST(1024),
      1ZZ(10,1024),GOBS(1024),DELRO(10),Z(1024),
      2ZINV(1024,2),IL1(2),IL2(2),ZAV(2),ZSD(2),ZINV0(2),
      3GCALC(1024),GRES(1024),NITS(2),DIFFA(2),
      4F3G(2),F4G(2),ROINV(2),ZBASE,TITLE(8)
      COMMON S(1024)
1000  READ(1,1000) TITLE
      FORMAT(8A4)
      READ(1,1100) L,INDIR
1100  FORMAT(2I5)
      READ(1,1150) THETA1,THETA2
1150  FORMAT(2F10.4)
      CALL FSRF(1,2)
      READ(1,1100) NLYR ,NINV
      READ(1,1200) (IZ(JINV),F3G(JINV),F4G(JINV),ROINV(JINV),ZB(JINV)),
      1ZINV0(JINV),XL1(JINV),XL2(JINV),JINV=1,2)
1200  FORMAT(1I,4X,7F10.4)
      IF(NLYR.GT.0) GOTO 5
      CALL FSRF(1,1)
      IF(IABS(INDIR).NE.1) STOP
      IF(INDIR.EQ.-1) II=L
      IF(INDIR.EQ.1) II=1
      DO 3 I=1,L
      READ(1,1300) RLAT(II),RLON(II),DIST(II),GOBS(II)
1300  FORMAT(F8.4,F9.4,F7.2,6X,F6.1,10F6.3)
      II=II+INDIR
      3   CONTINUE
      GOTO 15
      5   READ(1,1400) (DELRO(J),J=1,NLYR)
1400  FORMAT(10F10.4)
      IF(IABS(INDIR).NE.1) STOP
      IF(INDIR.EQ.-1) II=L
      IF(INDIR.EQ.1) II=1
      DO 10 I=1,L
      READ(1,1300) RLAT(II),RLON(II),DIST(II),GOBS(II),
      1(ZZ(J,II),J=1,NLYR)
      II=II+INDIR
10   CONTINUE
15   DX=(DIST(L)-DIST(1))/(L-1)
      DO 16 JINV=1,NINV
      IF(ZINV0(JINV).LE.0) GOTO 16
      IL1(JINV)=(XL1(JINV)-DIST(1))/DX
      IL2(JINV)=(XL2(JINV)-DIST(1))/DX
      IL1(JINV)=MIN0(IL1(JINV),L)
      IL1(JINV)=MAX0(IL1(JINV),1)
      IL2(JINV)=MIN0(IL2(JINV),L)
      IL2(JINV)=MAX0(IL2(JINV),1)
16   CONTINUE
      DX=ABS(DX)*COS(0.017453*(THETA2-THETA1))
      LL=L*5/4
      LX=1
      DO 17 K=1,1000
      LX=2*LX
      IF(LX.GT.LL) GOTO 18
17   CONTINUE
18   CALL SINTAB(LX,S)
      CALL CONST(L,GCALC,0.0)
      CALL SAVE(L,GRES,GOBS)
      IF(NLYR.EQ.0) GOTO 35
      DO 30 J=1,NLYR
      DO 20 I=1,L
      Z(I)=ZZ(J,I)
20   CONTINUE
      CALL TAPJN(LX,L,Z)
      CALL FORWD(LX,DX,DELRO(J),GJ,Z)
      CALL ADDSUB(L,GCALC,GCALC,GJ,1)
      CALL ADDSUB(L,GRES,GRES,GJ,-1)
30   CONTINUE
35   CONTINUE
      DO 50 JINV=1,NINV
      CALL TAPJN(LX,L,GRES)
      IF(ZINV0(JINV).LE.0) GOTO 38
      GAVL=0.
      I1=IL1(JINV)
      I2=IL2(JINV)
      DO 37 I=I1,I2
      GAVL=GAVL+GRES(I)
37   CONTINUE
      GAVL=GAVL/(IL2(JINV)-IL1(JINV))
      ZB(JINV)=ZINV0(JINV)-GAVL/41.91/ROINV(JINV)
38   CALL MEAN(LX,GRES,GAV)
      IF(JINV.EQ.2.AND.IZ(JINV).EQ.0) ZB(2)=ZB(2)-ZB(1)*ROINV(1)
      1/ROINV(2)
      Z0P=ZB(JINV)+GAV/41.91/ROINV(JINV)

```

```

CALL INV(LX,NITS(JINV),DX,ROINV(JINV),F3G(JINV),F4G(JINV),ZOP,
IGRES,ZINV(1,JINV))
CALL MSD(L,ZINV(1,JINV),ZAV(JINV),ZSD(JINV))
CALL FORWD(LX,DX,ROINV(JINV),GINV,ZINV(1,JINV))
GSLAB=-41.91*ROINV(JINV)*ZB(JINV)
DO 40 I=1,LX
GINV(I)=GINV(I)+GSLAB
GRES(I)=GRES(I)+GAV
40 CONTINUE
CALL RMS(L,DIFFA(JINV),GRES,GINV)
CALL ADDSUB(LX,GRES,GRES,GINV,-1)
CALL ADDSUB(LX,GCALC,GCALC,GINV,1)
WRITE(6,6000)JINV,NITS(JINV),DIFFA(JINV),ZAV(JINV),ZSD(JINV)
6000 FORMAT(1X,'INVERTED LAYER NO.',I3,' ',/,
11X,'NO. OF ITERATIONS=',I3,' ',RMS DIFF='F10.2',/,
21X,'ZAV='F5.2,' ',ZSD='F5.2',/)
50 CONTINUE
IF(NINV.EQ.1) ZBASE=ZB(1)
IF(NINV.EQ.2.AND.IZ(1).EQ.0) ZBASE=ZB(1)+ZB(2)*ROINV(2)/ROINV(1)
IF(NINV.EQ.2.AND.IZ(1).EQ.1) ZBASE=ZB(2)+ZB(1)*ROINV(1)/ROINV(2)
WRITE(6,6100) ZBASE
6100 FORMAT(' ZBASE='F5.2)
CALL PLOTG(L,DX,INDIR,NLYR,NINV,THETA1,THETA2)
STOP
END
SUBROUTINE PLOTG(L,DX,INDIR,NLYR,NINV,THETA1,THETA2)
COMMON /PLOT/ RLAT(1024),RLON(1024),DIST(1024),
IZZ(10,1024),GOBS(1024),DELRO(10),Z(1024),
ZZINV(1024,2),IL1(2),IL2(2),ZAV(2),ZSD(2),ZINV0(2),
3GCALC(1024),GRES(1024),NITS(2),DIFFA(2),
4F3G(2),F4G(2),ROINV(2),ZBASE,TITLE(8)
NDIM=1024
IST=1
IST1=2
IEND=L
WRITE(6,600)
600 FORMAT(' TYPE 1 TO INPUT DISTANCE LIMITS,OTHERWISE RETURN')
READ(5,500) IFLAG
500 FORMAT(11)
IF(IFLAG.EQ.0) GOTO 5
WRITE(6,610)
610 FORMAT(' TYPE IN DISTANCE LIMITS FOR PLOT,KM,2F10.0')
READ(5,510) DST,DEND
510 FORMAT(2F10.0)
I1=(1-INDIR)/2*L+(1+INDIR)/2+INDIR*IFIX((DST-DIST(1)))/
1(DIST(L)-DIST(1))*FLOAT(L)+0.5)
I2=(1-INDIR)/2*L+(1+INDIR)/2+INDIR*IFIX((DEND-DIST(1)))/
1(DIST(L)-DIST(1))*FLOAT(L)+0.5)
IST=MIN0(I1,I2)
IST=MAX0(1,IST)
IEND=MAX0(I1,I2)
IEND=MIN0(IEND,L)
L=IEND-IST+1
IST1=IST+1
5 XP1=0.3
XP2=XP1+3.937E-03*ABS(DIST(IEND)-DIST(IST))
CALL PAPER(1)
CALL CSPACE(0.0,XP2+0.1,0.0,1.0)
CALL PSPACE(XP1,XP2,0.0787,0.2362)
CALL MAP(DIST(IST),DIST(IEND),8.0,0.0)
CALL CTRMAG(12)
IF(RLAT(1).EQ.0.0) GOTO 6
CALL PLOTCS(DIST(IST)-INDIR*69.67,10.651,'LATITUDE',8)
CALL PLOTCS(DIST(IST)-INDIR*69.67,11.522,'LONGITUDE',9)
6 CALL PLOTCS(DIST((IST+IEND)/2)-INDIR*15.,9.781,'DISTANCE(KM)',12)
IF(RLAT(1).EQ.0.0) GOTO 8
DO 7 I=IST,IEND,50
CALL PLOTNC(DIST(I),10.20,54)
CALL PLOTNF(DIST(I),10.651,RLAT(I),2)
CALL PLOTNF(DIST(I),11.522,RLON(I),2)
7 CONTINUE
8 CALL PSPACE(XP1,XP2,0.0787,0.5906)
CALL MAP(DIST(IST),DIST(IEND),26.0,0.0)
J=1
9 CALL POSITN(DIST(IST),ZZ(J,IST))
DO 15 I=IST1,IEND
IF(J.EQ.1) GOTO 10
IF(ZZ(J,I).GT.ZZ(J-1,I).OR.ZZ(J,I-1).GT.ZZ(J-1,I-1)) GOTO 10
CALL POSITN(DIST(I),ZZ(J,I))
GOTO 15
10 CALL JOIN(DIST(I),ZZ(J,I))
15 CONTINUE
J=J+1
IF(J.GT.NLYR) GOTO 30
CALL POSITN(DIST(IEND),ZZ(J,IEND))
DO 25 I=IST1,IEND
L11=IEND-I+1

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LI2=LI1+1
IF(ZZ(J,LI1).GT.ZZ(J-1,LI1).OR.ZZ(J,LI2).GT.ZZ(J-1,LI2)) GOTO 20
CALL POSITN(DIST(LI1),ZZ(J,LI1))
GOTO 25
20 CALL JOIN(DIST(LI1),ZZ(J,LI1))
25 CONTINUE
J=J+1
IF(J.GT.NLYR) GOTO 30
GOTO 9
30 CONTINUE
DO 32 JINV=1,NINV
CALL SAVE(L,Z(IST),ZINV(IST,JINV))
CALL PTPLOT(DIST,Z,IST,IEND,-2)
IF(ZINV0(JINV).LE.0.) GOTO 32
CALL POSITN(DIST(IL1(JINV)),ZINV0(JINV))
CALL BROKEN(10,10,10,10)
CALL JOIN(DIST(IL2(JINV)),ZINV0(JINV))
CALL FULL
32 CONTINUE
35 CALL SCALES
CALL BORDER
CALL MINMAX(L,GOBS(IST),GOMIN,GOMAX)
CALL MINMAX(L,GCALC(IST),GCMIN,GCMAX)
CALL MINMAX(L,GRES(IST),GRMIN,GRMAX)
TMIN=AMIN1(GOMIN,GCMIN,GRMIN)
TMAX=AMAX1(GOMAX,GCMAX,GRMAX)
TMIN=TMIN-0.1*ABS(TMIN)
TMAX=TMAX+0.1*ABS(TMAX)
CALL PSPACE(XP1,XP2,0.6299,0.9843)
CALL MAP(DIST(IST),DIST(IEND),TMIN,TMAX)
CALL PTPLOT(DIST,GCALC,IST,IEND,-2)
CALL PTPLOT(DIST,GOBS,IST,IEND,43)
CALL PTPLOT(DIST,GRES,IST,IEND,-2)
CALL POSITN(DIST(IEND),0.0)
CALL JOIN(DIST(IST),0.0)
CALL SCALES
CALL BORDER
CALL CSPACE(0.0206,0.3,0.0,1.0)
CALL PLACE(1,4)
CALL TYPECS('FREE AIR ANOMALY (MGAL)',23)
CALL PLACE(1,7)
CALL TYPECS('OBSERVED AS CROSSES',19)
CALL CRLNFD
CALL TYPECS('CALCULATED AND RESIDUAL',23)
CALL CRLNFD
CALL TYPECS('AS SOLID LINES',14)
CALL PLACE(1,28)
CALL TYPECS('DEPTH (KM)',10)
CALL PSPACE(XP2+0.0394,XP2+0.3543,0.0787,0.9843)
CALL BORDER
CALL CSPACE(XP2+0.0600,XP2+0.3543,0.0,1.0)
CALL PLACE(1,3)
CALL TYPECS(TITLE,32)
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('AVERAGE BEARING',15)
CALL CRLNFD
CALL TYPECS('OF PROFILE=',12)
CALL TYPENF(THETA1,2)
CALL CTRSET(2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CTRSET(1)
CALL CRLNFD
CALL CRLNFD
IF(THETA1.EQ.THETA2) GOTO 37
CALL TYPECS('DATA PROJECTED ONTO',19)
CALL CRLNFD
CALL TYPECS('LINE BEARING',12)
CALL TYPENF(THETA2,2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CTRSET(1)
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('PROJECTED',9)
CALL CRLNFD
37 CALL TYPECS('SAMPLE SPACING=',15)
CALL TYPENF(DX,3)
CALL CTRSET(2)
CALL TYPECS(' KM',3)
CALL CTRSET(1)
CALL CRLNFD
CALL CRLNFD
IF(NLYR.EQ.0) GOTO 50

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GRVINV 3

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CALL TYPECS('DENSITY CONTRAST ACROSS',24)
CALL CRLNFD
CALL TYPECS('KNOWN INTERFACES:',17)
DO 40 J=1,NLYR
CALL CRLNFD
CALL HLINFD(1)
CALL CTRSET(3)
CALL TYPENC(14)
CALL CTRSET(4)
CALL TYPENC(28)
CALL SUFFIX
CALL TYPENI(J)
CALL NORMAL
CALL CTRSET(2)
CALL TYPENC(47)
CALL TYPENF(DELO(J),3)
CALL TYPECS('*10',3)
CALL SUFFIX
CALL TYPECS('3',1)
CALL NORMAL
CALL TYPECS(' KG/M',5)
CALL SUFFIX
CALL TYPECS('3',1)
CALL NORMAL
40 CONTINUE
50 CALL CTRSET(1)
DO 60 JINV=1,NINV
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('INVERTED LAYER',14)
CALL TYPENI(JINV)
CALL CRLNFD
CALL HLINFD(1)
CALL TYPECS('MEAN DEPTH=',11)
CALL TYPENF(ZAV(JINV),1)
CALL CTRSET(2)
CALL TYPECS(' KM',3)
CALL CTRSET(1)
CALL CRLNFD
CALL TYPECS('S.D. OF DEPTH=',14)
CALL TYPENF(ZSD(JINV),1)
CALL CTRSET(2)
CALL TYPECS(' KM',3)
CALL CTRSET(1)
CALL CRLNFD
CALL HLINFD(1)
CALL CTRSET(3)
CALL TYPENC(14)
CALL CTRSET(4)
CALL TYPENC(28)
CALL CTRSET(2)
CALL TYPENC(47)
CALL TYPENF(ROINV(JINV),3)
CALL TYPECS('*10',3)
CALL SUFFIX
CALL TYPECS('3',1)
CALL NORMAL
CALL TYPECS(' KG/M',5)
CALL SUFFIX
CALL TYPECS('3',1)
CALL NORMAL
CALL CTRSET(1)
CALL HLINFD(1)
CALL CRLNFD
CALL TYPECS('FILTER VALUES',13)
CALL CRLNFD
CALL TYPECS('F3=',3)
CALL TYPENF(F3G(JINV),4)
CALL CTRSET(2)
CALL TYPECS(' KM',3)
CALL CTRSET(1)
CALL SUFFIX
CALL TYPECS('-1',2)
CALL NORMAL
CALL CRLNFD
CALL TYPECS('F4=',3)
CALL TYPENF(F4G(JINV),4)
CALL CTRSET(2)
CALL TYPECS(' KM',3)
CALL CTRSET(1)
CALL SUFFIX
CALL TYPECS('-1',2)
CALL NORMAL
CALL CRLNFD
CALL HLINFD(1)
CALL TYPECS('NO. OF ITERATIONS=',18)
CALL TYPENI(NITS(JINV))

