The transition between the Sheba Ridge and Owen Basin: rifting of old oceanic lithosphere

Carol A. Stein* and James R. Cochran *Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964, USA*

Summary. Magnetic quiet zones are present along the margins of the entire length of the Gulf of Aden to the Owen fracture zone. This includes the easternmost 300 km between the eastern edges of Arabia and Africa to the Owen fracture zone where old oceanic lithosphere was rifted to form the Sheba Ridge. Within this easternmost region the boundary between the quiet zone and the old oceanic lithosphere is marked by ridge complexes, the Sharbithat Ridge Complex to the north and the Error Ridge Complex to the south. These ridge complexes, which lack a magnetic signature, occupy a structural position similar to the hinge zones at the continental margins to the west. They appear to have formed early in the opening of the Gulf of Aden or perhaps to have been pre-existing features. The boundary between Sheba Ridge and the northern magnetic quiet zone is often marked by an abrupt end of the Sheba Ridge seafloor spreading magnetic anomaly pattern and a sharp basement deepening to the north. The boundary between the East Sheba Ridge and northern magnetic quiet zone becomes less distinct near the Owen fracture zone. This is also accompanied by changes in the East Sheba Ridge, specifically a decrease in magnetic anomaly amplitudes, increase in ridge flank depths and a loss of the 'cooling-curve' ridge flank shape. This may be the result of lower mantle temperatures in the vicinity of the Owen fracture zone. The sediments within the magnetic quiet zone can be divided into a lower disturbed unit and an upper unit consisting of flat-lying reflectors. The disturbance of the lower sediments may have resulted from a period of diffuse extension in the magnetic quiet zone prior to the establishment of the Sheba Ridge spreading centre.

The similarity of the easternmost quiet zone to quiet zones at rifted continental margins leads to the suggestion that these regions were formed by diffuse extension of old oceanic lithosphere. Using a two-layer lithospheric attenuation model and assuming extension during a period of 15 Myr followed by 10 Myr of cooling, the basement depths and heat flow measure-

*Formerly Carol A. Geller. Now at: Department of Geological Sciences, Northwestern University, Evanston, Illinois 60201, USA.

ments can be adequately matched. The modelling implies 45 per cent crustal extension in the quiet zone. The amount of extension calculated is compatible with documented motion between Arabia and Africa. The old oceanic lithosphere thus must have been substantially thinned to a thickness similar to 10—14 Myr old ocean before seafloor spreading was initiated at the Sheba Ridge.

Introduction

The rifting of continental lithosphere and subsequent seafloor spreading creates a characteristic set of features commonly recognized at 'passive' margins. Specifically, between the clearly defined continental and oceanic lithosphere is a zone resulting from the margin's rifting and subsequent evolution, which is frequently called a magnetic quiet zone because of the generally low amplitude and lack of correlation of the magnetic anomalies. The transition from undisturbed continent to the magnetic quiet zone is marked by the hinge zone, a major structural boundary across which there is a rapid basement deepening. The magnetic quiet zone often appears to consist of thinned and faulted continental crust, as found in the Bay of Biscay (de Charpel *et al.* 1978; Montadert *et al.* 1979), the southern margin of Australia (Talwani *et al.* 1979) and the Newfoundland margin (Keen & Barrett 1981). At some margins, the rifting process is associated with large-scale dyke intrusion and volcanic activity, such as found in the westernmost Gulf of Aden and Red Sea (Beydoun 1970; Coleman *et al.* 1975, 1979) and on the Outer Voring Plateau, Norway, where layered volcanic flows have been detected from seismic data (Mutter, Talwani & Stoffa 1982) and from drilling results (Talwani & Udintsev 1976). The magnetic quiet zone—oceanic lithosphere boundary is usually sharp, often defined by a basement depth discontinuity and magnetic and gravity gradients (Talwani & Eldholm 1973).

The Gulf of Aden (Fig. 1) is a young ocean basin resulting from the separation of Africa and Arabia beginning in the late Oligocene or the earliest Miocene (Somaliland Oil Exploration Co., Ltd 1954; Azzaroli & Fas 1964; Beydoun 1982). The eastern edge of continental rifting, which defines what we shall call the 'geographical' Gulf of Aden, is marked by a line extending from the edge of the continental shelf near Ras Sharbithat (Arabia) to the eastern edge of the submerged peninsula containing the island of Socotra (Africa). However, from a tectonic point of view the Gulf of Aden extends an additional 300 km east to the Owen fracture zone. Seafloor spreading is occurring throughout the Gulf of Aden from Afar to the Owen fracture zone at the Sheba Ridge spreading centre which is divided into the West and East Sheba Ridge by the major Alula—Fartak fracture zone at 51°E. The East Sheba ridge is offset right-laterally by 300 km from the Carlsberg Ridge at the Owen fracture zone.

Ridge complexes extend eastward from both Ras Sharbithat and Socotra to the Owen fracture zone; the Sharbithat Ridge Complex to the north, and the Error Ridge Complex to the south. The ridge complexes separate crust associated with the Sheba Ridge from older oceanic basins, the Owen Basin (described by Whitmarsh 1979) to the north, and the northern Somali Basin (described by Bunce *et al.* 1967) to the south. At the intersection of both ridge complexes with the Owen fracture zone, the trend of the fracture zone changes from a nearly N—S trend along the boundaries of the Owen and northernmost Somali basins to a NNE—SSW trend between the two ridge complexes. This direction is parallel to the present Sheba Ridge spreading direction and the trend of both ridge complexes is perpendicular to the present Sheba Ridge spreading direction. The topographic expression of the Owen fracture zone is subdued near the intersection with Sharbithat Ridge and immediately to the south it is completely buried by sediments (Fig. 2).
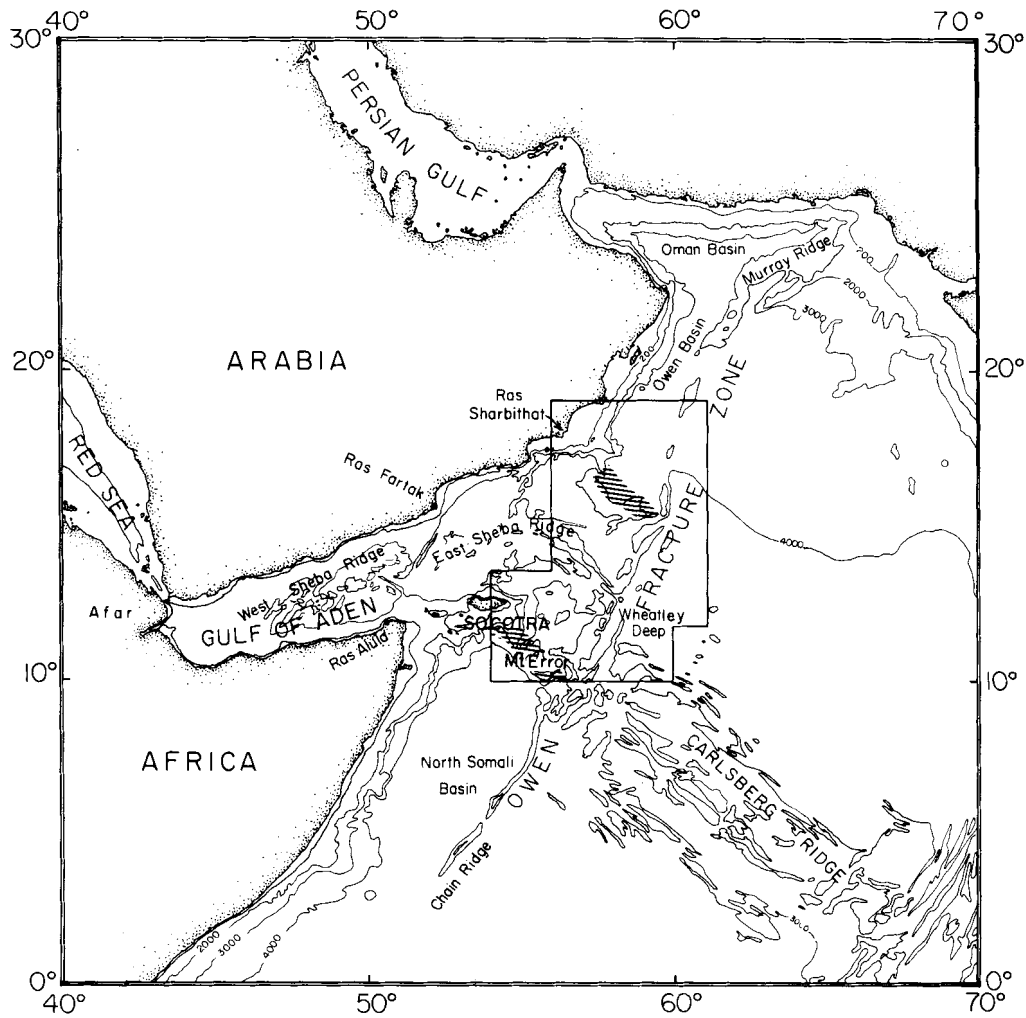
Figure 1. Location and bathymetry of the Gulf of Aden—East Sheba Ridge area and surrounding regions: Study area is outlined. Bathymetric contours in metres from the *General Bathymetric Chart of the Oceans* (GEBCO), 1975. Locations of Sharbithat and Error Ridge complexes are represented by the hatched lines.