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GRVINV 4

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CALL CRLNFD
CALL TYPECS('RMS RESIDUAL=',13)
CALL TYPENF(DIFFA(JINV),2)
CALL CTRSET(2)
CALL TYPECS('MGAL',5)
CALL CTRSET(1)
60 CONTINUE
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('BASE LEVEL=',11)
CALL TYPENF(ZBASE,1)
CALL CTRSET(2)
CALL TYPECS('KM',3)
CALL CTRSET(1)
CALL GREND
RETURN
END
SUBROUTINE INV(LX,NITS,DX,ROCM,F3G,F4G,Z0,G,ZINV)
DIMENSION G(LX),ZINV(1024),FGR(1024),FGI(1024),RK(1024),F(1024),
1H(1024),EKZOP(1024),FHOR(1024),FHOI(1024),SUMR(1024),SUMI(1024)
COMMON S(1024)
CALL SAVE(LX,FGR,G)
CALL CONST(LX,FGI,0.0)
CALL FURRY(LX,FGR,FGI,1.0,S)
CALL SCALE(LX,FGR,-1.0/41.91/ROCM,FGR)
CALL SCALE(LX,FGI,-1.0/41.91/ROCM,FGI)
DK=6.2831853/DX/FLOAT(LX)
RK(1)=0.0
LX21=LX/2+1
DO 10 I=2,LX21
RK(I)=RK(I-1)+DK
10 CONTINUE
CALL EVEN(LX,RK)
CALL BDPASS(LX,1.0/DX,0.0,0.0,F3G,F4G,F,IFAIL)
ZOP=Z0
DELZ=0.0
CALL CONST(LX,H,0.0)
DIFF0=30.0
DIFFMN=0.5E-03
NITER=50
DO 50 ITER=1,NITER
NITS=ITER
DEKZOP=EXP(DK*ZOP)
EKZOP(1)=1.0
DO 20 I=2,LX21
EKZOP(I)=EKZOP(I-1)*DEKZOP
20 CONTINUE
CALL EVEN(LX,EKZOP)
CALL MULT(LX,FHOR,FGR,EKZOP)
CALL MULT(LX,FHOI,FGI,EKZOP)
FHOR(1)=FHOR(1)+LX*DELZ
IF(ITER.EQ.1) GOTO 25
CALL SUMG(LX,RK,H,SUMR,SUMI)
CALL ADDSUB(LX,FHOR,FHOR,SUMR,-1)
CALL ADDSUB(LX,FHOI,FHOI,SUMI,-1)
25 CALL MULT(LX,FHOR,FHOR,F)
CALL MULT(LX,FHOI,FHOI,F)
CALL FURRY(LX,FHOR,FHOI,-1.0,S)
CALL RMS(LX,DIFF1,FHOR,H)
CALL SAVE(LX,H,FHOR)
IF(DIFF1.LE.DIFF0) GOTO 30
WRITE(6,6000) ITER
6000 FORMAT(' AFTER',I3,' ITERATIONS DIVERGENCE IS OCCURRING',/,
1' TYPE 1 FOR CONTINUATION,0 TO STOP')
READ(5,5000) INDIT
5000 FORMAT(I1)
IF(INDIT.EQ.0) GOTO 60
30 IF(DIFF1.LT.DIFFMN) GOTO 60
DIFF0=DIFF1
CALL MINMAX(LX,H,HMIN,HMAX)
DELH=-(HMIN+HMAX)/2.0
ZOP=ZOP+DELH
DELZ=ZOP-Z0
DO 40 I=1,LX
H(I)=H(I)+DELH
40 CONTINUE
50 CONTINUE
60 DO 70 I=1,LX
ZINV(I)=ZOP-H(I)
70 CONTINUE
RETURN
END
SUBROUTINE FORWD(LX,DX,RO,GR,Z)
DIMENSION GR(LX),Z(LX),H(1024),RK(1024),EKZ0(1024),GI(1024),
1SUMR(1024),SUMI(1024)
COMMON S(1024)
CALL MINMAX(LX,Z,ZMIN,ZMAX)

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GRVINV 5

GRVINV 6

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Z0=(ZMIN+ZMAX)/2.0
DO 10 I=1,LX
H(I)=Z0-Z(I)
10 CONTINUE
DK=6.2831853/DX/FLOAT(LX)
DEKZ0=EXP(-DK*Z0)
RK(1)=0.0
EKZ0(1)=-41.91*RO
LX21=LX/2+1
DO 20 I=2,LX21
RK(I)=RK(I-1)+DK
EKZ0(I)=EKZ0(I-1)*DEKZ0
20 CONTINUE
CALL EVEN(LX,RK)
CALL EVEN(LX,EKZ0)
CALL SAVE(LX,GR,H)
CALL CONST(LX,GI,0.0)
CALL FURRY(LX,GR,GI,1.0,S)
CALL SUMG(LX,RK,H,SUMR,SUMI)
CALL ADDSUB(LX,GR,GR,SUMR,1)
CALL ADDSUB(LX,GI,GI,SUMI,1)
CALL MULT(LX,GR,GR,EKZ0)
CALL MULT(LX,GI,GI,EKZ0)
GR(1)=GR(1)+41.91*LX*RO*Z0
CALL FURRY(LX,GR,GI,-1.0,S)
RETURN
END
SUBROUTINE SUMG(LX,RK,H,SUMR,SUMI)
DIMENSION RK(LX),H(LX),SUMR(LX),SUMI(LX),RKN(1024),HNR(1024),
1FHNR(1024),FHNI(1024),TERMR(1024),TERMI(1024)
COMMON S(1024)
CALL SCALE(LX,RKN,0.5,RK)
CALL MULT(LX,HNR,H,H)
CALL SAVE(LX,FHNR,HNR)
CALL CONST(LX,FHNI,0.0)
CALL FURRY(LX,FHNR,FHNI,1.0,S)
CALL MULT(LX,TERMR,RKN,FHNR)
CALL MULT(LX,TERMI,RKN,FHNI)
CALL MINMAX(LX,TERMR,TMINR,TMAXR)
CALL MINMAX(LX,TERMI,TMINI,TMAXI)
S2=AMAX1(ABS(TMINR),ABS(TMAXR),ABS(TMINI),ABS(TMAXI))
CALL SAVE(LX,SUMR,TERMR)
CALL SAVE(LX,SUMI,TERMI)
IF(S2.EQ.0) RETURN
NSUM=30
EPS=0.5E-02
DO 100 N=3,NSUM
CALL MULT(LX,HNR,HNR,H)
CALL SAVE(LX,FHNR,HNR)
CALL CONST(LX,FHNI,0.0)
CALL MULT(LX,RKN,RKN,RK)
CALL SCALE(LX,RKN,1.0/FLOAT(N),RKN)
CALL FURRY(LX,FHNR,FHNI,1.0,S)
CALL MULT(LX,TERMR,RKN,FHNR)
CALL MULT(LX,TERMI,RKN,FHNI)
CALL ADDSUB(LX,SUMR,SUMR,TERMR,1)
CALL ADDSUB(LX,SUMI,SUMI,TERMI,1)
CALL MINMAX(LX,TERMR,TMINR,TMAXR)
CALL MINMAX(LX,TERMI,TMINI,TMAXI)
SN=AMAX1(ABS(TMINR),ABS(TMAXR),ABS(TMINI),ABS(TMAXI))
EPSN=SN/S2
IF(EPSN.LT.EPS) RETURN
100 CONTINUE
WRITE(6,6000) NSUM,EPSN
6000 FORMAT(' WARNING: AFTER ',I3,' TERMS, SUM HAS NOT CONVERGED',/,
1' EPSN=',E10.2,' ,RETURNING FROM SUBROUTINE SUMG')
RETURN
END

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C *****
C PROGRAM MAGINV
C *****
C A.G.NUNNS 1978
C THIS PROGRAM PERFORMS AN INVERSION OF TOTAL FIELD MAGNETIC ANOMAL-
C IES USING THE PROCEDURE GIVEN IN R.L.PARKER AND S.P.HUESTIS,1974
C (THE INVERSION OF MAGNETIC ANOMALIES IN THE PRESENCE OF TOPOGRAPHY
C JGR VOL.79,NO.11,1587-1593.)
C
C IT IS ASSUMED THAT THE MAGNETIC ANOMALY ARISES FROM A "TWO-DIMENS-
C IONAL" STRUCTURE AND THAT THE ANOMALY IS KNOWN AT EQUALLY SPACED
C POINTS ALONG A PROFILE PERPENDICULAR TO THIS STRUCTURE.IT IS ALSO
C ASSUMED THAT ALL MAGNETISATION IS CONFINED TO A LAYER OF CONSTANT
C THICKNESS WHOSE UPPER SURFACE IS KNOWN AT EACH POINT ON THE PROFILE
C AND THAT THE DIRECTION OF THE MAGNETISATION IS CONSTANT AND KNOWN.
C SUBJECT TO THESE CONSTRAINTS THE PROGRAM ATTEMPTS TO COMPUTE THE
C MAGNETISATION MAGNITUDE AT EACH POINT ALONG THE PROFILE.BECAUSE
C OF A LACK OF UNIQUENESS IN THE INVERSE PROBLEM(SEE PARKER&HUESTIS)
C TWO SEPARATE MAGNITUDES ARE COMPUTED;THE FIRST IS THE ZERO MEAN
C MAGNETISATION,I.E.THE MAGNITUDE COMPUTED SUBJECT TO THE CONSTRAINT
C THAT ITS MEAN LEVEL OVER THE WHOLE PROFILE IS ZERO (WHICH MAKES IT
C UNIQUE) AND THE SECOND IS THE ANNIHILATOR MAGNETISATION WHICH HAS
C UNIT MEAN BUT CONTRIBUTES NOTHING TO THE MAGNETIC ANOMALY DUE TO
C THE CANCELLATION EFFECT OF THE TOPOGRAPHY.THE ANNIHILATOR MAGNET-
C ISATION CAN BE MULTIPLIED BY A SCALAR AND ADDED TO THE ZERO MEAN
C MAGNETISATION TO GIVE A MAGNETISATION WITH A GEOPHYSICALLY
C FEASIBLE MEAN WHICH STILL YIELDS THE SAME ANOMALY.
C
C
COMMON /MAITSU/ RMR(1024),RMI(1024),SUMR(1024),SUMI(1024),S(1024)
1,FMR(1024),FMI(1024)
COMMON /MAIT/ FL(1024),FMOR(1024),FMOI(1024),EKHT(1024)
COMMON /MASU/ H(1024),RK(1024)
DIMENSION A(1024),AI(1024),FAR(1024),FAI(1024),ZMMR(1024)
1,ZMMI(1024),W(1024),DIST(1024),EKH0(1024),ACOMR(1024),ACOMI(1024),
2,RMOR(1024),FACOMR(1024),FACOMI(1024),RLAT(1024),RLON(1024),
3,Z(1024),FH(1024),FV(4),TITLE(8)
EQUIVALENCE (H(1),Z(1)),(FH(1),FL(1))
C
C FOR DETAILS OF INPUT SEE SUBROUTINE READ
CALL READ(L,INDIR,LX,DX,BETA1,BETA2,RIO,D0,RIM,DM,FV,H0
1,AV,RLAT,RLON,DIST,A,Z,TITLE)
C
C THE ARRAY Z(I),I=1,2,...,L CONTAINS DEPTHS TO THE TOP
C SURFACE OF THE MAGNETIC LAYER.FOLLOWING PARKER(1974) THE OPTIMUM
C CONVERGENCE OF THE INFINITE SUM IS OBTAINED WITH A DATUM HALFWAY
C BETWEEN THE TOPOGRAPHIC EXTREMES,AT DEPTH ZMED.THE ARRAY H IS
C DEFINED SO THAT IT CONTAINS HEIGHTS MEASURED FROM THIS DATUM.THE
C OPTIMUM CONVERGENCE OF THE ITERATIVE SEQUENCE IS OBTAINED WITH A
C DATUM JUST ABOVE THE HIGHEST POINT OF THE MAGNETIC LAYER,AT DEPTH
C Z0 BELOW THE OBSERVATION SURFACE.THE DIFFERENCE BETWEEN THE TWO
C DATUMS IS HTILDE.
C PAD OUT DEPTHS TO LX VALUES USING SUBROUTINE TAPJN.
CALL TAPJN(LX,L,Z)
CALL MINMAX(LX,Z,ZMIN,ZMAX)
ZMED=(ZMIN+ZMAX)/2.0
HTILDE=(ZMAX-ZMIN)/1.98
Z0=ZMED-HTILDE
DO 100 I100=1,LX
H(I100)=ZMED-Z(I100)
100 CONTINUE
C PAD OUT ARRAY OF MAGNETIC ANOMALIES A(I) TO LX ELEMENTS
C AND REMOVE MEAN VALUE.
CALL TAPJN(LX,L,A)
CALL MEAN(LX,A,AAV)
C CREATE TABLE OF SINE VALUES S.
CALL SINTAB(LX,S)
C
CALL SAVE(LX,FAR,A)
CALL CONST(LX,FAI,0.0)
CALL FURRY(LX,FAR,FAI,1.0,S)
IF(FV(1).EQ.0.0.AND.FV(2).EQ.0.0) GOTO 150
CALL BDPASS(LX,1.0/DX,FV(1),FV(2),0.5/DX,0.5/DX,FH,IFAIL)
CALL MULT(LX,FAR,FAR,FH)
CALL MULT(LX,FAI,FAI,FH)
CALL SAVE(LX,A,FAR)
CALL SAVE(LX,AI,FAI)
CALL FURRY(LX,A,AI,-1.0,S)
150 CALL BDPASS(LX,1.0/DX,0.0,0.0,FV(3),FV(4),FL,IFAIL)
C CALCULATE REAL AND IMAGINARY PARTS OF THE COMPLEX NUMBER V,
C (SEE EQUATION (3),PARKER & HUESTIS,1974)AND ITS INVERSE VINV
C (EARTHS FIELD),D,(THE DECLINATION OF THE EARTHS FIELD WRT THE
C POSITIVE PROFILE DIRECTION) AND RIP,(THE INCLINATION OF THE
C MAGNETISATION IN A VERTICAL PLANE CONTAINING THE PROFILE).
C CONVERT DEGREES TO RADIANS
RIO=0.017453*RIO

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RIM=0.017453*RIM
D0=0.017453*D0
DM=0.017453*DM
C10=COS(RI0)
S10=SIN(RI0)
CIM=COS(RIM)
SIM=SIN(RIM)
CICD0=C10*COS(D0)
CICDM=CIM*COS(DM)
C EVALUATE REAL AND IMAGINARY PARTS OF V.
  VR=S10*SIM-CICD0*CICDM
  VI=CICD0*SIM+CICDM*S10
C EVALUATE REAL AND IMAGINARY PARTS OF 1/V.
  VMOD2=VR*VR+VI*VI
  VINVR=VR/VMOD2
  VINVI=VI/VMOD2
C INCORPORATE CONSTANT 10**9*MU0/2, TAKING CARE OF PERMEABILITY
C OF FREE SPACE AND THE CONVERSION FROM NANOTESLA TO TESLA, SO THAT
C FINAL MAGNETISATION VALUES WILL BE IN AMPS/M.
  VR=628.31853*VR
  VI=628.31853*VI
  VINVR=1.59155E-03*VINVR
  VINVI=1.59155E-03*VINVI
C CALCULATE THE INITIAL ESTIMATE OF THE TRANSFORM OF THE MAGNETISATION,
C GIVEN BY THE FIRST TERM ON THE RHS OF EQU.4 (PARKER, 1974). THE REAL
C PART OF THIS TRANSFORM IS IN THE ARRAY FM0R(I), THE IMAGINARY IN
C FM0I(I). THE ARRAYS, RK(I), I=1, 2, ..., LX OF SPATIAL WAVENUMBERS AND
C EKHT(I) CONSISTING OF ADJUSTMENT FACTORS EXP(-K*TILDE) ARE ALSO
C CALCULATED THIS PROGRAM SEGMENT.
C DK IS THE CHANGE IN SPATIAL FREQUENCY FOR AN INDEX INCREMENT OF
C 1,  $DK=2*PI/DX/LX$  WHERE DX IS THE ORIGINAL SAMPLING INTERVAL IN M.
  DK=6.2831853/DX/FLOAT(LX)
C DEKZ0, DEKH0 AND DEKHT ARE THE MULTIPLICATIVE CHANGES IN
C  $EXP(K*Z0)$ ,  $EXP(-K*H0)$ , (WHERE H0 IS THE INPUT THICKNESS OF THE
C MAGNETIC LAYER IN M) AND  $EXP(-K*HTILDE)$  RESPECTIVELY,
C CORRESPONDING TO AN INDEX INCREMENT OF 1.
  DEKZ0=EXP(DK*Z0)
  DEKH0=EXP(-DK*H0)
  DEKHT=EXP(-DK*HTILDE)
  EKZ0=1.0
  EKH0(1)=1.0
  EKHT(1)=1.0
C DC LEVEL OF MAGNETISATION ESTIMATE IS ZERO.
  FM0R(1)=0.0
  FM0I(1)=0.0
  RK(1)=0.0
C FREQUENCIES ABOVE THE NYQUIST FREQUENCIES (INDEX=LX/2+1) ARE
C TAKEN CARE OF BY THE INDEX LXI2=LX+I-2.
  LXI2=LX
  NN=LX/2
  DO 200 I200=2, NN
    RK(I200)=RK(I200-1)+DK
    RK(LXI2)=RK(I200)
    EKZ0=EKZ0*DEKZ0
    EKH0(I200)=DEKH0*EKH0(I200-1)
    EKHT(I200)=DEKHT*EKHT(I200-1)
    EKHT(LXI2)=EKHT(I200)
C FACT1 IS THE FACTOR  $EXP(K*Z0)/(1-EXP(-K*H0))$  IN EQU.4
  FACT1=EKZ0/(1.0-EKH0(I200))
  FM0R(I200)=FACT1*(VINVR*FAR(I200)-VINVI*FAI(I200))
  FM0I(I200)=FACT1*(VINVR*FAI(I200)+VINVI*FAR(I200))
  FM0R(LXI2)=FACT1*(VINVR*FAR(LXI2)+VINVI*FAI(LXI2))
  FM0I(LXI2)=FACT1*(VINVR*FAI(LXI2)-VINVI*FAR(LXI2))
  LXI2=LXI2-1
200 CONTINUE
C TIDY UP VALUES AT NYQUIST FREQUENCY.
  NN1=NN+1
  RK(NN1)=RK(NN)+DK
  EKH0(NN1)=DEKH0*EKH0(NN)
  EKHT(NN1)=EKHT(NN)*DEKHT
  FACT1=DEKZ0*EKZ0/(1.0-EKH0(NN1))
  FM0R(NN1)=FACT1*(VINVR*FAR(NN1)-VINVI*FAI(NN1))
  FM0I(NN1)=0.0
C
C MULTIPLY FOURIER TRANSFORM FM0<R, I> BY THE BANDPASS FILTER F
C AND CREATE INVERSE TRANSFORM, THAT IS, THE FIRST ESTIMATE OF THE
C MAGNETISATION, RM<R, I>(I), I=1, 2, ..., LX. RMI(I) SHOULD BE ZERO.
  CALL MULT(LX, RMR, FM0R, FL)
  CALL MULT(LX, RMI, FM0I, FL)
  CALL FURRY(LX, RMR, RMI, -1.0, S)
  CALL SAVE(LX, RMOR, RMR)
C CALL ITRATE TO CALCULATE ZERO MEAN MAGNETISATION ESTIMATE, REAL
C PART ZMMR(I), IMAGINARY PART ZMMI(I).
  CALL ITRATE(LX, NITSM, DIFF0)
  CALL SAVE(LX, ZMMR, RMR)
  CALL SAVE(LX, ZMMI, RMI)
  WRITE(6, 600) NITSM, DIFF0

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600 FORMAT(/,1X,'COMPUTATION OF ZERO MEAN MAGNETISATION FINISHED'
1,/, 'AFTER',I3,' ITERATIONS,RMS DIFFERENCE IS',E10.2)
C
C NOW COMPUTE ANOMALY CORRESPONDING TO INVERTED MAGNETISATION
C AND COMPARE IT WITH ORIGINAL ANOMALY
CALL SUM(LX)
CALL SAVE(LX,FACOMR,SUMR)
CALL SAVE(LX,FACOMI,SUMI)
C DEKZM IS MULTIPLICATIVE CHANGE IN EXP(-K*ZMED) FOR AN INDEX
C INCREMENT OF 1.
DEKZM=EXP(-DK*ZMED)
EKZM=1.0
ACOMR(1)=0.0
ACOMI(1)=0.0
C MULTIPLY INFINITE SUM BY FACTOR GIVEN IN EQU.1 TO GIVE FOURIER
C TRANSFORM OF COMPUTED ANOMALY
LXI2=LX
DO 300 I300=2,NN
EKZM=DEKZM*EKZM
C FACT2 IS FACTOR EXP(-K*ZMED)*(1-EXP(-K*H0))
FACT2=EKZM*(1.0-EKH0(I300))
ACOMR(I300)=FACT2*(VR*FACOMR(I300)+VI*FACOMI(I300))
ACOMI(I300)=FACT2*(VR*FACOMI(I300)-VI*FACOMR(I300))
ACOMR(LXI2)=FACT2*(VR*FACOMR(LXI2)-VI*FACOMI(LXI2))
ACOMI(LXI2)=FACT2*(VR*FACOMI(LXI2)+VI*FACOMR(LXI2))
LXI2=LXI2-1
300 CONTINUE
FACT2=DEKZM*EKZM*(1.0-EKH0(NN1))
ACOMR(NN1)=FACT2*(VR*FACOMR(NN1)+VI*FACOMI(NN1))
ACOMI(NN1)=0.0
C PERFORM INVERSE TRANSFORM TO GIVE COMPUTED ANOMALY ACOM<R,I>(I)
CALL FURRY(LX,ACOMR,ACOMI,-1.0,S)
C FIND AND PRINT OUT DIFFERENCE BETWEEN ACTUAL AND COMPUTED ANOMALY.
CALL RMS(LX,DIFFA,A,ACOMR)
WRITE(6,605) DIFFA
605 FORMAT(1X,'RMS DIFFERENCE BETWEEN ACTUAL ANOMALY AND',/,1X,
1'RECOMPUTED ANOMALY IS',F8.2)
C
C NOW COMPUTE ANNIHILATOR MAGNETISATION.
C FMOR(I) FOR ANNIHILATOR IS SIMPLY A KRONECKER DELTA,FMOI(I) IS ZERO
CALL CONST(LX,FMOR,0.0)
CALL CONST(LX,FMOI,0.0)
FMOR(1)=LX
CALL CONST(LX,RMR,1.0)
CALL CONST(LX,RMI,0.0)
C ANNIHILATOR MAGNETISATION HAS REAL PART RMR(I),IMAGINARY,RMI(I).
CALL ITRATE(LX,NITSA,DIFF0)
WRITE(6,610) NITSA,DIFF0
610 FORMAT(/, 'COMPUTATION OF ANNIHILATOR MAGNETISATION FINISHED'
1,/, 'AFTER',I3,' ITERATIONS,RMS DIFFERENCE IS',E10.2)
C ADD AV*RMR TO ZMMR TO ADJUST MEAN VALUE OF MAGNETISATION TO AV.
IF(AV.EQ.0.0) GOTO 500
DO 400 I400=1,LX
ZMMR(I400)=ZMMR(I400)+AV*RMR(I400)
400 CONTINUE
C PLOT OUT RESULTS
500 CALL PLOTM(L,INDIR,DX,RIO,D0,RIM,DM,H,A,FV,H0,ZMED,ACOMR,
1ZMMR,RMR,DIST,AV,DIFFA,BETA1,BETA2,NITSM,NITSA,RLAT,RLON,TITLE)
STOP
END
SUBROUTINE ITRATE(LX,NITS,DIFF0)
C THE SUBROUTINE PERFORMS THE ITERATIVE CALCULATION OF MAGNETISATION
C GIVEN BY EQUATION (4) OF PARKER(1974).ON ENTRY TO THE SUBROUTINE
C THE FOLLOWING ARRAYS ARE DEFINED IN COMMON:
C RM<R,I>(I),REAL AND IMAGINARY PARTS OF FIRST MAGNETISATION ESTIMATE.
C FM0<R,I>(I),REAL AND IMAGINARY PARTS OF UNFILTERED TRANSFORM
C OF FIRST ESTIMATE OF MAGNETISATION.
C F(I),BANDPASS FILTER USED TO SMOOTH MAGNETISATION ESTIMATES.
C H(I),ARRAY OF TOPOGRAPHY VALUES.
C RK(I),ARRAY OF SPATIAL WAVENUMBER VALUES.
C EKHT(I),DATUM LEVEL ADJUSTMENT ARRAY.
C S(I),THE ARRAY OF SINE VALUES.
C THE ARRAYS RM<R,I>(I) REPRESENTING THE REAL AND IMAGINARY PARTS
C OF THE FINAL MAGNETISATION ESTIMATE ARE DEFINED IN COMMON ON
C RETURN.
C
C
COMMON /MAITSU/ RMR(1024),RMI(1024),SUMR(1024),SUMI(1024),S(1024)
1,FMR(1024),FMI(1024)
COMMON /MAIT/ FL(1024),FMOR(1024),FMOI(1024),EKHT(1024)
DIMENSION RMNR(1024),RMNI(1024)
C ENTER MAJOR ITERATION LOOP,WITH INDEX ITER.EXIT FROM THE
C LOOP OCCURS IF;(1)THE NUMBER OF ITERATIONS EXCEEDS THE SET
C VALUE NITER,(2)THE RMS DIFFERENCE BETWEEN TWO SUCCESSIVE
C MAGNETISATION ESTIMATES IS DIVERGING,OR IF (3)THE RMS DIFFERENCE
C CONVERGES TO LESS THAN A SET VALUE DIFFMN.
DIFFMN=5.0E-03

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      NITER=100
C   INITIAL RMS VALUE SET TO 20 A/M
      DIFF0=20.0
      CALL SAVE(LX,FMR,FMOR)
      CALL SAVE(LX,FMI,FMOI)
C
C
      DO 200 ITER=1,NITER
      NITS=ITER
C
C   CALL SUBROUTINE SUM TO EVALUATE INFINITE SUM IN EQU. 4.
      CALL SUM(LX)
C   MULTIPLY ARRAY SUM<R,I>(I) BY ADJUSTMENT ARRAY EKHT(I) TO TAKE
C   CARE OF HEIGHT DIFFERENCE HTILDE BETWEEN DATUM LEVELS.
      CALL MULT(LX,SUMR,SUMR,EKHT)
      CALL MULT(LX,SUMI,SUMI,EKHT)
      CALL ADDSUB(LX,SUMR,SUMR,FMR,-1)
      CALL ADDSUB(LX,SUMI,SUMI,FMI,-1)
C   SUBTRACT ADJUSTED ARRAY SUM<R,I>(I) FROM FMO<R,I>(I) TO GIVE
C   NEW UNFILTERED ESTIMATE OF TRANSFORM OF MAGNETISATION.
      CALL ADDSUB(LX,FMR,FMOR,SUMR,-1)
      CALL ADDSUB(LX,FMI,FMOI,SUMI,-1)
C   MULTIPLY BY FILTER F(I) AND PERFORM INVERSE TRANSFORMATION.
      CALL MULT(LX,FMR,FMR,FL)
      CALL MULT(LX,FMI,FMI,FL)
      CALL SAVE(LX,RMNR,FMR)
      CALL SAVE(LX,RMNI,FMI)
      CALL FURRY(LX,RMNR,RMNI,-1.0,S)
C   COMPUTE RMS DIFFERENCE DIFF1 BETWEEN PRESENT AND PREVIOUS
C   ESTIMATES OF MAGNETISATION.
      CALL RMS(LX,DIFF1,RMR,RMNR)
C   CONTRAST DIFF1 WITH PREVIOUS RMS DIFFERENCE,DIFF0,TO SEE IF
C   ITERATIONS ARE CONVERGING.IF THEY ARE DIVERGING CALL SUBROUTINE
C   CHOICE WHICH DISPLAYS VALUES OF ITER,DIFF0,DIFF1 AND ASKS IF
C   ITERATION SHOULD PROCEED (INDIT=1),OR STOP (INDIT=0).
      INDIT=1
      IF(DIFF1.GT.DIFF0) CALL CHOICE(DIFF0,DIFF1,ITER,INDIT)
C   REDEFINE DIFF0,RM<R,I>(I)
      DIFF0=DIFF1
      CALL SAVE(LX,RMR,RMNR)
      CALL SAVE(LX,RMI,RMNI)
      IF(INDIT.EQ.0) RETURN
C   *****
C   CHECK IF DIFF1 IS LESS THAN DIFFMN,IN WHICH CASE CONVERGENCE
C   HAS OCCURRED AND EXIT IS MADE FROM ITERATION LOOP.
      IF(DIFF1.LT.DIFFMN) RETURN
C   *****
C
200  CONTINUE
C
C
C   ARRIVAL AT THIS POINT IMPLIES ITERATION HAS NOT
C   CONVERGED SATISFACTORILY AFTER NITER ITERATIONS.THE
C   VALUES OF NITER AND DIFF0 ARE PRINTED OUT AND THEN
C   RETURN IS MADE TO THE MAIN PROGRAM.
      WRITE(6,6100) NITER,DIFF0
6100  FORMAT(1X,'AFTER',I3,' ITERATIONS RMS DIFFERENCE IS',
     1 '1 STILL',E10.2,'/','RETURNING TO MAIN PROGRAM')
      RETURN
C   *****
C   END
      SUBROUTINE SUM(LX)
C   THE SUBROUTINE PERFORMS THE SUMMATION IN EQU. 4.ON ENTRY TO THE
C   SUBROUTINE THE FOLLOWING ARRAYS ARE DEFINED IN COMMON;
      COMMON /MAITSU/ RMR(1024),RMI(1024),SUMR(1024),SUMI(1024),S(1024)
      1,FMR(1024),FMI(1024)
      COMMON /MASU/ H(1024),RK(1024)
      DIMENSION HNMR(1024),FHNMR(1024),FHNMI(1024),RKN(1024),
     1TERMR(1024),TERMI(1024)
C   BEFORE ENTERING SUMMATION LOOP,SET UP INITIAL VALUES OF
C   THE VARIABLES AND ARRAYS IMPLICIT IN THE SECOND TERM OF EQU.4.
C   HNMR(I) IS THE ARRAY DEFINED BY M*H**N,FHNM<R,I>(I) IS ITS
C   FOURIER TRANSFORM.RKN(I) IS THE ARRAY DEFINED BY K**N/N! AND
C   TERM<R,I>(I) IS THE ARRAY DEFINED BY THE PRODUCT OF RKN(I)
C   AND FHNM<R,I>(I),I.E.IT REPRESENTS THE NTH TERM IN THE
C   SUMMATION.THE NTH PARTIAL SUM IS REPRESENTED BY THE ARRAY
C   SUM<R,I>(I).
      CALL SAVE(LX,HNMR,RMR)
      CALL CONST(LX,RKN,1.0)
      CALL SAVE(LX,TERMR,FMR)
      CALL SAVE(LX,TERMI,FMI)
C   S1 IS MAXIMUM CONTRIBUTION TO THE TRANSFORM OF THE MAGNETISATION
C   FROM THE FIRST TERM IN THE SUMMATION.
      CALL MINMAX(LX,TERMR,TMINR,TMAXR)
      CALL MINMAX(LX,TERMI,TMINI,TMAXI)
      S1=AMAX1(ABS(TMINR),ABS(TMAXR),ABS(TMINI),ABS(TMAXI))
      CALL SAVE(LX,SUMR,TERMR)
      CALL SAVE(LX,SUMI,TERMI)

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C NOW ENTER SUMMATION LOOP, WITH INDEX N.EXIT FROM THE
C LOOP OCCURS IF: (1) THE NUMBER OF ITERATIONS EXCEEDS THE SET
C VALUE NSUM WITHOUT CONVERGENCE HAVING OCCURRED, (2) IF THE RATIO
C SN/S1 BECOMES LESS THAN THE SET VALUE EPS, WHERE SN IS THE
C MAXIMUM CONTRIBUTION TO THE TRANSFORM OF THE MAGNETISATION FROM
C THE NTH TERM IN THE SUMMATION.
  NSUM=30
  EPS=5.0E-05