Seafloor spreading at the Sheba Ridge axis has produced correlatable magnetic anomalies with NNE-trending fracture zones, identified by Laughton, Whitmarsh & Jones (1970) and Cochran (1981) as indicating organized seafloor spreading since magnetic anomaly 5 time (about 10 Myr BP) through most of the Gulf of Aden. It has also been suggested that the Red Sea and the Gulf of Aden opened in two stages, an initial seafloor spreading phase 30—15 Myr BP and a second phase beginning 5 Myr BP (Girdler & Styles 1978, 1982; Styles & Hall 1980; Girdler *et al.* 1980) with oceanic lithosphere, in some places, as close as 20 km from the coastline. We do not agree with this interpretation (see Cochran 1981, 1982a and Girdler & Styles 1982 for a complete discussion).

Inside the geographic Gulf of Aden, magnetic quiet zones are located between the oldest identifiable magnetic anomaly (usually anomaly 5) and the steep continental margins, with boundaries marked by basement depth discontinuities (Cochran 1981). Magnetic quiet zones
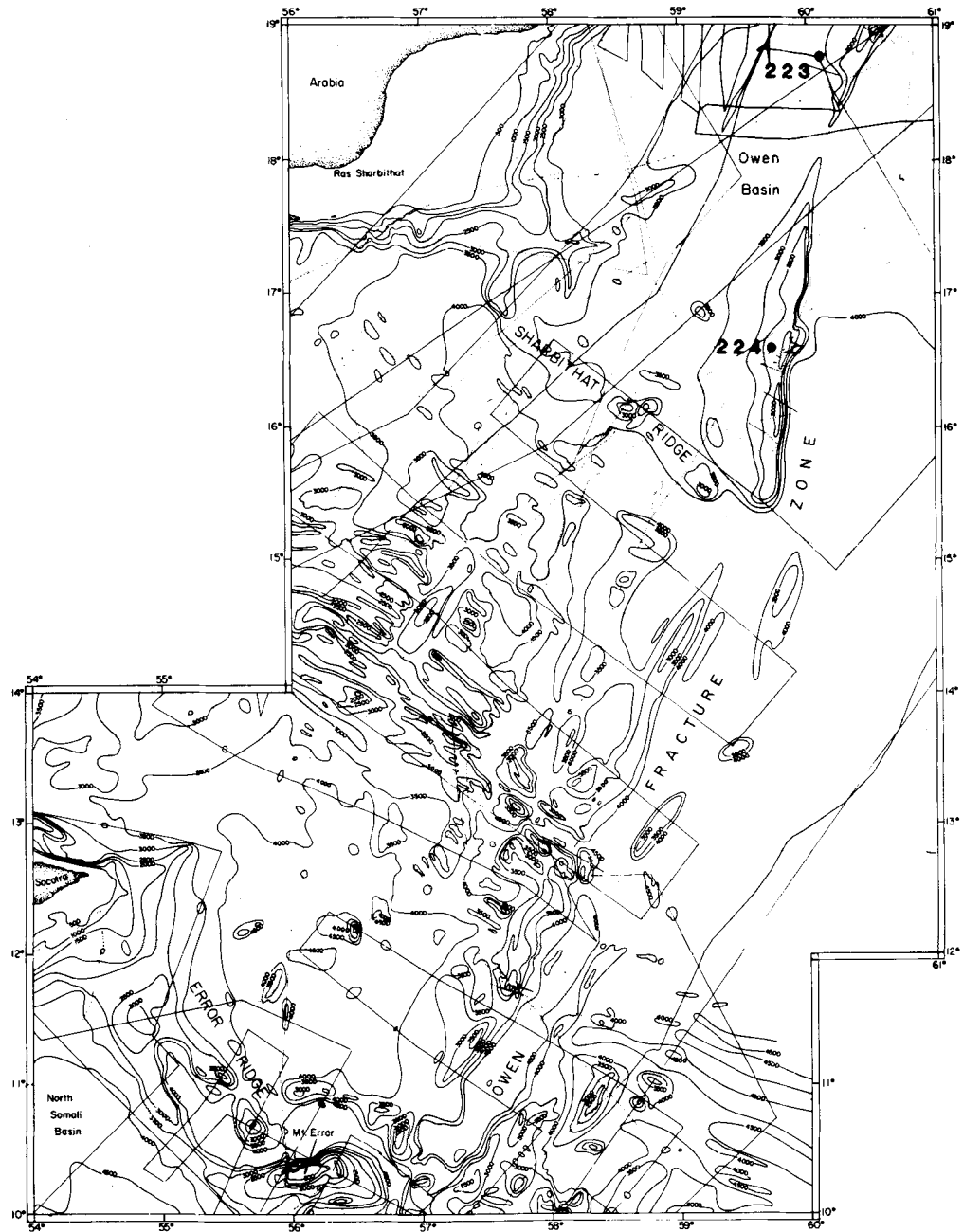
Figure 2. Bathymetry of the study area (Fig. 1). Contour interval is 500 m. Location of ship tracks used in constructing this map is shown. Contours from GEBCO (1975) were used where data are not available. Circles with dot inside show the location of DSDP sites.

exist not only along the continental margins of the geographical Gulf of Aden but also extend 300 km eastward to the Owen fracture zone with the position of the stable continental crust replaced by the oceanic crust of the Owen and Somali Basins (Cochran 1981). The position of the continental hinge zones is occupied by the Sharbithat and Error Ridge Complexes (Cochran 1981).
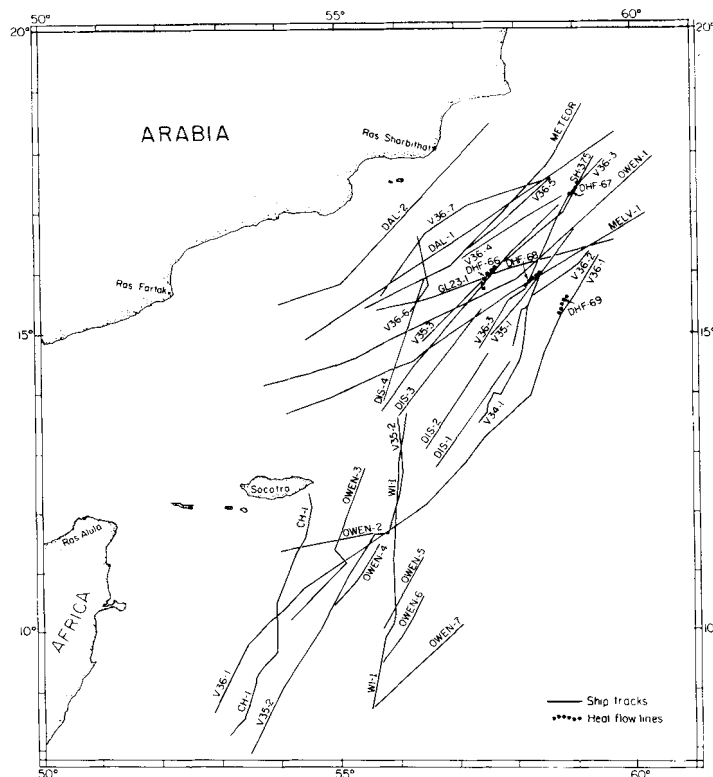
Figure 3. Location of ship tracks and heat flow stations. Track identifications correspond to those for profiles in subsequent figures.

We propose that the easternmost magnetic quiet zones were formed by rifting of old oceanic lithosphere during the early stages of the opening of the Gulf of Aden. The purpose of this paper is to study the boundary between the Sheba Ridge and the older oceanic crust to the north and south. A comparison of these oceanic transition zones with the magnetic quiet zones associated with the rifted continental margins should yield information on the process of lithospheric rifting. We will concentrate on examining the northern magnetic quiet zone, the transition between the East Sheba Ridge and the Owen Basin, because there is much better data coverage there than in the south.

Northern magnetic quiet zone

The northern boundary of the magnetic quiet zone east of the geographical Gulf of Aden is formed by the Sharbithat Ridge, a 300 km long continuous feature which extends from the Owen fracture zone to a complex intersection with the Arabian continental rise east of Ras Sharbithat (Fig. 2). Sharbithat Ridge appears to act as a relatively unbroken barrier to sediment transport from the north (Cochran 1981). A number of lower basement ridges, also elongated perpendicular to the Sheba Ridge (Fig. 2), are also present within the magnetic quiet zone south of the Sharbithat Ridge (Fig. 4), but these are not continuous and are generally less than 40 km long.

Free-air gravity anomaly data contoured at 10 mgal intervals from the areas of the East Sheba Ridge and the Owen Basin are presented in Fig. 5. The dominant features of the gravity map are the steep gradients and large amplitude positive and negative anomalies
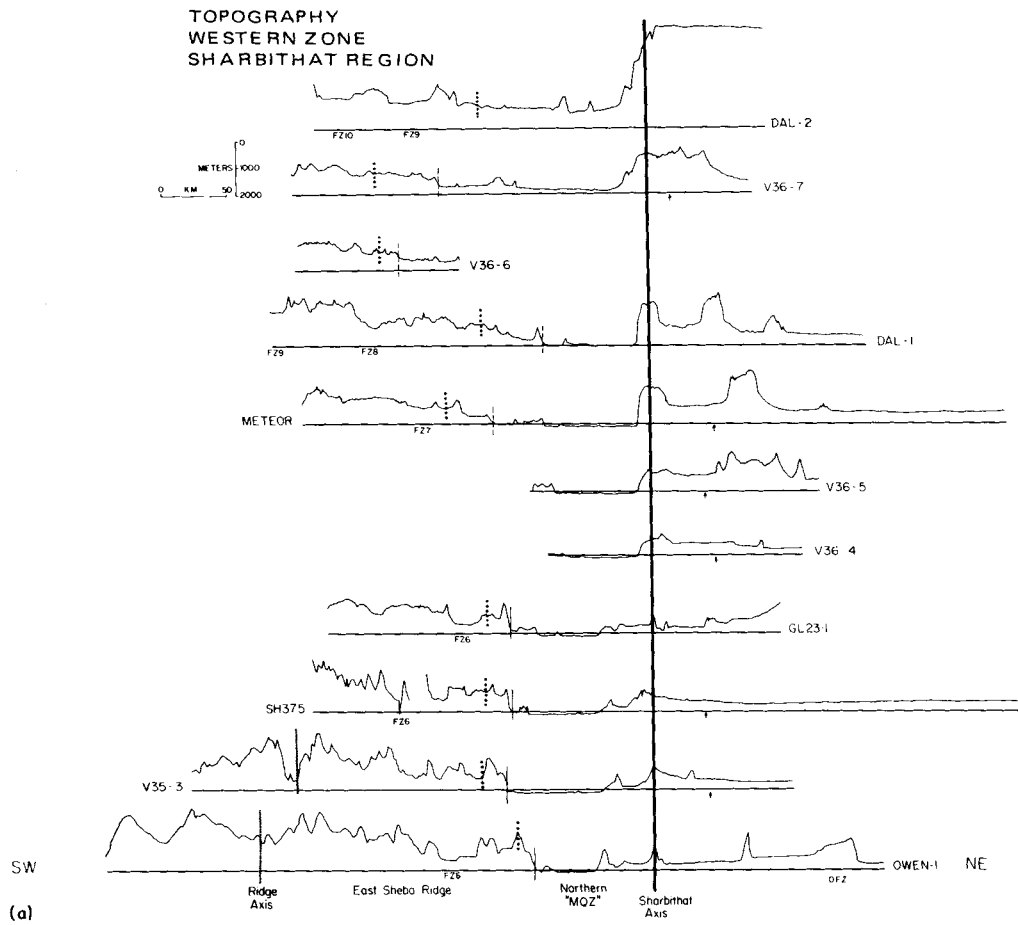
Figure 4. (a) Topographic profiles from the western zone across the Sharbithat Ridge Complex. Location of all profiles is shown in Fig. 3. Profiles have been projected at N32°E and aligned along the Sharbithat Ridge axis. The thin horizontal line through each profile is the 4000 m level. The vertical hatched line marks the East Sheba Ridge axis. The dotted line represents the expected location of magnetic anomaly 5 based on 1.5 cm yr^{-1} spreading rate on the northern flank. The short thin vertical line represents the boundary between Sheba Ridge and the magnetic quiet zone and is dashed where the boundary is not as certain. Fracture zones named according to Cochran (1981). The small arrow indicates the location of the free air gravity low associated with the northern flank of Sharbithat for ship tracks with gravity data. Note that the shape of Sharbithat Ridge changes from track to track but the ridge complex is present on all profiles. (b) Topographic profiles from the eastern zone across the Sharbithat Ridge Complex. See caption of (a) for explanation of symbols. Profiles have been projected along N32°E.

associated with the Owen fracture zone. The Sharbithat Ridge is marked by a relative gravity high, the magnitude varying with the height of the ridge. A large broad gravity low with maximum amplitude of about −50 mgal is located on the northern flank of Sharbithat Ridge and is not associated with any acoustic basement feature. (The gravity minimum locations are shown by an arrow on the topography profiles in Fig. 4.) The gravity low extends westward from 59°E to near 58°E in the 'bight' between two ridges projecting from the continental margin. These ridges separate this Sharbithat gravity low from similar amplitude 'edge effect' lows extending along the continental margins of the Owen Basin and Gulf of Aden (Fig. 5) near the 3–3.5 km isobaths. East of 59°E a gravity low exceeding