C
  DO 100 N=1, NSUM
C   REDEFINE ARRAYS HNMR(I), PHNM<R,I>(I), RKN(I) FOR NEXT ITERATION.
    CALL MULT(LX, HNMR, HNMR, H)
    CALL SAVE(LX, FHNMR, HNMR)
    CALL CONST(LX, FHNMI, 0.0)
    CALL MULT(LX, RKN, RKN, RK)
    CALL SCALE(LX, RKN, 1.0/FLOAT(N), RKN)
    CALL FURRY(LX, FHNMR, FHNMI, 1.0, S)
C   CALCULATE NTH TERM AND NTH PARTIAL SUM.
    CALL MULT(LX, TERMR, RKN, FHNMR)
    CALL MULT(LX, TERMI, RKN, FHNMI)
    CALL ADDSUB(LX, SUMR, SUMR, TERMR, 1)
    CALL ADDSUB(LX, SUMI, SUMI, TERMI, 1)
    CALL MINMAX(LX, TERMR, TMINR, TMAXR)
    CALL MINMAX(LX, TERMI, TMINI, TMAXI)
    SN=AMAX1(ABS(TMINR), ABS(TMAXR), ABS(TMINI), ABS(TMAXI))
C   EXIT FROM LOOP IF CONVERGENCE CRITERION SATISFIED.
    IF((SN/S1).LT.EPS) RETURN
    *****
C 100 CONTINUE
C
C ARRIVAL AT THIS POINT IMPLIES NSUM TERMS HAVE BEEN EVALUATED
C WITHOUT CONVERGENCE HAVING TAKEN PLACE. VALUES OF NSUM
C AND SN/S1 ARE DISPLAYED ON UNIT 6.
  SNS1=SN/S1
  WRITE(6, 6000) NSUM, SNS1
6000 FORMAT(1X, I4, ' TERMS OF SUM HAVE BEEN COMPUTED ', /
1, ' WITHOUT CONVERGENCE TAKING PLACE. ', /,
2 ' THE VALUE OF SN/S1 IS ', E10.2, ' , RETURNING FROM SUM ')
  RETURN
  END
  SUBROUTINE CHOICE(DIFF0, DIFF1, ITER, INDIT)
  WRITE(6, 600) ITER, DIFF0, DIFF1
600 FORMAT(1X, ' AFTER ', I3, ' ITERATIONS DIVERGENCE IS OCCURRING ',
1, ' PREVIOUS RMS DIFFERENCE WAS ', E10.2, ' PRESENT IS ', E10.2, /,
2 ' DO YOU WANT ITERATION TO CONTINUE? ', /,
3 ' TYPE 1(FORMAT1) FOR YES, RETURN FOR NO. ')
  READ(5, 500) INDIT
500 FORMAT(I1)
  RETURN
  END
  SUBROUTINE READ(L, INDIR, LX, DX, BETAL, BETA2, RIO, DO, RIM, DM, FV, H0
1, AV, RLAT, RLON, DIST, A, Z, TITLE)
C THE SUBROUTINE READS FROM UNIT 1, THE VARIABLES AND ARRAYS;
C L, NO. OF EQUALLY SPACED OBSERVATION POINTS IN PROFILE.
C INDIR, INDICATOR; =-1 IF PROFILE IS TO BE REVERSED FOR PLOT
C =1, OTHERWISE
C ILAYER=1 IF LAYER 1(SEA-BOTTOM) IS USED FOR INVERSION
C =2 IF LAYER 2 IS USED
C BETAL, AVERAGE BEARING OF PROFILE IN DEGREES
C BETA2, BEARING OF LINE PERPENDICULAR TO STRUCTURE, ONTO WHICH
C DATA IS TO BE PROJECTED. NOTE: BOTH ANGLES ARE FOR REVERSED
C LINE IF INDIR=-1
C RIO, INCLINATION OF THE EARTH'S MAGNETIC FIELD, POSITIVE DOWNWARDS
C IN DEGREES.
C DO, DECLINATION OF EARTH'S FIELD IN DEGREES, CLOCKWISE FROM NORTH
C RIM, DM, INCLINATION AND DECLINATION OF MAGNETISATION, AS ABOVE
C FV(I) I=1, 4, THESE VALUES OF SPATIAL FREQUENCY (NOT WAVENUMBER)
C SPECIFY THE BANDPASS FILTER WHICH MUST BE EMPLOYED TO PREVENT
C LONG AND SHORT WAVELENGTH INSTABILITY. SEE FURPAK, BDPASS FOR DETAILS
C UNITS ARE 1/KM.
C H0, THICKNESS OF MAGNETIC LAYER IN KM.
C AV, MAGNITUDE OF ANNIHILATOR TO BE ADDED TO COMPUTED MAGNETISATION
C RLAT(I), I=1, L, ARRAY OF LATITUDE VALUES
C RLON(I), ARRAY OF LONGITUDE VALUES.
C DIST(I), I=1, L, ARRAY OF DISTANCES, IN KM
C A(I), I=1, L, ARRAY OF MAGNETIC ANOMALIES IN NANOTESLAS.
C Z(I), I=1, L, ARRAY OF DEPTHS TO MAGNETIC BASEMENT IN KM.
C
  DIMENSION RLAT(1024), RLON(1024), DIST(1024), A(1024), Z(1024), FV(4)
1, TITLE(8)
  REAL*8 FMT(4), FMT1(4), FMT2(4)
  DATA FMT1 / (F8.4, F9.4, F7.2, F6.0, 6X, F6.3) /,
1 FMT2 / (F8.4, F9.4, F7.2, F6.0, 12X, F6.3) /
C
  READ(1, 1000) TITLE
1000 FORMAT(8A4)
  READ(1, 100) L, INDIR, ILAYER

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MAGINV 6

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100  FORMAT(3I5)
      DO 2 JF=1,4
      IF (ILAYER.EQ.1) FMT(JF)=FMT1(JF)
      IF (ILAYER.EQ.2) FMT(JF)=FMT2(JF)
2    CONTINUE
      READ(1,110) BETA1,BETA2
110  FORMAT(4F10.0)
      READ(1,110) RIO,D0,RIM,DM
      D0=BETA2-D0
      DM=BETA2-DM
      READ(1,120) FV,H0,AV
120  FORMAT(6F10.0,////)
      IF (IABS(INDIR).NE.1) STOP
      IF (INDIR.EQ.-1) II=L
      IF (INDIR.EQ.1) II=1
      DO 10 I=1,L
      READ(1,FMT) RLAT(II),RLON(II),DIST(II),A(II),Z(II)
      II=II+INDIR
10  CONTINUE
      DX=(DIST(L)-DIST(1))/FLOAT(L-1)*COS(0.017453*(BETA2-BETA1))
      LX=1
      LL=L*5/4
      DO 30 K=1,1000
      LX=2*LX
      IF (LX.GT.LL) GOTO 40
30  CONTINUE
40  RETURN
      END
      SUBROUTINE PLOTM(L,INDIR,DX,RIO,D0,RIM,DM,H,A,FV,H0,ZMED
1,ACOMR,ZMMR,RMR,DIST,AV,DIFFA,BETA1,BETA2,NITSM,NITSA,RLAT,
2RLON,TITLE)
      DIMENSION H(L),A(L),ACOMR(L),ZMMR(L),RMR(L),DIST(L),RLAT(1024)
1,RLON(1024),FV(4),TITLE(8)
      DO 5 I=1,L
      H(I)=ZMED-H(I)
5  CONTINUE
      XP1=0.3
      XP2=XP1+3.937E-03*ABS(DIST(L)-DIST(1))
      CALL PAPER(1)
      CALL CSPACE(0.0,XP2+0.1,0.0,1.0)
      CALL PSPACE(XP1,XP2,0.0787,0.2362)
      CALL MAP(DIST(1),DIST(L),8.0,0.0)
      CALL CTRMAG(12)
      CALL PLOTCS(DIST(1)-INDIR*69.67,10.651,'LATITUDE',8)
      CALL PLOTCS(DIST(1)-INDIR*69.67,11.522,'LONGITUDE',9)
      CALL PLOTCS(DIST(L/2)-INDIR*15.0,9.781,'DISTANCE(KM)',12)
      DO 7 I=1,L,50
      CALL PLOTNC(DIST(I),10.20,54)
      CALL PLOTNF(DIST(I),10.651,RLAT(I),2)
      CALL PLOTNF(DIST(I),11.522,RLON(I),2)
7  CONTINUE
      CALL POSITN(DIST(1),H(1))
      DO 10 I=2,L
      CALL JOIN(DIST(I),H(I))
      IF (ZMMR(I).LT.0.0) GOTO 10
      CALL JOIN(DIST(I),H(I)+H0)
      CALL JOIN(DIST(I),H(I))
10  CONTINUE
      CALL POSITN(DIST(1),H(1)+H0)
      DO 20 I=2,L
      CALL JOIN(DIST(I),H(I)+H0)
20  CONTINUE
      CALL SCALES
      CALL BORDER
      CALL PSPACE(XP1,XP2,0.2756,0.5118)
      CALL MAP(DIST(1),DIST(L),0.0,0.0)
      CALL PTPLOT(DIST,A,1,L,-2)
      CALL PTPLOT(DIST,ACOMR,1,L,43)
      CALL POSITN(DIST(L),0.0)
      CALL JOIN(DIST(1),0.0)
      CALL SCALES
      CALL BORDER
      CALL PSPACE(XP1,XP2,0.5512,0.6693)
      CALL MAP(DIST(1),DIST(L),0.00,0.00)
      CALL PTPLOT(DIST,RMR,1,L,-2)
      CALL POSITN(DIST(L),1.0)
      CALL JOIN(DIST(1),1.0)
      CALL SCALSI(0.01,0.2)
      CALL BORDER
      CALL PSPACE(XP1,XP2,0.7087,0.9449)
      IF (AV.EQ.0.0) GOTO 35
      DO 30 I=1,L
      ZMMR(I)=ZMMR(I)+AV*RMR(I)
30  CONTINUE
35  CALL MAP(DIST(1),DIST(L),0.0,0.0)
      CALL PTPLOT(DIST,ZMMR,1,L,-2)
      CALL POSITN(DIST(L),0.0)

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CALL JOIN(DIST(1),0.0)
CALL SCALES
CALL BORDER
CALL CSPACE(0.0206,0.3,0.0,1.0)
CALL PLACE(1,5)
CALL TYPECS('COMPUTED',8)
CALL CRLNFD
CALL TYPECS('MAGNETISATION(A/M)',18)
CALL CRLNFD
CALL TYPECS('MEAN VALUE=',11)
CALL TYPENF(AV,2)
CALL CRLNFD
CALL TYPECS('NO. OF ITERATIONS=',18)
CALL TYPENI(NITSM)
CALL PLACE(1,21)
CALL TYPECS('MAGNETIC',8)
CALL CRLNFD
CALL TYPECS('ANNIHILATOR(A/M)',16)
CALL CRLNFD
CALL TYPECS('NO. OF ITERATIONS=',18)
CALL TYPENI(NITSA)
CALL PLACE(1,30)
CALL TYPECS('TOTAL FIELD MAGNETIC',20)
CALL CRLNFD
CALL TYPECS('ANOMALY(NANOTESLA)',18)
CALL CRLNFD
CALL TYPECS('SOLID LINE,OBSERVED',19)
CALL CRLNFD
CALL TYPECS('CROSSES,CALCULATED',18)
CALL CRLNFD
CALL TYPECS('RMS RESIDUAL=',13)
CALL TYPENF(DIFFA,2)
CALL PLACE(1,46)
CALL TYPECS('DEPTH TO MAGNETIC',17)
CALL CRLNFD
CALL TYPECS('BASEMENT(KM)',12)
CALL CRLNFD
CALL TYPECS('LAYER THICKNESS=',16)
CALL TYPENF(H0,1)
CALL PSPACE(XP2+0.0394,XP2+0.3543,0.0787,0.9449)
CALL BORDER
CALL CSPACE(XP2+0.0600,XP2+0.3543,0.0,1.0)
CALL PLACE(1,5)
CALL TYPECS(TITLE,32)
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('AVERAGE BEARING',15)
CALL CRLNFD
CALL TYPECS('OF PROFILE=',12)
CALL TYPENF(BETA1,2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('DATA PROJECTED ONTO',19)
CALL CRLNFD
CALL TYPECS('LINE BEARING',12)
CALL TYPENF(BETA2,2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('DECLINATION OF',14)
CALL CRLNFD
CALL TYPECS('EARTHS FIELD=',13)
CALL TYPENF(BETA2-57.2958*D0,2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('INCLINATION OF',14)
CALL CRLNFD
CALL TYPECS('EARTHS FIELD=',13)
CALL TYPENF(57.2958*RI0,2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('DECLINATION OF',14)
CALL CRLNFD
CALL TYPECS('MAGNETISATION=',14)
CALL TYPENF(BETA2-57.2958*DM,2)
CALL SUPFIX

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CALL TYPENC(25)
CALL NORMAL
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('INCLINATION OF',14)
CALL CRLNFD
CALL TYPECS('MAGNETISATION=',14)
CALL TYPENF(57.2958*RIM,2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('INCLINATION OF',14)
CALL CRLNFD
CALL TYPECS('MAGNETISATION IN',16)
CALL CRLNFD
CALL TYPECS('PLANE OF PROFILE=',18)
CALL TYPENF(57.2958*ATAN2(SIN(RIM),COS(RIM)*COS(DM)),2)
CALL SUPFIX
CALL TYPENC(25)
CALL NORMAL
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('PROJECTED AVERAGE',17)
CALL CRLNFD
CALL TYPECS('SAMPLE SPACING=',15)
CALL TYPENF(DX,3)
CALL TYPECS(' KM',3)
CALL CRLNFD
CALL CRLNFD
CALL TYPECS('FILTER VALUES',13)
CALL CRLNFD
CALL TYPECS('F0=',3)
CALL TYPENF(1.0/DX,4)
CALL TYPECS(' KM',3)
CALL SUPFIX
CALL TYPECS('-1',2)
CALL NORMAL
CALL CRLNFD
CALL TYPECS('F1=',3)
CALL TYPENF(FV(1),4)
CALL TYPECS(' KM',3)
CALL SUPFIX
CALL TYPECS('-1',2)
CALL NORMAL
CALL CRLNFD
CALL TYPECS('F2=',3)
CALL TYPENF(FV(2),4)
CALL TYPECS(' KM',3)
CALL SUPFIX
CALL TYPECS('-1',2)
CALL NORMAL
CALL CRLNFD
CALL TYPECS('F3=',3)
CALL TYPENF(FV(3),4)
CALL TYPECS(' KM',3)
CALL SUPFIX
CALL TYPECS('-1',2)
CALL NORMAL
CALL CRLNFD
CALL TYPECS('F4=',3)
CALL TYPENF(FV(4),4)
CALL TYPECS(' KM',3)
CALL SUPFIX
CALL TYPECS('-1',2)
CALL NORMAL
CALL CRLNFD
CALL GREND
RETURN
END

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      SUBROUTINE ADDSUB(LX,Z,X,Y,IND)
C THE SUBROUTINE ADDS THE ARRAY X TO THE ARRAY Y ELEMENT BY ELEMENT
C FOR I=1,2,...,LX,IF IND IS POSITIVE OR SUBTRACTS Y FROM X IF IND
C IS NEGATIVE.IN BOTH CASES THE ANSWER IS RETURNED IN Z.
      DIMENSION X(LX),Y(LX),Z(LX)
      IF(IND.LT.0) GOTO 15
      DO 10 I=1,LX
      Z(I)=X(I)+Y(I)
10 CONTINUE
      RETURN
15 DO 20 I=1,LX
      Z(I)=X(I)-Y(I)
20 CONTINUE
      RETURN
      END
      SUBROUTINE BDPASS(LX,F0,F1,F2,F3,F4,F,IFAIL)
C THE SUBROUTINE DEFINES A BANDPASS FILTER F(I),I=1,2,...,LX IN
C THE FREQUENCY DOMAIN.F0 IS THE ORIGINAL SAMPLING FREQUENCY AND
C THE ELEMENT F(LX/2+1) IS THE FILTER VALUE AT THE NYQUIST FREQUENCY.
C THE FILTER IS DEFINED AS FOLLOWS;FOR FREQUENCIES FROM 0 TO F1 THE
C FILTER HAS VALUE 0,FROM F1 TO F2 IT IS COSINE TAPERED FROM 0 TO 1,FROM
C F2 TO F3 IT HAS VALUE 1,FROM F3 TO F4 IT IS COSINE TAPERED FROM 1 TO 0
C AND FROM F4 TO THE NYQUIST FREQUENCY IT HAS VALUE 0 AGAIN.FILTER VALUES
C FOR FREQUENCIES BETWEEN THE NYQUIST FREQUENCY AND F0 (I.E,EFFECTIVELY
C NEGATIVE FREQUENCIES) ARE GENERATED BY CALLING SUBROUTINE EVEN.
      DIMENSION F(LX)
      DEL=FLOAT(LX)/F0
      N1=IFIX(F1*DEL)+1
      N2=IFIX(F2*DEL)+1
      N3=IFIX(F3*DEL)+1
      N4=IFIX(F4*DEL)+1
C ERROR TRAPS,RETURN WITH IFAIL=1
      IFAIL=1
      IF(N1.LT.1) RETURN
      IF(N2.LT.N1) RETURN
      IF(N3.LT.N2) RETURN
      IF(N4.LT.N3) RETURN
      NN1=LX/2+1
      IF(N4.GT.NN1) RETURN
      IFAIL=0
      N1M=N1-1
      N2M=N2-1
      N3M=N3-1
      N4M=N4-1
      IF(N1.EQ.1) GOTO 12
      DO 10 I1=1,N1M
      F(I1)=0.0
10 CONTINUE
12 IF(N1.EQ.N2) GOTO 14
      CALL COSTAP(LX,F,N1,N2M,+1)
14 IF(N2.EQ.N3) GOTO 22
      DO 20 I2=N2,N3M
      F(I2)=1.0
20 CONTINUE
22 IF(N3.EQ.N4) GOTO 24
      CALL COSTAP(LX,F,N3,N4M,-1)
24 IF(N3.EQ.NN1) GOTO 32
      DO 30 I3=N4,NN1
      F(I3)=0.0
30 CONTINUE
      GOTO 34
32 F(NN1)=1.0
34 CALL EVEN(LX,F)
      RETURN
      END
      SUBROUTINE BDSTOP(LX,F0,F1,F2,F3,F4,F,IFAIL)
C THE SUBROUTINE DEFINES A BANDSTOP FILTER WHOSE CHARACTERISTICS ARE
C COMPLEMENTARY TO THAT GENERATED BY BDPASS.
      DIMENSION F(LX)
      CALL BDPASS(LX,F0,F1,F2,F3,F4,F,IFAIL)
      DO 10 I=1,LX
      F(I)=1.0-F(I)
10 CONTINUE
      RETURN
      END
      SUBROUTINE CONST(LX,X,VAL)
C THE SUBROUTINE ASSIGNS THE VALUE,VAL TO EACH ELEMENT X(I),I=1,2,...,LX
      DIMENSION X(LX)
      DO 10 I=1,LX
      X(I)=VAL
10 CONTINUE
      RETURN
      END
      SUBROUTINE COSTAP(LX,F,L1,L2,ISIGN)
C THE SUBROUTINE DEFINES THE ELEMENTS F(I),I=L1,L1+1,...,L2 BY A RISING
C COSINE TAPER BETWEEN 0 AND 1 IF ISIGN=+1,OR AS A FALLING COSINE TAPER

```

```
C BETWEEN 1 AND 0 IF ISIGN=-1
    DIMENSION F(LX)
    DELARG=3.14159265/FLOAT(L2-L1+1)
    ARG=0.0
    DO 10 I=L1,L2
        F(I)=0.5*(1-ISIGN*COS(ARG))
        ARG=ARG+DELARG
10 CONTINUE
    RETURN
END
SUBROUTINE EVEN(LX,X)
C THE SUBROUTINE TAKES AS INPUT AN ARRAY X(I) WHICH IS DEFINED FOR
C I=1,2,...,LX/2+1.THE ARRAY IS EXTENDED TO CREATE AN EVEN FUNCTION IN
C THE TRANSFORM DOMAIN,I.E.X(LX-I+2)=X(I)
    DIMENSION X(LX)
    NN=LX/2
    LXI2=LX
    DO 10 I=2,NN
        X(LXI2)=X(I)
        LXI2=LXI2-1
10 CONTINUE
    RETURN
END
SUBROUTINE FURRY(LX,X,Y,SIGNI,S)
C
C FAST FOURIER TRANSFORM SUBROUTINE
C
C A.G.NUNNS 1977,MODIFIED FROM CLAERBOUT (1976),FUNDAMENTALS OF
C GEOPHYSICAL DATA PROCESSING
C
C THE COOLEY-TUKEY ALGORITHM IS USED TO CALCULATE THE FOURIER TRANSFORM
C OF A COMPLEX TIME SERIES (X(J),Y(J)) OF LX EQUALLY SPACED POINTS WHERE
C LX=2**INTEGER.THE COMPLEX TRANSFORM,(X(K),Y(K)) IS OBTAINED AT LX
C EQUALLY SPACED FREQUENCY POINTS FROM ZERO TO 2*PI.
C
C      LX
C X(K)=SC*(SUM(X(J)*COS(ARG)-Y(J)*SIN(ARG)))
C      J=1
C                                     FOR K=1,2,...,LX
C      LX
C Y(K)=SC*(SUM(X(J)*SIN(ARG)+Y(J)*COS(ARG)))
C      J=1
C
C WHERE   ARG=2*PI*SIGNI*(J-1)*(K-1)/LX
C
C IF SIGNI=+1.0,THEN SC=1.0 AND FORWARD TRANSFORM IS OBTAINED
C IF SIGNI=-1.0,THEN SC=1.0/LX AND REVERSE TRANSFORM IS OBTAINED
C
C LIST OF SUBROUTINE ARGUMENTS
C LX     THE NUMBER OF DATA POINTS
C X      THE ARRAY X(1),X(2),...,X(LX) WHICH CONTAINS THE REAL PART
C       OF THE TIME SERIES WHEN THE SUBROUTINE IS CALLED AND ON
C       RETURN THE REAL PART OF THE FOURIER TRANSFORM
C Y      THE ARRAY Y(1),Y(2),...,Y(LX) CONTAINING THE IMAGINARY PARTS
C       OF THE TIME SERIES AND TRANSFORM ON CALL AND RETURN RESPECTIVELY
C SIGNI  INDICATOR FOR FORWARD OR INVERSE TRANSFORM
C S      THE ARRAY S(1),S(2),...,S(LX/4+1) CONTAINING TABLE OF
C       SINE VALUES EVALUATED BY SUBROUTINE SINTAB,CALLED FROM MAIN
C
    DIMENSION X(LX),Y(LX),S(LX)
C
    SC=1.0
    IF(SIGNI.LT.0.0) SC=1.0/LX
    NN=LX/2
    N=LX/4
C
C REORDER DATA POINTS BY REVERSING ORDER OF THE BINARY DIGITS OF THEIR
C INDICES(INDEX OF X(J) IS (J-1)) AND MULTIPLY BY SCALE CONSTANT SC
C
    J=1
    DO 30 I=1,LX
        IF(I.GT.J) GO TO 10
        TEMPX=X(J)*SC
        TEMPY=Y(J)*SC
        X(J)=X(I)*SC
        Y(J)=Y(I)*SC
        X(I)=TEMPX
        Y(I)=TEMPY
10 M=NN
20 IF(J.LE.M) GO TO 30
    J=J-M
    M=M/2
    IF(M.GE.1) GO TO 20
30 J=J+M
C
C COMPUTE TRANSFORM USING DOUBLING PROCEDURE.THE REQUIRED SINE AND
C COSINE VALUES FROM ZERO TO PI ARE ALL OBTAINED BY APPROPRIATE INDEXING
```


C FROM THE ARRAY S WHICH HAS SINES EVALUATED FROM ZERO TO PI/2

C

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```
      L=1
      NL=NN
40  ISTEP=2*L
      IND=1
      DO 55 M=1,L
        INDN=IND-N-1
        IF(INDN) 41,42,43
41  WX=S(1-INDN)
      WY=S(IND)*SIGNI
      GO TO 45
42  WX=0.
      WY=SIGNI
      GO TO 45
43  WX=-S(INDN+1)
      WY=S(N+1-INDN)*SIGNI
45  DO 50 I=M,LX,ISTEP
      TEMPX=WX*X(I+L)-WY*Y(I+L)
      TEMPY=WX*Y(I+L)+WY*X(I+L)
      X(I+L)=X(I)-TEMPX
      Y(I+L)=Y(I)-TEMPY
      X(I)=X(I)+TEMPX
50  Y(I)=Y(I)+TEMPY
55  IND=IND+NL
      L=ISTEP
      NL=NL/2
      IF(L.LT.LX) GO TO 40
      RETURN
      END
      SUBROUTINE MEAN(L,X,XAV)
```

C

C A.G.NUNNS 1977

C

C THE SUBROUTINE CALCULATES THE MEAN VALUE,XAV, OF AN ARRAY,

C X(I) I=1,....,L

C AND SUBTRACTS IT FROM EACH ELEMENT,X(I).THE NEW ARRAY X IS

C RETURNED TOGETHER WITH XAV.

C

```
      DIMENSION X(L)
      XAV=0.0
      DO 10 I=1,L
        XAV=XAV+X(I)
10  CONTINUE
      XAV=XAV/L
      DO 20 I=1,L
        X(I)=X(I)-XAV
20  CONTINUE
      RETURN
      END
      SUBROUTINE MINMAX(LX,X,XMIN,XMAX)
```

C

C A.G.NUNNS

C

C THE SUBROUTINE RETURNS THE MINIMUM VALUE,XMIN AND THE MAXIMUM VALUE

C XMAX,OF AN ARRAY,X OF LX ELEMENTS

C

```
      DIMENSION X(LX)
      XMIN=X(1)
      XMAX=X(1)
      DO 10 I=1,LX
        IF(X(I).LT.XMIN) XMIN=X(I)
        IF(X(I).GT.XMAX) XMAX=X(I)
10  CONTINUE
      RETURN
      END
      SUBROUTINE MSD(L,X,AV,SD)
```

C THE SUBROUTINE CALCULATES THE MEAN,AV AND THE STANDARD

C DEVIATION,SD(BIASED) OF THE ARRAY X(I),I=1,2,....,L

C USING FIRST VALUE AS WORKING MEAN

```
      DIMENSION X(L)
      SSQ=0.0
      AV=0.0
      DO 10 I=1,L
        DUM=X(I)-X(1)
        AV=AV+DUM
        SSQ=SSQ+DUM**2
10  CONTINUE
      AV=AV/L
      SD=SQRT(SSQ/L-AV**2)
      AV=AV+X(1)
      RETURN
      END
      SUBROUTINE MULT(LX,Z,X,Y)
```

C THE SUBROUTINE MULTIPLIES THE ARRAY X(I) BY THE ARRAY Y(I),ELEMENT

C BY ELEMENT FOR I=1,2,....,LX,RETURNING THE ANSWER IN THE ARRAY Z(I).

DIMENSION X(LX),Y(LX),Z(LX)

```

DO 10 I=1,LX
Z(I)=X(I)*Y(I)
10 CONTINUE
RETURN
END
SUBROUTINE ODD(LX,X)
C THE SUBROUTINE TAKES AS INPUT AN ARRAY X(I) WHICH IS DEFINED FOR
C I=1,2,...,LX/2+1. THE ARRAY IS EXTENDED TO CREATE AN ODD FUNCTION IN
C THE TRANSFORM DOMAIN, I.E. X(LX-I+2)=-X(I)
DIMENSION X(LX)
NN=LX/2
LXI2=LX
DO 10 I=2,NN
X(LXI2)=-X(I)
LXI2=LXI2-1
10 CONTINUE
X(NN+1)=0.0
RETURN
END
SUBROUTINE RMS(LX,DIFF,X,Y)
C THE SUBROUTINE CALCULATES THE RMS DIFFERENCE BETWEEN THE ARRAYS
C X(I) AND Y(I) AND RETURNS IT AS DIFF.
DIMENSION X(LX),Y(LX)
DIFF=0.0
DO 10 I=1,LX
DIFF=DIFF+(X(I)-Y(I))**2
10 CONTINUE
DIFF=SQRT(DIFF/LX)
RETURN
END
SUBROUTINE SAVE(LX,X,Y)
C THE SUBROUTINE SAVES THE ARRAY Y(I) AS THE ARRAY X(I), I=1,2,...,LX
DIMENSION X(LX),Y(LX)
DO 10 I=1,LX
X(I)=Y(I)
10 CONTINUE
RETURN
END
SUBROUTINE SCALE(LX,Z,CONST,X)
C THE SUBROUTINE MULTIPLIES EVERY ELEMENT OF THE ARRAY X(I), I=1,2,...,LX
C BY THE CONSTANT CONST AND RETURNS THE ANSWER IN ARRAY Z(I).
DIMENSION X(LX),Z(LX)
DO 10 I=1,LX
Z(I)=CONST*X(I)
10 CONTINUE
RETURN
END
SUBROUTINE SINTAB(LX,S)
C
C SINE TABLE SUBROUTINE
C
C THE SUBROUTINE RETURNS VALUES
C
C S(I)=SIN(2*PI*(I-1)/LX) I=1,2,...,LX/4+1
C
C WHERE LX=2**INTEGER IS THE NUMBER OF POINTS IN THE TIME SERIES IN MAIN
C
DIMENSION S(LX)
DOUBLE PRECISION ARG,DELARG
C
N1=LX/4+1
ARG=0.0
DELARG=0.6283185307179586D01/LX
DO 10 I=1,N1
S(I)=DSIN(ARG)
ARG=ARG+DELARG
10 CONTINUE
RETURN
END
SUBROUTINE TAPJN(LX,L,X)
C THE SUBROUTINE PADS OUT THE ARRAY X(I), I=1,2,...,L TO LX
C ELEMENTS BY JOINING THE END POINTS IN CYCLIC FASHION BY
C HALF A COSINE WAVE.
DIMENSION X(LX)
DELX=(X(1)-X(L))/2.0
XL1=0.5*(X(L)+X(1))
L1=L+1
DELARG=3.14159/FLOAT(LX-L+1)
ARG=0.0
DO 10 I=L1,LX
ARG=ARG+DELARG
X(I)=XL1-DELX*COS(ARG)
10 CONTINUE
RETURN
END
SUBROUTINE UPCON(LX,F,FREQ0,Z)
C

```

C A.G.NUNNS 1977

C

C THE SUBROUTINE GENERATES A FREQUENCY DOMAIN FILTER,
C F(I), I=1,2,...,LX FOR UPWARD CONTINUATION. INPUT ARGUMENTS ARE
C LX, THE NUMBER OF POINTS IN THE FOURIER TRANSFORM,
C FREQ0, THE ORIGINAL SPATIAL SAMPLING FREQUENCY,
C Z, THE HEIGHT TO WHICH WAVEFORM IS TO BE UPWARD CONTINUED.
C FREQ0 AND Z MUST BE IN CONSISTENT UNITS, E.G. WAVELENGTHS/KM AND KM.

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 DIMENSION F(LX)

 NN1=LX/2+1

 FACT=-6.28319*FREQ0*Z/LX

 DO 10 I=2,NN1

 F(I)=EXP(FACT*(I-1))

 C F(I)=EXP(-K*Z), WHERE K IS WAVENUMBER(=2*PI/WAVELENGTH)

 LI2=LX-I+2

 F(LI2)=F(I)

 C SO THAT FILTER IS SYMMETRIC FOR NEGATIVE FREQUENCIES

 10 CONTINUE

 F(1)=1.0

 RETURN

 END

B.2 Description of programs for seismic processing

SYNSEI - synthetic seismogram generator

INT - interpolation

VEL - velocity analysis

CDP - common depth point stacking

See Chapter 3 for a description of how these programs operate. The details of input/output for the seismic trace and semblance data will not be given here, as in order to be comprehended, they require a knowledge of the RT-11 operating system. A brief description of the basic input parameters follows. Many of these are common to a number of the programs. They are described in the order that they appear and are underlined in the program listings. The variable names do not correspond to the symbols used in Section 3.4. Any consistent system of units may be used.

L - number of points in each seismic trace

NCHAN - number of seismic channels

NSTART - sample number of first point in seismic trace, relative to 0 for time zero.

NLYR - number of reflectors in synthetic seismogram

XSTART - offset for receiver 1

XSTEP - receiver spacing

FSAMP - sampling frequency

FRICK - frequency of synthetic pulse (see page 76)

FNOISE - frequency limit of noise for synthetic seismogram

SN - peak signal/rms noise for synthetic seismogram

TOLYR - array of vertical incidence times

VLyr - array of move out velocities

M - number of times sampling rate is to be increased by interpolation

NTAP - length of cosine taper applied to ends of trace

INDEN - if negative, no scaling of trace; if zero, scale to unit rms; if positive, use inverse energy scaling

T01, T02, T0STEP, TGATE - initial and final vertical incidence times, increment and gate width for velocity analysis

VSTEP, V1, V2MIN, V2MAX, VICPT, VGRAD - velocity increment and parameters describing velocity limits for velocity analysis (see figure)

INDMUT - if negative, no muting

MUTE - array giving number of samples to be muted for each channel

For a description of the array processor subroutines, reference should be made to the technical publications of Floating Point Systems Inc., Beaverstone, Oregon 97005 and in particular, the "AP Math Library Manual, volume 2, 860-7288-004, 1979", a copy of which is held by the department.

Details of the final implementation of these programs will be given by M.J.Poulter (thesis in preparation). Listings of the programs follow.