TOPOGRAPHY
EASTERN ZONE
SHARBITHAT REGION

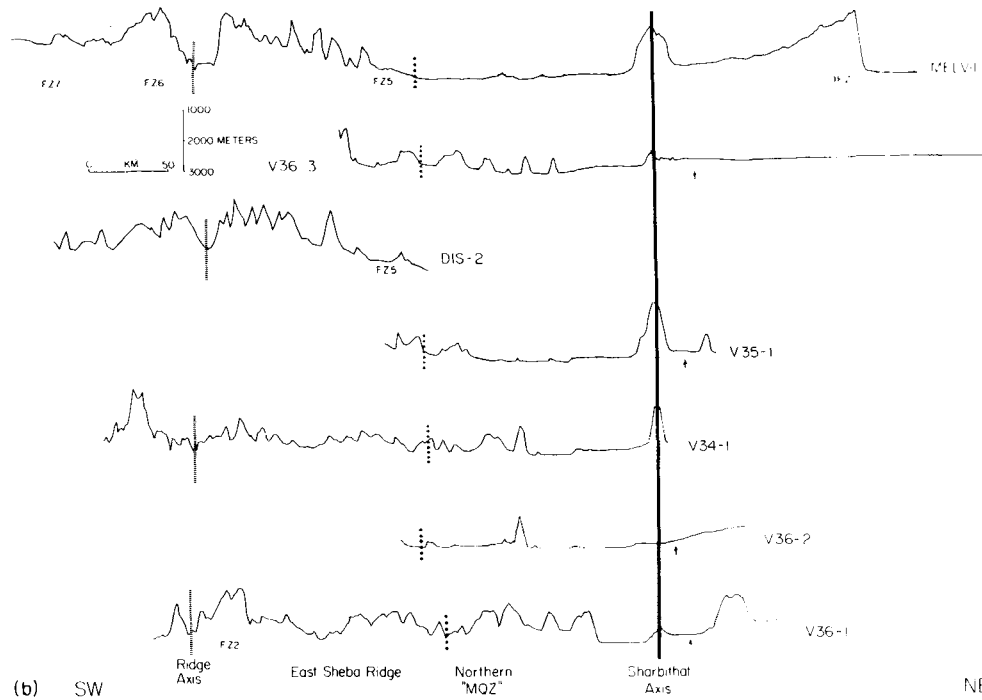F21 F26 F25 MELVI

1000
2000 METERS
KM 50 3000 V36 3

DIS-2
FZ5

V35-1

V34-1

V36-2

V36-1
F22

(b) SW Ridge Axis East Sheba Ridge Northern "MQZ" Sharbithat Axis NE

Figure 4 – *continued*

−40 mgal is also associated with Sharbithat Ridge, but is here located south of the ridge complex over a deep trough which parallels it to the south. Sediment thicknesses in the trough exceed 2 s in places. This gravity low is narrower with much steeper gradients than the gravity low to the west. There is also a lower amplitude gravity low (about −10 to −20 mgal) over the magnetic quiet zone south of Sharbithat Ridge in the western zone where sediment thicknesses average 0.5 s.

The gravity pattern over the magnetic quiet zone in some profiles (Fig. 6) has the appearance of a broad low 100–150 km wide extending from near the boundary between the Sheba Ridge and magnetic quiet zone to well north of Sharbithat Ridge. The gravity high of the Sharbithat Ridge is superimposed on the broad low. This is similar to the gravity anomaly pattern resulting from flexural loading. However, simple calculations show that for reasonable basement and sediment densities and a range of lithospheric rigidities, the gravity anomalies observed over the quiet zone cannot result simply from flexural loading (Fig. 6). Calculations also show that the gravity anomaly cannot result from a thermal isostatic 'edge effect' (Karner & Watts 1982). Specifically, thermal isostasy (Karner & Watts 1982) predicts a positive anomaly of similar amplitude to the negative, which is not observed, and a longer wavelength anomaly than is observed.

The highly stratified sediments of the Owen Basin abut against the northern flank of the Sharbithat Ridge without any evidence of tectonic disturbance (Fig. 7). This observation suggests either that the Sharbithat Ridge at least in part predates the beginning of rifting in the late Oligocene or that it formed early in the development of the quiet zone, prior to
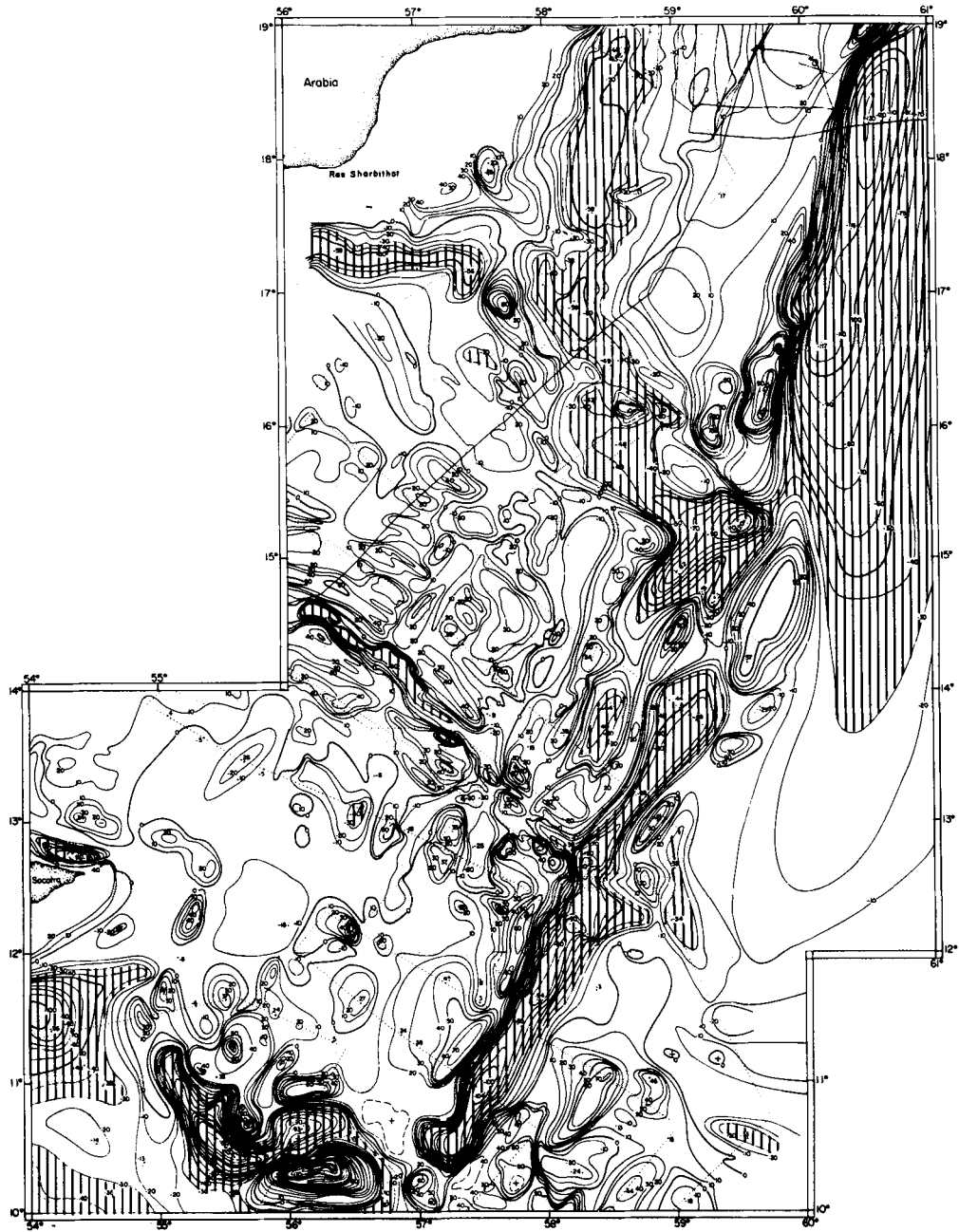
Figure 5. Free-air gravity anomalies in the study area (Fig. 1) contoured at 10 mgal intervals. Areas of gravity anomalies less than −30 mgal are shaded. Data points are shown as fine dots. Anomalies are referred to the 1930 International Ellipsoid (flattening = 1/297).