[illegible]

```

INT 1
VIRTUAL S0(23552),S1(23552)
DIMENSION DUM(2048)
DIMENSION FBUFR(3),FBUFW(3),FSTAP(23)
REAL*8 FSPLCR,FSPLCW
DATA K1M,K1,K2,KTOP,IA,IB,IB1,IB2,IC,IC1,IC1
17-1,1,2,0191,3,2010,2049,2050,4096,4097,6148/
DATA DEV/JOBLK /
C
CALL ASSIGN(1,'BK2:ANINV.PAT',13)
IF(1.FE.CHDEV).NE.D)STOP'FETCH ERROR'
IDCHR=IGETC()
IDCHW=IGETC()
C
C READ IN REQUIRED INPUT PARAMETERS
READ(1,1000) L,NCHAN,M,NHAP,INDEX
1000 FORMAT(215,5X,315)
READ(1,1100) FBUFR,FBUFW
1100 FORMAT(23A4)
C VALIDATE DATA
C NREAD IS THE NUMBER OF BLOCKS READ IN/TRACE
NREAD=L/128
C STOP IF L IS NOT AN INTEGER NUMBER OF BLOCKS
IF(L.NE.128*NREAD)STOP'L NOT INTEGER NO. OF BLOCKS'
C L2 IS NUMBER OF WORDS READ/TRACE
L2=2*L
C IAL IS POINTER NEEDED FOR COSINE TAPERING
IAL=IA+L-1
C FIND L21, INTEGER POWER OF TWO G.E. THAN L.
L21=2
DO 5 I=1,1000
IF(L21.GE.L) GOTO 10
L21=2*L21
5 CONTINUE
C STOP IF L21 IS GREATER THAN 2048
10 IF(L21.GT.2048)STOP'L21.GT.2048'
L21=L21/2-1
C STOP IF M.GT.23
IF(M.GT.23)STOP'M.GT.23'
C LH IS NUMBER OF INTERPOLATED SAMPLES/TRACE
LM=L*M
C STOP IF LM.GT.23552
IF(LM.GT.23552)STOP'LM.GT.23552'
C LM2 IS THE TOTAL NUMBER OF WORDS WRITTEN OUT/INTERPOLATED TRACE
LM2=2*LM
C NWRITE IS THE NUMBER OF BLOCKS WRITTEN/TRACE
NWRITE=LM/128
C FST0 IS ADDRESS NEEDED FOR DISK READ INTO S0(1)
FST0=ADGET(S0(1))
C FST1 IS ADDRESS NEEDED FOR DISK WRITE FROM S1
FST1=ADGET(S1(1))
C FSTAP IS ARRAY OF ADDRESSES NEEDED FOR AP TRANSFERS
KL=1
DO 20 K=1,M
FSTAP(K)=APGAD(S0(KL))
KL=KL+1
20 CONTINUE
C OPEN READ FILE
CALL IRAD50(12,FBUFR,FSPECR)
IF(LOOKUP(IDCHR,FSPECR).LT.0)STOP'LOOKUP ERROR'
C OPEN WRITE FILE
CALL IRAD50(12,FBUFW,FSPECW)
IF(ENTER(IDCHW,FSPECW,NWRITE*NCHAN+1).LT.0)STOP'ENTER ERROR'
C

```

```

C-----
C THIS PART OF THE PROGRAM SETS UP THE AP FOR INTERPOLATING THE TRACES.
C A COMPLEX EXPONENTIAL VECTOR IS FORMED IN REGION C, TO BE USED IN
C THE NEXT PART OF THE PROGRAM TO PRODUCE TIME SHIFTS OF TSAMP/M BY
C MULTIPLICATION IN THE FREQUENCY DOMAIN.
C-----

```

```

C INITIALISE AP
CALL APINIT
C PUT 1./FLOAT(M*L21) IN IC AND 1./FLOAT(NTAP) IN KTOP
CALL APVR
CALL APPUT(1./FLOAT(M*L21),IC,K1,K2)
CALL APPUT(1./FLOAT(NTAP),KTOP,K1,K2)
CALL APWD
C MULTIPLY BY 2*PI FROM TM
CALL VTSMUL(IC,K1,2317,IC,K1,K1)
C FORM VECTOR RAMP IN IA USING STARTING VALUE AND RAMP INCREMENT
C BOTH EQUAL TO VALUE IN IC. TAKE SIN AND COS OF RAMP AND
C PLACE IN C.
CALL VRAMP(IC,IC,IA,K1,L2121)
CALL CVEXP(IA,K1,IC1,K2,L2121)
C FORM COSINE TAPER NTAP LONG STARTING AT ID1
CALL VCLR(ID1,K1,K1)
CALL VTSADD(ID1,K1,2306,ID1,K1,K1)
CALL VTSMUL(KTOP,K1,2306,KTOP,K1,K1)
CALL VRAMP(ID1,KTOP,ID1,K1,NTAP)
CALL VCOS(ID1,K1,ID1,K1,NTAP)
CALL VTSADD(ID1,K1,2049,ID1,K1,NTAP)
CALL VTSMUL(ID1,K1,2327,ID1,K1,NTAP)

```

```

C-----
C IN THIS PART THE SEISMIC TRACES ARE READ IN ONE AT A TIME OFF DISK.
C EACH TRACE HAS ITS MEAN REMOVED, IS SCALED (IF REQUIRED), IS PADDED OUT
C WITH ZEROES FROM L TO L21 SAMPLES AND THEN IS INTERPOLATED SO THAT THE
C NO. OF SAMPLES BECOMES LM. THE INTERPOLATED TRACES ARE STORED ON DISK.
C-----

```

```

C JOBLK IS THE DISK BLOCK ABOUT TO BE READ
C JOBLK1 IS THE DISK BLOCK ABOUT TO BE WRITTEN
JOBLK1=1
JOBLK=1
C
C NOW ITERATE THROUGH CHANNELS INTERPOLATING THE TRACES
DO 000 JCHAN=1,NCHAN
C READ IN TRACE FROM UNIT 2
IF(IREAD(IDCHR,L2,JOBLK,FST0).LT.0)STOP'READA ERROR'
JOBLK=JOBLK+NREAD
C CLEAR A AND TRANSFER TRACE INTO IT
CALL VCLR(IA,K1,L21)
CALL APVR
CALL IVAT(IDCHR)
CALL APPUT(IA,L,K2,FSTAP(1))
CALL APWD
C FIND MEAN VALUE OF TRACE AND PLACE IN AP(KTOP)
CALL MEANV(IA,K1,KTOP,L)
C SUBTRACT MEAN FROM TRACE
CALL VHLG(KTOP,K1,KTOP,K1,K1)
CALL VSADD(IA,K1,KTOP,IA,K1,L)
C MULTIPLY ENDS OF DATA BY COSINE TAPER
CALL VHLG(IA,K1,ID1,K1,IA,K1,NTAP)
CALL VHLG(IA,K1,IB1,K1,IAL,K1,NHAP)
O IF INDEX IS NEGATIVE, NO SCALING OF TRACE

```

```

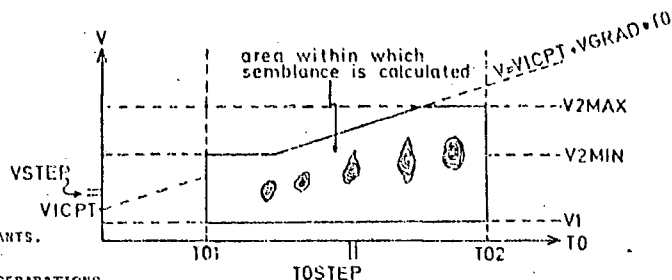
INT 2
C IF INDEEN IS ZERO SCALE TO UNIT R.M.S. VALUE
C IF INDEEN IS POSITIVE, USE INVERSE ENERGY SCALING (DIVERSITY STACK)
  IF (INDEEN.LT.0) GOTO 350
  CALL RMSEV(IA,K1,KTOP,L)
  IF (INDEEN.GT.0) CALL VSQ(KTOP,K1,KTOP,K1,K1)
C PLACE L.D FROM 1H IN IC
  CALL VCR(IC,K1,K1)
  CALL VISADD(IC,K1,VSQ,IC,K1,K1)
C USE THIS VALUE TO CREATE RECIPROCAL IN KTOP
  CALL VOIV(KTOP,K1,IC,K1,KTOP,K1,K1)
C MULTIPLY TRACE BY RECIPROCAL
  CALL VSMUL(IA,K1,KTOP,IA,K1,L)
C TRANSFER MODIFIED TRACE BACK TO S0
350  CALL APWR
  CALL APGETA(IA,L,K2,FSTAP(1))
  CALL APWD
  IF (H.EQ.1) GOTO 600
C TAKE TRANSFORM OF TRACE AND PUT IN D
  CALL RFFTR(IA,10,L21,K1)
  CALL RFFSC(10,L21,K2,K1)
C ITERATE FROM 2 TO H
  DO 400 L=2,H
C MULTIPLY TRANSFORM BY COMPLEX EXPONENTIAL VECTOR
  CALL CMUL(102,K2,101,K2,102,K2,L2121)
C MOVE D TO A AND TAKE 1H PLACE IFFT
  CALL VMOV(10,K1,1A,K1,L21)
  CALL RFFT(1A,L21,K1H)
C TRANSFER SHIFTED TRACE BACK TO S0
  CALL APWR
  CALL APGETA(1A,L,K2,FSTAP(K))
  CALL APVD
400  CONTINUE
C INTERPOLATED TRACE IS NOW IN S0 IN SCRAMBLED ORDER.
C UNSCRAMBLE INTO S1 SO THAT RECORD IS IN CORRECT TIME SEQUENCE.
500  J1=1
  DO 700 J1=1,L
  J0=J1
  DO 600 JM=1,M
  S1(J1)=S0(J0)
  J0=J0+L
  J1=J1+1
600  CONTINUE
700  CONTINUE
C NOW WRITE OUT INTERPOLATED TRACE ONTO DISK
  IF (1WRIT(1DCHW,LM2,JBLKW,FST1),LT.0) STOP 'WRIT ERROR'
  JBLKW=JBLKW+MWRITE
  CALL INAIT(1DCHW)
800  CONTINUE
  CALL CLOSEC(1DCHR)
  CALL CLOSEC(1DCHW)
  STOP
  END

```

```

VICI 1
NUMBER NMOCS(4096),TOSQ(8192),NV(612),VINSQ(178),SMBC(178),
ISQ(178),S(178),XSO(24),
JSTCH(24),JSTC(24),NCH(24),JULK(24),CONST(7),FBUF(3)
LOGICAL*FV
REAL*8 FSPEC
DATA K1,K2,KNCHIN,KGLK,KOLKIN,KADD2,KADD1,ENF,NFS,NBUF
1/1,2,0.005,0.005,0.007,0.008,0.009,0.010,0.011,0.012,0.013,0.014
DATA DEV/3RDK /
C
C ASSIGN INPUT AND OUTPUT CHANNELS
CALL ASSIGN(1,'BK2:AHINV.DAT',13)
CALL ASSIGN(2,'BK2:ARQUIV.DAT',14)
IF(1) GOTO(DEV).NE.01STOP 'FETCH ERROR'
IDCD=IGETC(1)
C READ IN REQUIRED INPUT PARAMETERS
READ(1,1000) L,NCHAN,H2TART,H1
1000 FORMAT(4I5)
READ(1,1000) FBUF
READ(1,1100) XSTART,XSTEP,FSAMP
1100 FORMAT(10F10.3)
READ(1,1100) T01,T02,TGATE
READ(1,1100) VSTEP,V1,V2MIN,V2MAX,VICPT,VGRAD
1200 FORMAT(12X,3A4)
C VALIDATE DATA AND CALCULATE REQUIRED ARRAYS AND CONSTANTS.
C FORM ARRAY JSTCH CORRESPONDING TO THE BEGINNING BLOCK
C NUMBERS FOR EACH TRACE ON DISK.
C AT SAME TIME FORM ARRAY XSO OF SQUARED SHOT RECEIVER SEPARATIONS
JSTCH(1)=1
NREAD=L*M/120
X=XSTART
XSO(1)=X**2
DO 10 JCHAN=2,NCHAN
JSTCH(JCHAN)=JSTCH(JCHAN-1)+NREAD
X=X+XSTEP
XSO(JCHAN)=X**2
10 CONTINUE
C CALCULATE FSAMP,THE SAMPLING FREQUENCY AFTER INTERPOLATION
FSAMP=FLOAT(M)*FSAMP
C CALCULATE TSAMP THE SAMPLE INTERVAL
TSAMP=1.0/FSAMP
C CALCULATE TSTART,THE BEGINNING TIME
TSTART=H2TART*TSAMP
C CALCULATE NSTH THE INITIAL SAMPLE NUMBER AFTER INTERPOLATION.
NSTH=H2TART
C TRUNCATE T0STEP TO BE INTEGRAL MULTIPLE OF TSAMP
T0STEP=TSAMP*IFIX(T0STEP/TSAMP)
C ROUND TGATE TO BE EVEN INTEGRAL MULTIPLE OF TSAMP
NGATE=IFIX(TGATE/TSAMP)*2
TGATE=FLOAT(NGATE)*TSAMP
NGATE=NGATE+1
NGTM2=M*(NGATE/2)
C ENSURE THAT T01 IS A LEAST TSTART+TGATE/2
T01=AMAX1(T01,TSTART+TGATE/2)
C ROUND T01 UP TO BE INTEGRAL MULTIPLE OF TSAMP
T01=TSAMP*IFIX(T01/TSAMP)+1
C TRUNCATE T02 IF NECESSARY SO THAT THERE IS SUFFICIENT
C DATA TO DO VELOCITY ANALYSIS AT T02
T02MAX=TSAMP*((T02-T01)/TSAMP-0.5*TSAMP)**2-
1XSO(NCHAN)/V1**2
T02=AMIN1(T02,T02MAX)
C CALCULATE NT0,THE NUMBER OF TWO WAY TIMES CONSIDERED
NT0=IFIX((T02-T01)/T0STEP)+1
T02=T01+(NT0-1)*T0STEP
IF(NT0.LE.0)STOP 'NT0 NOT POSITIVE'
C FORM ARRAYS TOSQ (SQUARED TWO WAY TIME) AND NV (NUMBER
C OF VELOCITY POINTS AT EACH TWO WAY TIME)
T0J=T01
DO 20 JTB=1,NT0
TOSQ(JTB)=T0J**2
V2=VICPT+VGRAD*T0J
V2=AMAX1(V2MIN,V2)
V2=AMIN1(V2,V2MAX)
NV(JTB)=IFIX((V2-V1)/VSTEP)+1
T0J=T0J+T0STEP
20 CONTINUE
C VINSQ IS THE ARRAY OF INVERSE SQUARED VELOCITIES
NVHX=NV(NT0)
IF(NVHX.GT.178)STOP 'NVHX GT 178'
V=V1
DO 30 JV=1,NVHX
VINSQ(JV)=1.0/V**2
V=V+VSTEP
30 CONTINUE
C NOW CALCULATE NVHX,AN UPPER BOUND ON THE SIZE OF THE
C DATA WINDOW,IF THIS EXCEEDS NBUF PROGRAM STOPS.
NVHX=IFIX(FSAMP*TSAMP*(TOSQ(1)+XSO(NCHAN)*VINSQ(1))+0.5)
1-IFIX(FSAMP*TSAMP*(TOSQ(1)+XSO(NCHAN)*VINSQ(NVHX))+0.5)+2*NGTM2+1
IF(NVHX.GT.NBUF-127)STOP 'NVHX TOO LARGE'
C CONST IS ARRAY OF CONSTANTS USED IN AP
CONST(1)=1.0/FLOAT(NCHAN)
CONST(2)=-128.
CONST(3)=1.0/128.
CONST(4)=FLOAT(NGTM2-NSTH+1)
CONST(5)=-FLOAT(NGTM2+NSTH)
CONST(6)=0.5
CONST(7)=FSAMP
C INITIALISE AP AND PLACE CONSTANTS IN TOP 7 LOCATIONS
CALL APINIT
CALL APWR
CALL APPUT(CONST,KNCHIN,7,K2)
C CALCULATE THOSE ADDRESSES WHICH ARE CONSTANT IN NMO
C AND SEMBLANCE COMPUTATIONS.
IVINSQ=NCHAN+1
ISSQ=NVHX
IEN=ISSQ+1
C LOOKUP FILE TO BE READ FROM
CALL TRADG(12,FBUF,FSPEC)
IF(LOOKUP(12CH,FSPEC).LT.0)STOP 'LOOKUP ERROR'
WRITE(2) L,NCHAN,H2TART,H1
WRITE(2) XSTART,XSTEP,FSAMP
WRITE(2) T01,T02,T0STEP,TGATE
WRITE(2) VSTEP,V1,V2MIN,V2MAX
WRITE(2) NT0
C-----
DO 500 JTB=1,NT0
NV0=NV(JTB)
C-----
C THE PURPOSE OF NMO IS TO CALCULATE ALL THE NORMAL MOVEOUTS AND
C AUXILIARY INFORMATION NECESSARY TO CALCULATE THE SEMBLANCE FOR
C THE REQUIRED RANGE OF VELOCITIES AT A GIVEN VERTICAL INCIDENCE
C TIME.
C-----
C-----
C OUTPUT ARGUMENTS AND METHOD OF COMPUTATION
C-----
C NMOCS IS ARRAY CONTAINING ALL THE REQUIRED NORMAL MOVEOUT OFFSETS.

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VEL 2

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C ALTHOUGH IT IS A LINEAR ARRAY IT IS USED IN A 'TWO DIMENSIONAL'
C CHANNEL TO SORT ALL THE OFFSETS AS A FUNCTION OF BOTH SEISMIC
C CHANNEL NUMBER AND VELOCITY. THUS THE FIRST NVD VALUES ARE
C THE OFFSETS FOR CHANNEL 1, IN ORDER OF INCREASING VELOCITY, THE NEXT
C NVD VALUES THE OFFSETS FOR CHANNEL 2 AND SO ON. THE OFFSETS
C ARE ROUNDED TO THE NEAREST (INTERPOLATED) SAMPLE AND, FOR A GIVEN
C SEISMIC CHANNEL, ARE RELATIVE TO THE BEGINNING OF THE DATA WINDOW
C THAT IS TO BE TAKEN INTO THE AP TO DO THE PARTIAL SEMBLANCE
C CALCULATIONS FOR THAT CHANNEL, AT TIME JTO.
C
C AT A TIME JTO, AND FOR A CHANNEL JCHAN, THE DATA WINDOW THAT IS
C TAKEN INTO THE AP IN ROUTINE SEMB1 MUST COVER ALL TIMES FROM
C THE CROSSING OF THE SHALLOWEST ARRIVAL TRAJECTORY (VELOCITY OF
C VINVD) TO THE CROSSING OF THE STEEPEST ARRIVAL TRAJECTORY
C (VELOCITY OF VINV(1)). IN ADDITION A HALF GATEWIDTH OF DATA IS
C REQUIRED AT EITHER END.
C
C SET UP INITIAL ADDRESSES
  NVNC=NVD*NCHAN
  NTOT=NVNC+4*NCHAN
  INNO=IVINSQ*NVD
  I1=INNO+NVNC
  I2=I1+NCHAN
  I3=I2+NCHAN
  I4=I3+NCHAN
  I5=I4+NCHAN
  IADD=I5+NCHAN
C TRANSFER TOSQ TO J
  CALL APPUT(VOSQ(JTO),J,K1,K2)
C TRANSFER XSO TO (1,NCHAN)
  CALL APPUT(XSO,1,NCHAN,K2)
C TRANSFER VINSQ TO (IVINSQ,NVD)
  CALL APPUT(VINSQ,IVINSQ,NVD,K2)
C SET UP INITIAL ADDRESSES FOR RESULTS
  IRES=INNO
  IRNV1=IRES+NVD-1
  I1J=I1
  I2J=I2
  CALL APVD
C ITERATE THROUGH CHANNELS
  DO 40 JCHAN=1,NCHAN
C FORM XSO(JCHAN)*VINSQ(JV)+TOSQ(JTO) IN (RES,NVD)
  CALL VMSA(IVINSQ,K1,JCHAN,J,IRES,K1,NVD)
C FORM SORT(RES,NVD)
  CALL VSORT(IRES,K1,IRES,K1,NVD)
C FORM INT(FSAMP)*SORT(TOSQ(JTO)+XSO(JCHAN)*VINSQ(JV))+D.5) USING FSAMP
C IN KFS AND D.5 IN KHF
  CALL VMSA(IRES,K1,KFS,KHF,IRES,K1,NVD)
  CALL VINT(IRES,K1,IRES,K1,NVD)
C FILL IADD WITH VALUE FROM IRNV1 AND NEGATE IT
  CALL VFILL(IRNV1,IADD,K1,K1)
  CALL VNEG(IADD,K1,IADD,K1,K1)
C FILL I1J WITH VALUE FROM IRNV1 (VELOCITY V2)
C AND I2J WITH VALUE FROM IRES (VELOCITY V1)
  CALL VFILL(IRNV1,I1J,K1,K1)
  CALL VFILL(IRES,I2J,K1,K1)
C ADD VALUE IN IADD TO (IRES,NVD), I.E. SUBTRACT MOVEOUT FOR VELOCITY V2
C TO GIVE INNOFS
  CALL VSADD(IRES,K1,IADD,IRES,K1,NVD)
C REDEFINE ADDRESSES SO THAT RESULTS FOR NEXT CHANNEL FOLLOW ON
  IRES=IRES+NVD
  IRNV1=IRNV1+NVD
  I1J=I1J+1
  I2J=I2J+1
40 CONTINUE
C ADD -NSTH*NGTH2 TO (I1,NCHAN) USING VALUE IN KADD1
  CALL VSADD(I1,K1,KADD1,I1,K1,NCHAN)
C ADD -NSTH*NGTH2+1 TO (I2,NCHAN) USING VALUE IN KADD2
  CALL VSADD(I2,K1,KADD2,I2,K1,NCHAN)
C SUBTRACT (I1,NCHAN) FROM (I2,NCHAN) TO FORM (I3,NCHAN)
  CALL VSUB(I1,K1,I2,K1,I3,K1,NCHAN)
C MULTIPLY (I1,NCHAN) BY 1.0/NSBLK FROM KBLKIN TO FORM (I4,NCHAN)
  CALL VSMUL(I1,K1,KBLKIN,I4,K1,NCHAN)
C TRUNCATE (I4,NCHAN) TO INTEGER VALUES
  CALL VINT(I4,K1,I4,K1,NCHAN)
C MULTIPLY (I4,NCHAN) BY -NSBLK FROM KBLK TO FORM (I5,NCHAN)
  CALL VSMUL(I4,K1,KBLK,I5,K1,NCHAN)
C ADD (I5,NCHAN) TO (I1,NCHAN)
  CALL VADD(I1,K1,I5,K1,I1,K1,NCHAN)
C ADD 1.0 FROM TH TO (I1,NCHAN)
  CALL VTSADD(I1,K1,2049,I1,K1,NCHAN)
C ADD (I5,NCHAN) TO (I2,NCHAN) AND MULTIPLY BY 2.0 FROM YTH
  CALL VADD(I2,K1,I5,K1,I2,K1,NCHAN)
  CALL VTSUL(I2,K1,2050,I2,K1,NCHAN)
C FIX ALL VALUES FROM INNO TO I5 PREPARATORY T
C TRANSFER TO HOST
  CALL VFIX(INNO,K1,INNO,K1,NTOT)
C TRANSFER TO HOST ARRAYS
  CALL APVR
  CALL APGET(NMOFS,INNO,NVNC,K1)
  CALL APGET(JST,I1,NCHAN,K1)
  CALL APGET(NVDS,I2,NCHAN,K1)
  CALL APGET(NEL,I3,NCHAN,K1)
  CALL APGET(JBLK,I4,NCHAN,K1)
C SET UP INITIAL ADDRESSES FOR SEMBLANCE CALCULATIONS
  ISENB=IEN+NVD
  ISENB1=ISENB-1
  ISMAX=ISENB+NVD
  ISMAX1=ISMAX+1
  ISTK=ISMAX+2
  CALL APVD
C CLEAR SECTIONS OF AP MEMORY THAT WILL HAVE DATA ADDED:
  CALL VCLR(IEN,K1,NVD)
  NVHG=NVD*NGATE
  CALL VCLR(ISTK,K1,NVHG)
C
C FILL BUFFER S# WITH DATA WINDOW FOR CHANNEL 1
  IF (IREAD(NVDS(1),S#(1),JBLK(1)+1,ICCH).LT.#)STOP'READ ERROR'
C DEFINE INITIAL VALUE OF BUFFER SWITCH
  C SW=.TRUE., READ FROM S#, INTO S1
  C SW=.FALSE., READ FROM S1, INTO S#.
  SW=.TRUE.
C
C ITERATE THROUGH CHANNELS
  JIMO=1
  DO 200 JCHAN=1,NCHAN
    JCHAN1=JCHAN+1
C TRANSFER DATA FROM BUFFER TO AP
    CALL APVR
    CALL IWAIT(ICCH)
    IF (SW) CALL APPUT(S#(JST(JCHAN)),J,NEL(JCHAN),K2)
    IF (.NOT.SW) CALL APPUT(S1(JST(JCHAN)),J,NEL(JCHAN),K2)
    IF (JCHAN.GT.NCHAN)GOTO 200
C TRANSFER NEW DATA FROM DISK TO BUFFER
    IF (SW)ERR=IREAD(NVDS(JCHAN1),S1(1),JSTCH(JCHAN1)+JBLK(JCHAN1),
    1,ICCH)
    IF (.NOT.SW)ERR=IREAD(NVDS(JCHAN1),S#(1),JSTCH(JCHAN1)+
    1,JBLK(JCHAN1),ICCH)

```



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C(1) 2
IF (VARIABLE) GOTO 40
DO 30 J=1,N
N1=N2+1
N2=N1+1
N3=N2+1
N4=N3+1
N5=N4+1
N6=N5+1
N7=N6+1
N8=N7+1
N9=N8+1
N10=N9+1
N11=N10+1
N12=N11+1
N13=N12+1
N14=N13+1
N15=N14+1
N16=N15+1
N17=N16+1
N18=N17+1
N19=N18+1
N20=N19+1
N21=N20+1
N22=N21+1
N23=N22+1
N24=N23+1
N25=N24+1
N26=N25+1
N27=N26+1
N28=N27+1
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N30=N29+1
N31=N30+1
N32=N31+1
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N36=N35+1
N37=N36+1
N38=N37+1
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N40=N39+1
N41=N40+1
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N224=N223+1
N225=N224+1
N226=N225+1
N227=N226+1
N228=N227+1
N229=N228+1
N230=N229+1
N231=N230+1
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N379=N378+1
N380=N379+1
N381=N380+1
N382=N381+1
N383=N382+1
N384=N383+1
N385=N384+1
N386=N385+1
N387=N386+1
N388=N387+1
N389=N388+1
N390=N
```

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CDP 3
C SUBTRACT B FROM C1 AND PUT RESULT IN B
  CALL VSUB(1B,K1,101,1,10,K1,L)
C FIX B AND TRANSFER BACK AS HOST ARRAY INDEX
  CALL VFIX(1B,K1,10,K1,L)
  CALL APWR
  CALL APGET(1INDEX,10,L,K1)
C TRANSFER BUN BACK TO K1
  CALL APPUT(1BUN,K1,K10,K2)
  CALL APWD
C
C HAVING COMPUTED INDEX ARRAY NOW START INTERPOLATION OF TRACE
C
C CLEAR B IN AP
  CALL VCLR(1B,K1,L2INT)
C TRANSFER TRACE TO B
  CALL APWR
  CALL APPUT(1B,L,K2,FSTAP(1))
  CALL APWD
C FIND MEAN VALUE OF TRACE AND PLACE IN AP(1TOP)
  CALL MEANV(1B,K1,1TOP,L)
C SUBTRACT MEAN FROM TRACE
  CALL VNSG(1TOP,K1,1TOP,K1,K1)
  CALL VSADD(1B,K1,1TOP,10,K1,1)
C IF INDEX IS NEGATIVE, NO SCALING OF TRACE
C IF INDEX IS ZERO SCALE TO UNIT R.M.S VALUE
C IF INDEX IS POSITIVE, USE INVERSE ENERGY SCALING (DIVERSITY STACK)
  IF(1INDEX.LT.0) GOTO 100
  CALL RMSQV(1B,K1,1TOP,L)
  IF(1INDEX.GT.0) CALL VSDI(1TOP,K1,1TOP,K1,K1)
C USE 1.0 RESIDING IN 10 (FROM SETAP)
  CALL VDI(1TOP,K1,10,K1,1TOP,K1,K1)
  CALL VSHUL(1B,K1,1TOP,10,K1,L)
C IF INDMUT IS NEGATIVE USE NO MUTING
100 IF(1INDMUT.LT.0) GOTO 200
  CALL VCLR(1B,K1,MUTE(1CHAN))
C TRANSFER MODIFIED TRACE BACK TO SEISM
200 CALL APWR
  CALL APGETA(1B,L,K2,FSTAP(1))
  CALL APWD
  IF(1M.EQ.1) GOTO 350
C TAKE TRANSFORM OF TRACE AND PUT IN C
  CALL RFFT(1B,1C,L2INT,K1)
  CALL RFTSC(1C,L2INT,K0,K1)
  CALL VCLR(1C,K1,K2)
C ITERATE FROM 2 TO N
  DO 300 K=2,N
C MULTIPLY TRANSFORM BY COMPLEX EXPONENTIAL ARRAY
  CALL CMUL(1C2,K2,101,K2,1C2,K2,L2I21,K1)
C MOVE C TO B AND TAKE IN PLACE IFFT
  CALL VMOV(1C,K1,1B,K1,L2INT)
  CALL RFFT(1B,L2INT,K1N)
C TRANSFER SHIFTED TRACE BACK TO SEISM
  CALL APWR
  CALL APGETA(1B,L,K2,FSTAP(K))
  CALL APWD
300 CONTINUE
C PICK REQUIRED ELEMENTS OUT OF SEISM AND PLACE IN STACK
350 DO 400 I=1,L
  STACK(I)=SEISM(1INDEX(I))
400 CONTINUE
C TRANSFER STACK INTO B1 AND ADD TO STACK IN A
  CALL APWR
  CALL APPUT(STACK,1B1,L,K2)
  CALL APWD
  IF(1WRITE(2*L,STACK,JBLK,1DCH1).LT.0) STOP 'WRITE ERROR 1'
  JBLK=JBLK+NBLKTR
  CALL VADD(1A,K1,101,K1,1A,K1,L)
500 CONTINUE
C SCALE STACK BY 1.0/NCHAN
  FNCINV=1.0/NCHAN
  CALL APWR
  CALL APPUT(FNCINV,1TOP,K1,K2)
  CALL APWD
  CALL VSHUL(1A,K1,1TOP,1A,K1,L)
C TRANSFER SCALED STACK BACK TO STACK
  CALL APWR
  CALL IWAIT(1DCH1)
  CALL APGET(STACK,1A,L,K2)
  CALL APWD
  IF(1WRITE(2*L,STACK,JBLK,1DCH1).LT.0) STOP 'WRITE ERROR 2'
  CALL IWAIT(1DCH1)
  CALL CLOSEC(1DCH1)
  RETURN
END

```

Appendix C

DETAILS OF NGSDC NORTHEAST ATLANTIC DATA TAPES

Purchased from:NGSDC,Code D-651,Marine Geology and Geophysics Branch,NOAA/EDS,
Boulder,Co.80302

Copies of correspondence:Finance Secretary,Department of Geological Sciences
(U.S. Dept. Comm. File)

Tapenames:Originals,RQ5046,MT083A-archived in Department of Geological Sciences
Working copies,QGPO4A,QGPO4B-Newcastle computer room; Owner GPO4

Physical characteristics:1600BPI,9 track,odd parity,record length of 80,
50 records/physical block

Data format:coded in EBCDIC in the Merged-merged format.
Tape marks separate each file with two tape
marks at the end of the tape.No volume name,
tape marks or labels at beginning of tape.