the deposition of the upper second of Owen Basin sediments. Sediments within the magnetic quiet zone and on the southern flank of Sharbithat Ridge can be divided into two units. The underlying sediments, which can be up to 0.6 s thick, show signs of disturbance and in places are upturned on to Sharbithat Ridge (Fig. 7). Although in some places these upturned layers
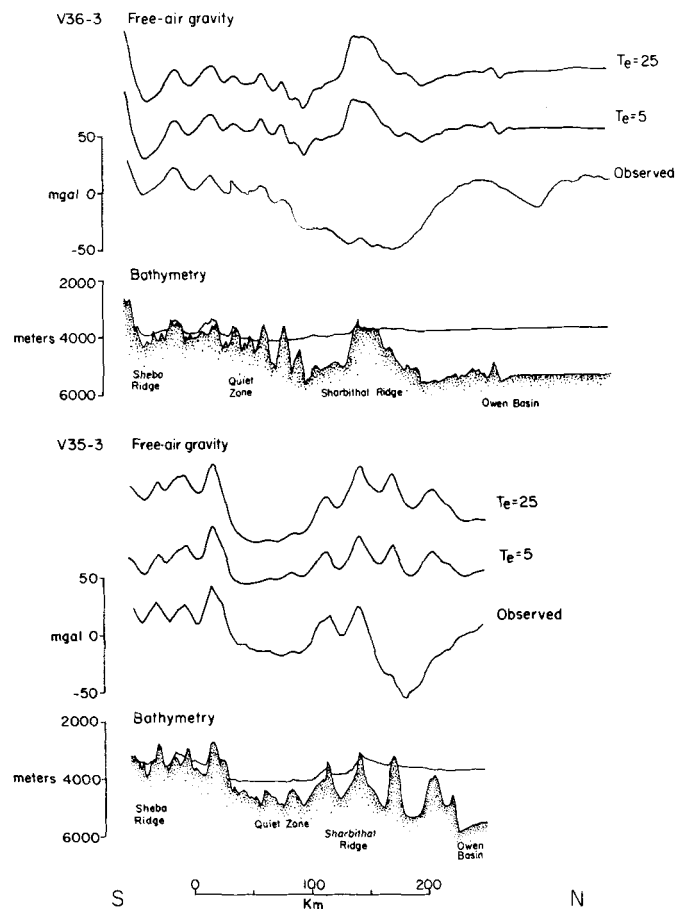
Figure 6. Observed and calculated free-air gravity anomalies across the northern magnetic quiet zone and Sharbithat Ridge. Free-air gravity anomalies are computed for profiles V35-3 and V36-3 (see Fig. 3 for locations) for 5 and 25 km elastic plates with the deflection of the crust−mantle boundary due to the load of the sediments and topography included. For profile V35-3 deflections due to the load of Sharbithat Ridge are up to about 2 and 1 km for the 5 and 25 km elastic thicknesses respectively. For profile V36-3 deflections due to the load of Sharbithat Ridge are up to 1 and 0.6 km for the 5 and 25 km elastic plate. Note that the gravity low on the northern flank of the Sharbithat Ridge is not reproduced in the modelling.

on the ridge may be the result of draping, the general appearance throughout the quiet zone suggests tectonic disturbance. The upper sediments, which are up to 0.5 s thick, are characterized by flat-lying reflectors. They appear in general undisturbed although there are a few minor faults with small offsets. The division of the sediments into two units may be a reflection of the tectonic history of the quiet zone. The upper unit reflects the present relative stability of the quiet zone, while the disturbance of the lower unit may be the result of an earlier active tectonic period associated with the early opening of the Gulf of Aden.

The boundary between the East Sheba Ridge and the magnetic quiet zone is well defined for all profiles more than about 130 km west of the Owen fracture zone (western zone) and is characterized by an increase in basement depth (Fig. 8a) and a decrease in the magnetic anomaly amplitude just north of magnetic anomaly 5 (Fig. 7a). The magnetic anomalies within the magnetic quiet zone are generally lower amplitude than magnetic anomalies
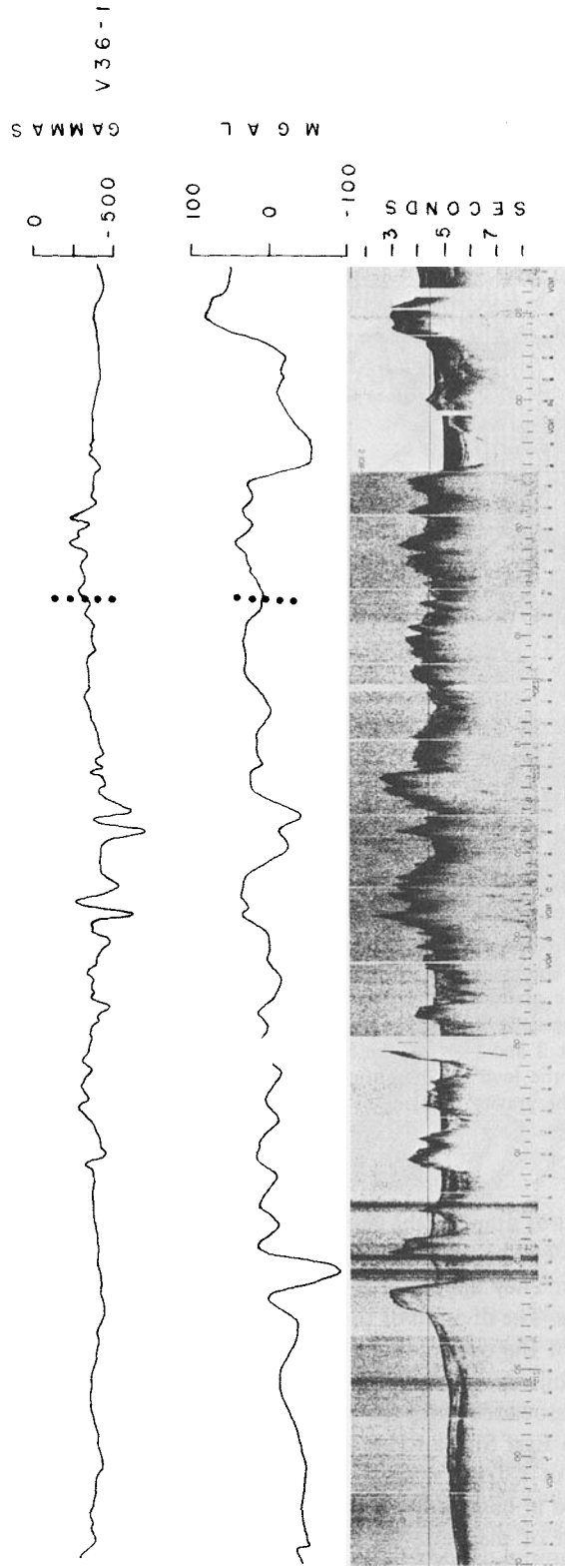
EASTERN ZONE (near Owen F.Z.)

Figure 7. Seismic reflection profiles with free-air gravity anomaly and total intensity magnetic anomaly profiles along track from the easternmost Gulf of Aden. Location of profiles shown in Fig. 3. The profiles are aligned on the Sharbithat Ridge Complex axis. The dotted line represents the predicted location of magnetic anomaly 5 based on a 1.5 cm yr^{-1} spreading rate on the northern flank of the East Sheba Ridge. Note that in the western zone the predicted and actual location of magnetic anomaly 5 agree well and are near the abrupt basement deepening to the north into the quiet zone but in the eastern zone the high and rough basement topography extends well north of the predicted magnetic anomaly 5 location. Also in the eastern zone there is no correlatable magnetic anomaly pattern for profile V36-1 and there is a clear departure in the shape of the East Sheba Ridge from that predicted from thermal cooling models.
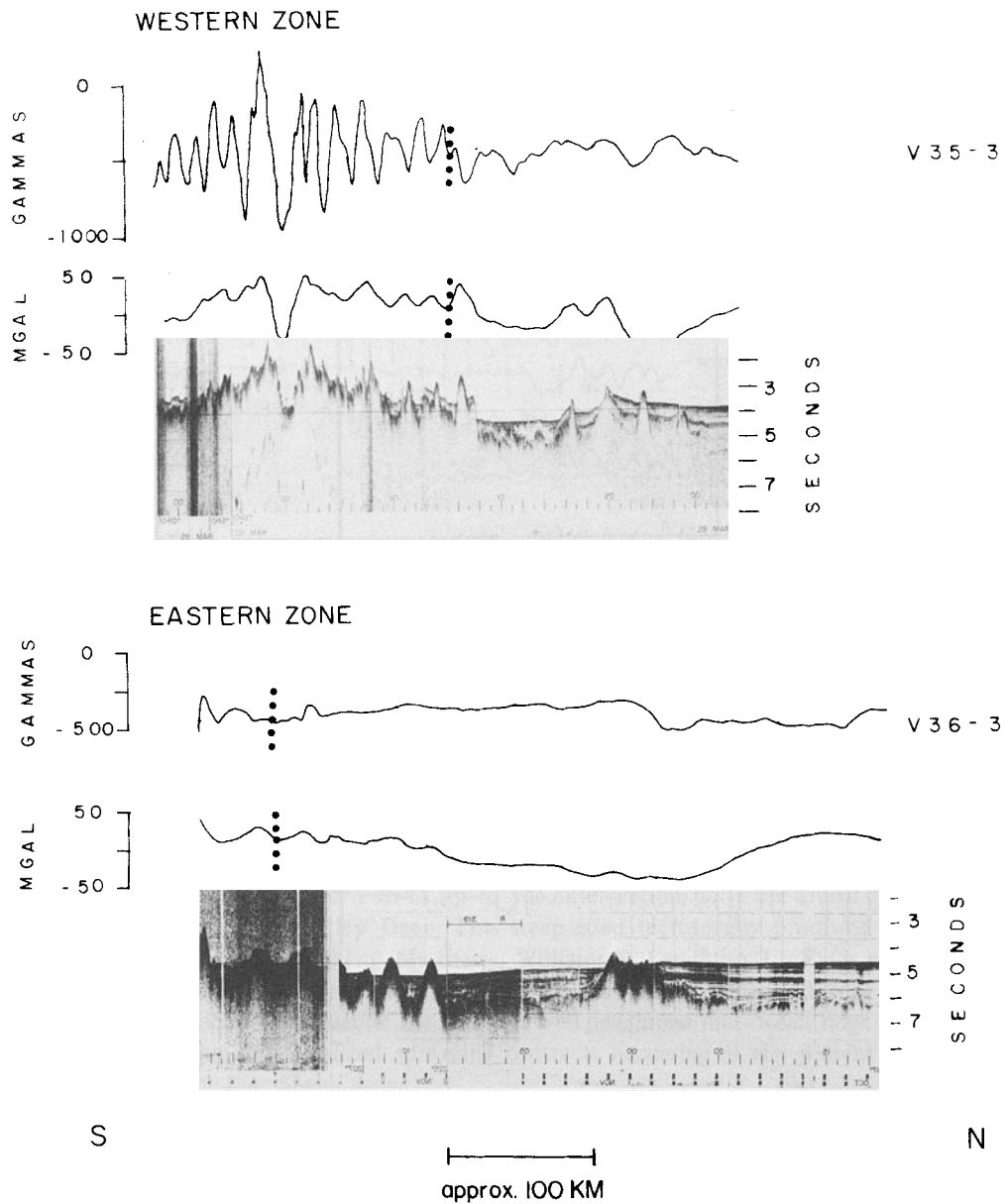
WESTERN ZONE

EASTERN ZONE

S

N

approx. 100 KM

Figure 7 – *continued*

formed in the last 10 Myr by seafloor spreading on the Sheba Ridge, and can only be traced between profiles where the tracks are close together (for example, profiles SH375 and V35-3, Fig. 8a). The far western section of the Sharbithat Ridge, made up of the N–S trending ridge on the continental slope has no magnetic anomaly associated with it. The main portion of the ridge is characterized by a broad low amplitude (< 150 gamma) magnetic anomaly.

The amplitude of the seafloor spreading magnetic anomalies decrease systematically eastward toward the Owen fracture zone starting with Profile DIS-3, about 190 km west of the Owen fracture zone. Near the Owen fracture zone, only the axial anomaly is clearly defined
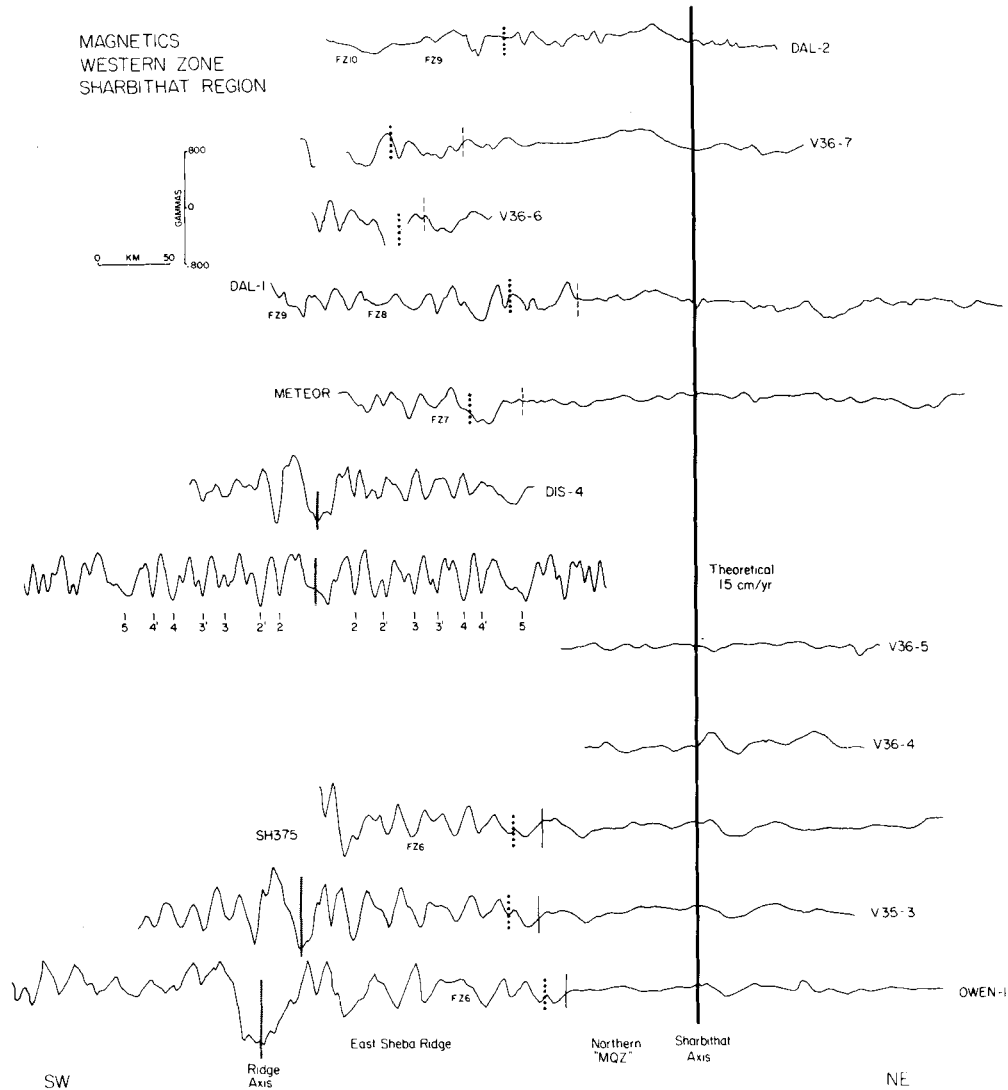
Figure 8. (a) Total intensity magnetic anomaly profiles across the Sharbithat Ridge Complex from the western zone. Location of all profiles shown in Fig. 3. All profiles have been projected along N32°E. Theoretical seafloor spreading anomaly sequence is also shown generated using LaBrecque, Kent & Cande (1977) reversal time-scale and inclination 17°, declination 0°. Symbols are the same as in Fig. 4. (b) Total intensity magnetic anomaly profile across the Sharbithat Ridge Complex from the eastern zone.