MTS mount instructions:\$MOUNT TAPENAME *T* NV C50

Exact data format is as follows

COLUMN DESCRIPTION	FORMAT	IMPLIED DECIMAL POINT BET.COLS.
01-08 Cruise identification	A8	
09-13 Time zone-number to be added for GMT	I5	11,12
14-15 Last two digits of year	I2	
16-17 Month	I2	
18-19 Day	I2	
21-22 Hour - 3pm=15	I2	
23-25 Minutes in decimal tenths	I3	24,25(for mins)
26-33 Latitude in decimal degrees,S is neg	F8.4	
34 42 Longitude in decimal degrees,W is neg	F9.4	
53-57 Uncorrected fathoms	I5	56,57(for fathoms)
58-62 Depth in corrected metres	I5	
63-64 Matthews zone for sound velocity corr.	I2	
66-70 Total magnetic intensity in nT	I5	
71 75 Residual magnetic intensity in nT	I5	
76-80 Free air anomaly in g.u. (1/10 mgal)	I5	79,80(for mgals)
Columns 20,43-52 and 65 are unused		

TAPE FILE		CRUISE	TOTAL	UNCORR	CORR	TOTAL	RESID	FREE
NAME	NO	AGENCY	ID	RECORDS	FATHOMS	METRES	MAG	MAG AIR ANOM
RQ5046	1	US NAVY	01804	5825				5825
(ORIG)			10021	158				158
QGPO4A			10164	490				490
(COPY)			11102	185				185
			11201	2000				2000
			11203	3680				3680
			11205	8040				8040
			11408	5790				5790
			11409	103				103
			11413	115				115
			11423	6299				6299
			11424	1710				1710
			11425	1847				1847
	2	BNDO	69005611	9260	5064	5064	4265	4265
		(FRANCE)	69005621	5060	3004	3004	1398	1398
			70005511	8757	4238	4238	3491	3491
			70005611	5955	2723	2723		
			70005711	9447	3437	3437	4268	4268
			70005811	4159	1524	1524	1858	1858
			72003811	976	869	869		
			73002511	5329	2314	2314	3797	3797
			73002521	3305	1600	1600	2162	2162
	3	L-DGO	V2111	10				7
			V2909	1752				1447
			V2910	4355				3485
			V2911	2577				2142
			V3009	4273				3612
			V3010	4904				3917
			V3011	2963				2297
			V3012	2330				1893
			V2302	2211	1025		1538	1538
			V2305	1158	505		1078	487
	4	L-DGO	C0912	15	9		5	5
			C0913	2322	1123		1045	1045
			V1716	3293	2105		1383	1383
			V2303	9145	4294		5000	5000
			V2304	6191	2433		4448	4448
			V2702	4784	3619		3062	3062
			V2703	6187	4317		3813	3813
			V2704	7992	5538		5024	5024
			V2705	2312	1617		1341	1341
			V2706	5169	3741		3195	3195
			V2801	3294	1993		1839	1839
			V2802	5743	3318		3328	3328
			V2804	8343	6234		5761	5761
			V2805	1648	1489		1463	1463
	5	SCRIPPS	DSDP37GC	37	37		34	34
			DSDP38GC	91	4691		4314	4314
MT083A	1	SCRIPPS	DSDP49GC	4624	2085		2078	2078
(ORIG)	2	US NAVY	LY21A	680	224		535	535
QGPO4B			LY21B	1827	461		1597	1597
(COPY)			LY21C	7451	2287		6086	6086
			LY21E	3924	1387		2635	2635
			LY21F	2531	2349			
			LY12A	573	377		164	164
			LY12B	2022	1385		798	798
			LY32A	10641	9632		853	853
			LY32B	954	643		322	322
	3	L-DGO	V2302	1073	471	471	749	749
			V2303	7830	3669	3669	4304	4304
			V2304	4634	1806	1806	3418	3418
			V2801	427	234	234	191	191
			V2802	4501	2667	2667	2718	2718
			V2803	7490	6228	6228	5993	5993
			V2804	4923	3364	3364	3242	3242
			V2909	1062	729	729	657	657
			V2910	8512	5826	5826	5661	5661
			V2911	4748	1995	1995	4259	4259
			V3009	1061	629	629	725	725
			V3010	7347	5013	5013	4889	4889
			V3011	4930	3327	3327	2908	2908
			V3012	1016	707	707	628	628
	4	L-DGO	V2705	2179	1526	1526	1314	1314
			V2706	2178	1504	1504	1403	1403
			V2702	994	633	633	719	719
			V2703	5853	4086	4086	3617	3617
			V2704	3010	2150	2150	2146	2146

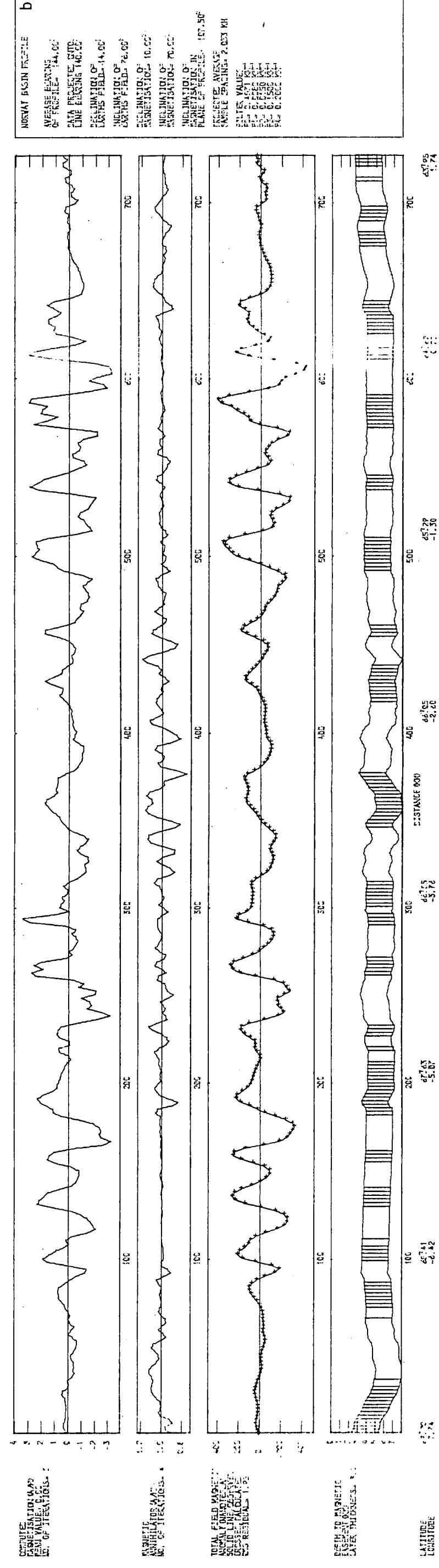
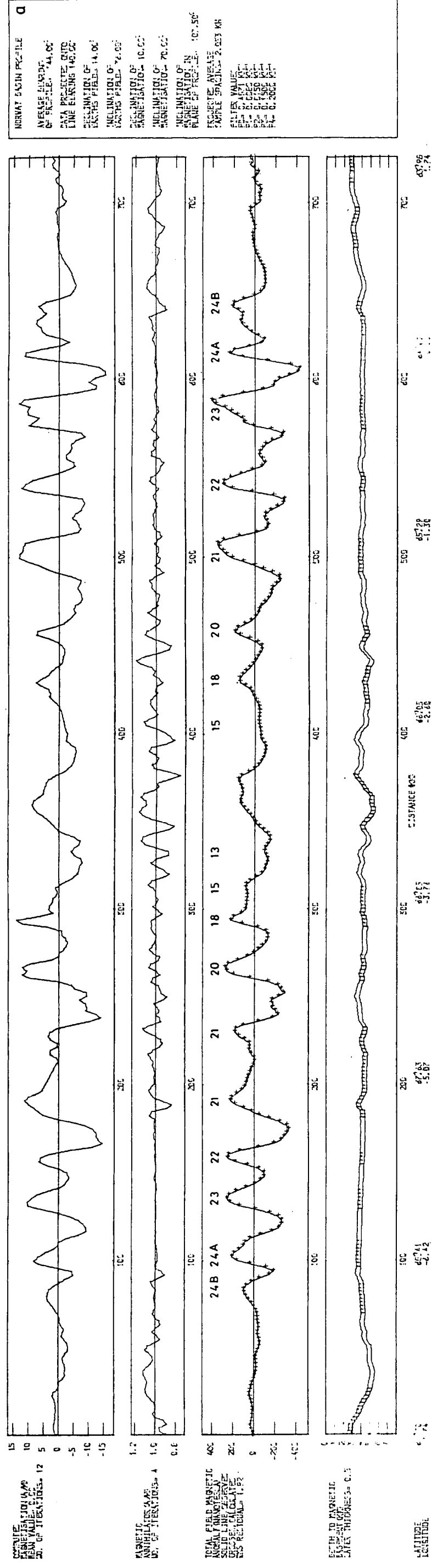


Figure 5.2

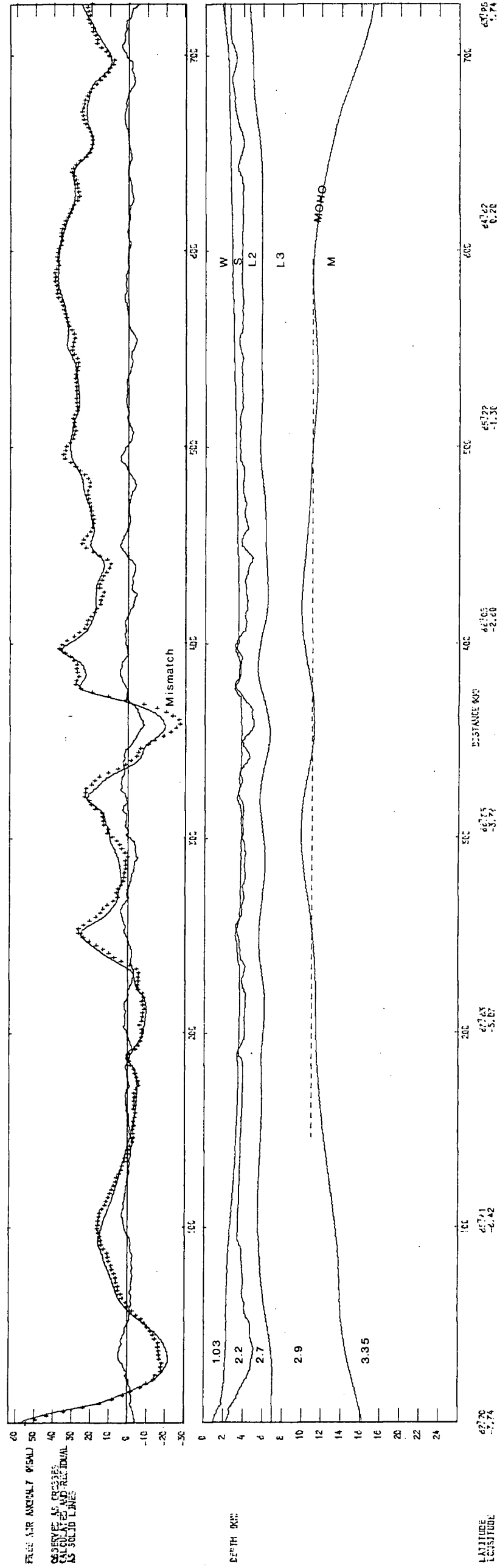
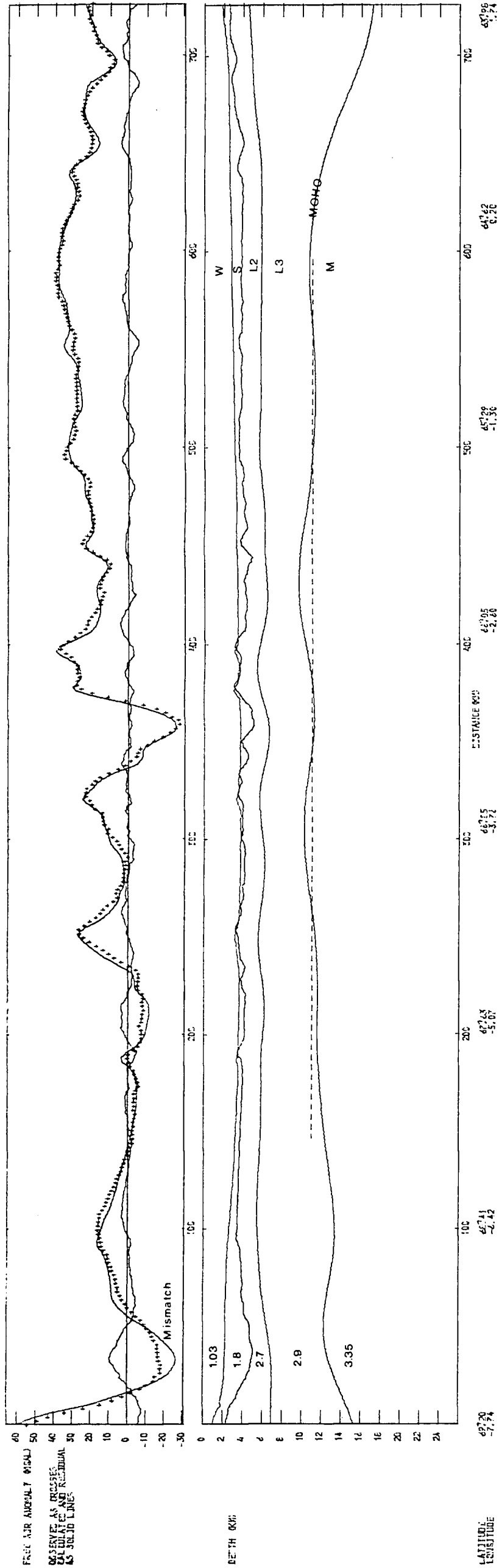


Figure 4.7

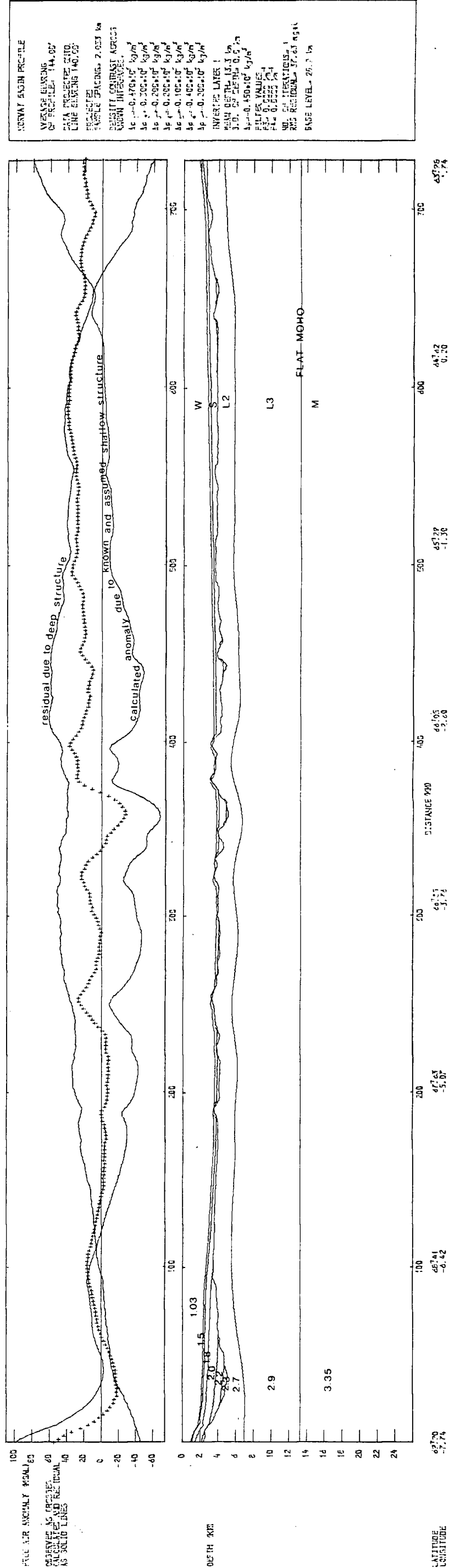
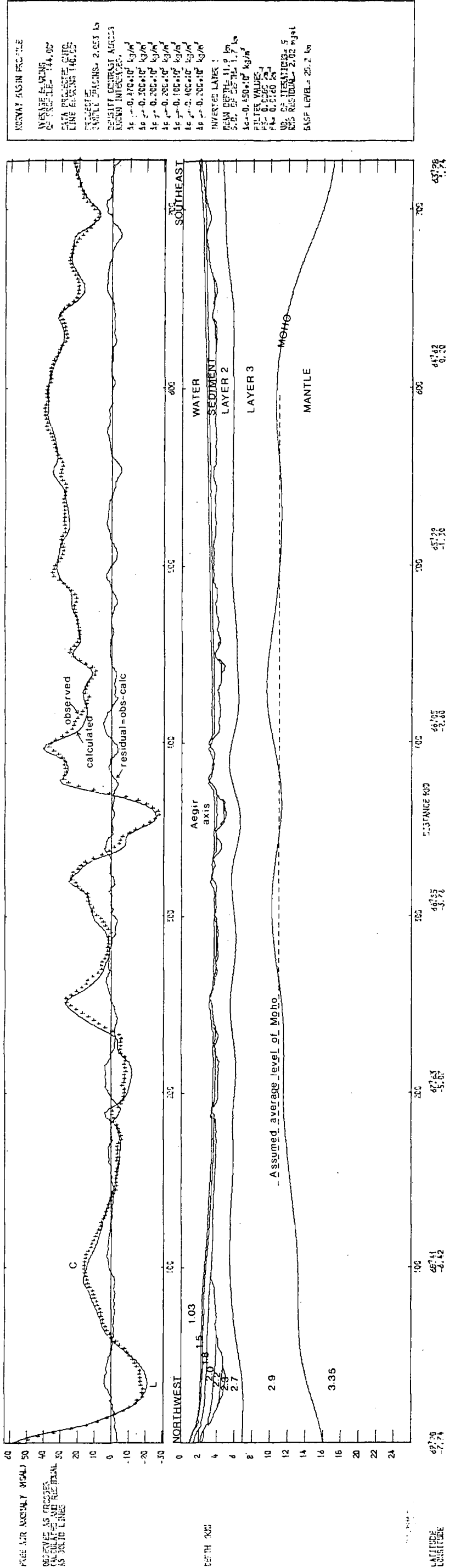


Figure 4.6