on profile V36-1 located about 90 km west of the Owen fracture zone (Fig. 8b). Topographic control from cross-lines clearly show that this line is not situated in a fracture zone (Fig. 2).

The decrease in the amplitude of the magnetic anomalies toward the Owen fracture zone is associated with an increase in the depth of the ridge crest and, between the Owen fracture region and approximately 130 km to the west (in the eastern zone) a loss of the characteristic mid-ocean ridge shape. In this region the East Sheba Ridge does not display the
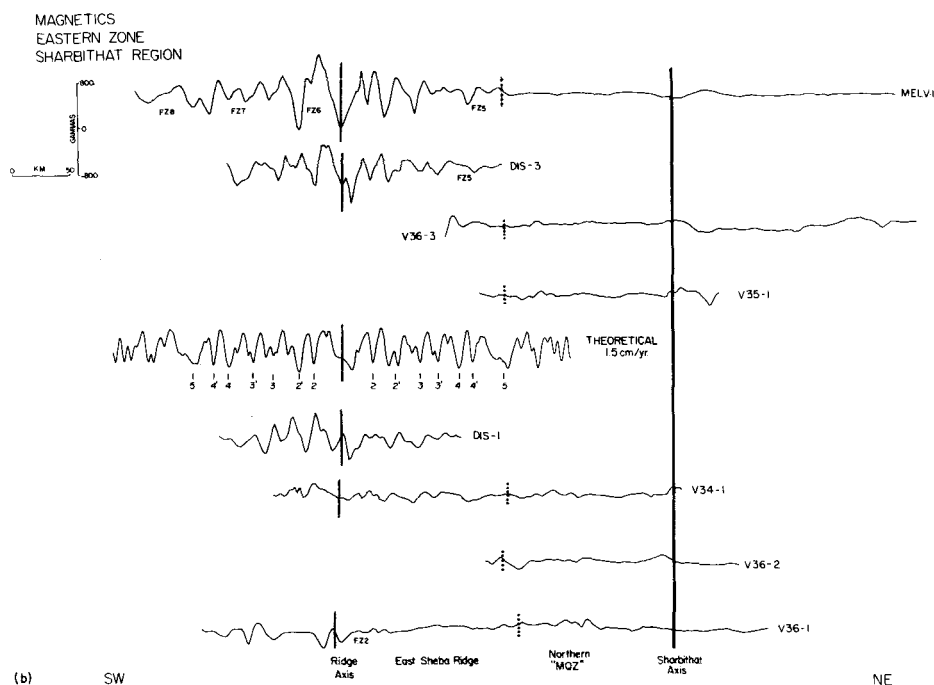
MAGNETICS
EASTERN ZONE
SHARBITHAT REGION

Figure 8 – *continued*

regular increase in depth away from the ridge crest typical of mid-ocean ridge spreading centres, but rather consists of a zone of rough bathymetry with a fairly constant average depth of about 3500 m. Sheba Ridge does, however, retain a well developed rift valley (Fig. 4b, profiles V34-1 and V36-1) up to the intersection with the Owen fracture zone which is marked by Wheatley Deep. This steep-sided tectonically produced trough with water depths exceeding 5600 m (Matthews, Williams & Laughton 1967), is similar to the depressions commonly observed at the intersection of fracture zones and ridge crests. These observations indicate that, despite the lack of a well developed mid-ocean ridge morphology, seafloor spreading at a localized spreading centre is occurring at the present time to the eastern end of the East Sheba Ridge.

In the eastern zone, the boundary between the East Sheba Ridge and magnetic quiet zone is not well defined. The ridge and trough Sheba Ridge morphology does still terminate at a down-to-the-north increase in depth, but this depth increase is located significantly north of the predicted location of the 10 Myr isochron (Figs 4b and 8b). The change in morphology also appears to be offset left laterally across fracture zones while the ridge crest is offset right laterally (see Fig. 2). Thus, near the Owen fracture zone it becomes difficult to distinguish 'Sheba Ridge' crust from 'magnetic quiet zone' crust except in the vicinity of the ridge crest.

Southern magnetic quiet zone

Far fewer data are available for the southern flank of the Sheba Ridge. A magnetic quiet zone and a ridge complex, Error Ridge, occupy analogous positions to the quiet zone and Sharbithat Ridge to the north. The Error Ridge complex is made up of two parallel ridges separated by a 30 km wide trough (Fig. 2) containing up to 1 km of sediment. The double
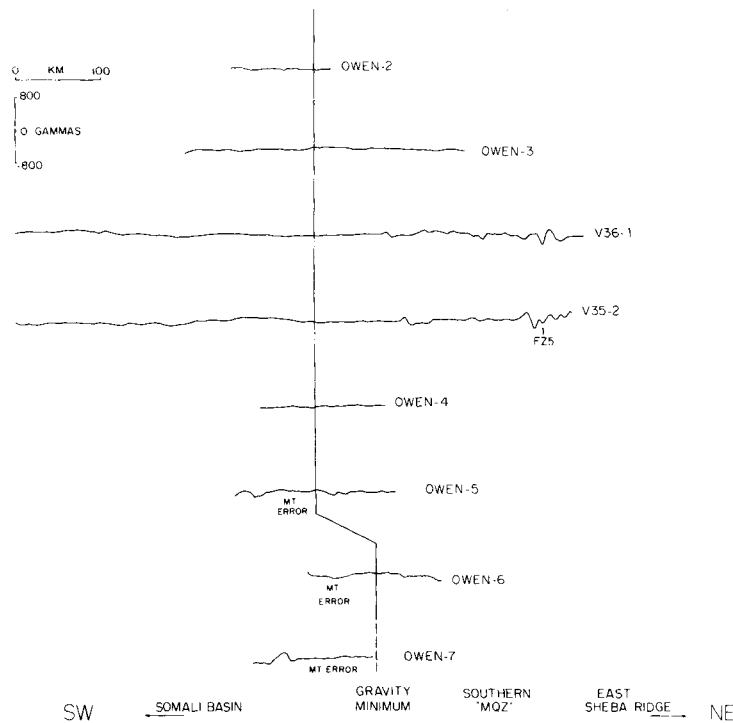
MAGNETICS
ERROR REGION

O   KM   100

800

O GAMMAS

-800

OWEN-2

OWEN-3

V36-1

V35-2

FZ5

OWEN-4

OWEN-5

MT
ERROR

OWEN-6

MT
ERROR

OWEN-7

MT ERROR

SW   SOMALI BASIN   GRAVITY   SOUTHERN   EAST
                              MINIMUM   "MQZ"   SHEBA RIDGE   NE

Figure 9. Total intensity magnetic anomalies across the Error Ridge Complex. Profiles have been projected N32°E. Profile locations shown in Fig. 3. The heavy line through each profile represents the location of the gravity minimum associated with the trough between the two ridges of Error Ridge Complex.

ridge structure terminates in the southeast at Mt Error, a large guyot capped by limestone (Laughton 1966), which reaches to within 200 m of the sea surface.

Neither Error Ridge, the southern magnetic quiet zone, nor the northern Somali Basin is associated with significant magnetic anomalies (Fig. 9). The northern Somali Basin is generally considered to be a northern continuation of the southern Somali Basin where magnetic anomalies M25 to M9 (Oxfordian—Hautervian) have been identified by Rabinowitz, Coffin & Falvey (1983). However, no magnetic anomalies have been identified in the northern Somali Basin and none are apparent in the two long N–S profiles (V36-1 and V35-2) shown in Fig. 9. Also the northern Somali Basin has significantly deeper basement and more sediment accumulation than the southern part (Bunce *et al.* 1967).

A prominent free-air gravity anomaly low with values less than −130 mgal is associated with the trough in the middle of Error Ridge (Fig. 5) and appears to be of a greater amplitude than can be explained simply by basement relief. This gravity low is, however, unlike the low associated with the northern quiet zone because it is directly associated with a basement feature. A broad gravity low similar to that associated with the northern flank of the Sharbithat Ridge is not found at Error Ridge. The regional level of the free-air gravity anomaly field increases from about −50 mgal in the northern Somali Basin about 50 km south of Error Ridge to about 0 mgal in the southern magnetic quiet zone.

There are insufficient data to define the nature of the boundary between the southern magnetic quiet zone and the Sheba Ridge or to determine whether the southern magnetic quiet zone changes character from west to east in a manner similar to the northern quiet zone. However, comparison of the bathymetry suggests that such a change may exist.

Model of lithospheric rifting

The magnetic quiet zones and ridge complexes at the eastern end of the Sheba Ridge are in the same structural position as the magnetic quiet zones and 'hinge zones' to the west within the geographical Gulf of Aden, formed by the rifting of the continental lithosphere of Arabia and Africa. The similarity of the features near the continental margins inside the Gulf of Aden and other rifted margins to the structures found between East Sheba Ridge and the oceanic basins to the north and south suggest the latter region resulted from the rifting of oceanic lithosphere.

The basic similarity between the morphology of the magnetic quiet zones in the East Sheba Ridge where oceanic lithosphere has been rifted and within the Gulf of Aden where continental lithosphere was rifted suggests that it is not the thickness or the composition of the original crust that is the major factor in the overall development of the margin and the resulting structural elements as much as the thickness and thermal structure of the entire lithosphere. Thus the conditions and processes required for rifting of old oceanic lithosphere must be similar to those for the rifting of continental lithosphere. A comparison between the results obtained from oceanic and continental lithosphere thus should give additional information on and insight into the rifting mechanisms.

Much of the subsidence of rifted continental margins appears to be due to thermal processes (Sleep 1971; Watts & Ryan 1976). A number of models have been proposed to explain this observation which predict temperature distribution, heat flow and elevation with time. The simplest model (McKenzie 1978) explains the vertical motions as the result of horizontal extension of a lithosphere with an initial linear temperature gradient. This simple model has since been modified to include two-layered extension, with the top portion of the lithosphere stretched by a different factor than below and also extension by dyke intrusion (Royden, Sclater & von Herzen 1980; Royden & Keen 1980) resulting in a different temperature distribution.

The initial change in elevation from isostatic readjustment after lithospheric extension is due to the combination of the effects of crustal thinning, that increases the average density, and to lithospheric heating, that decreases the average density. Models of the thermal and mechanical development of continental margins typically have assumed a 30–35 km thick continental crust, which makes up approximately a quarter of the lithospheric thickness. Uniform extension of the continental lithosphere containing a pre-rift crust of this thickness always produces initial subsidence. However, if the crustal thickness is roughly half this amount, heating effects will be more significant than crustal thinning and the result is initial uplift. Thus one difference between the uniform extension of continental and oceanic lithosphere, according to these models, is that initial uplift will result for the latter, because the oceanic crustal thickness is typically 5–7 km. Regardless of the initial crustal and lithospheric thickness after rifting, heat flux and elevation will decrease with time and at equilibrium (assuming the crust is thinned), an extended area will be deeper than before rifting since the equilibrium depth depends only on the crustal thickness.

The procedure used in the thermal modelling follows that developed by Steckler (1981, 1985) and Cochran (1983). It explicitly includes the effects of lateral heat conduction across the horizontal temperature gradients set up by the rifting (Steckler 1981) and of an extended rather than instantaneous rifting event (Cochran 1983).

Most models assume instantaneous rifting; however, geological evidence suggests that rifting of the continents and initiation of seafloor spreading generally occurs over tens of millions of years. A finite length period of rifting is specifically included in our numerical calculations during which the extension parameter β increases linearly. The effects of finite periods of slow extension have been discussed by Jarvis & McKenzie (1980) and Cochran (1983). During and just after a period of extension lateral heat flow will significantly redistribute heat and change the temperature distribution, especially near the boundaries of the extended lithosphere (Steckler & Watts 1980, 1981).

Also, most models of rifted continental lithosphere have assumed an initial, linear temperature distribution with depth. This greatly simplifies the mathematical treatment. However, the method which we use permits an initial temperature gradient appropriate to the oceanic lithosphere of a given age to be assumed at the beginning of rifting. The values of constants used for these numerical calculations are given in Table 1.

Table 1. Values of constant model parameters.

Parameter	Value
Lithospheric thickness	125 km
Oceanic crustal thickness	5 km
Crustal density (0°C)	2.8 g cm^{-3}
Mantle density (0°C)	3.33 g cm^{-3}
Water density	1.03 g cm^{-3}
Coefficient of thermal expansion	3.4×10^{-5} °C^{-1}
Asthenospheric temperature	1333°C
Thermal conductivity	3.14 W m^{-1} °C^{-1}

Model constraints

BASEMENT DEPTHS

The most easily observable quantity which depends on the specific mechanism of rifting is the basement depth within the rifted region. At a well sedimented passive continental margins, the subsidence history can be determined using biostratigraphic data obtained from deep wells or by extrapolation of known stratigraphy to the study area using seismic reflection records. These sort of data are not available for the East Sheba Ridge quiet zone. However, the present depths can be used as a constraint providing one point on the subsidence curve 10 Myr after the end of the rifting event.

The basement depths used to constrain the modelling must be corrected for sediment loading. The sediment cover north of anomaly 5 on the East Sheba Ridge in the quiet zone and Owen Basin generally ranges from 0.5 to 2 s (Fig. 10). Sonobuoy data collected during V3617 and published data from Whitmarsh (1979) were used to determine sediment seismic velocities and estimated densities with depth in the sediment column (Table 2) for these calculations. An Airy type compensation was considered adequate because of the moderately small thickness and even distribution of the sediment.

The height and appearance of the Sharbithat Ridge complex varies substantially from profile to profile along its length. It consists of a fairly continuous ridge with a number of large peaks spaced along it and has the appearance of a volcanic ridge extruded on top of the crust, although it lacks the large magnetic anomaly often associated with rapidly constructed volcanic features. Thus, its relief would not be predicted from the simple thermal models. Therefore, no attempt will be made to match the specific shape of the ridge, but rather a smooth surface through its base will be assumed as a datum.
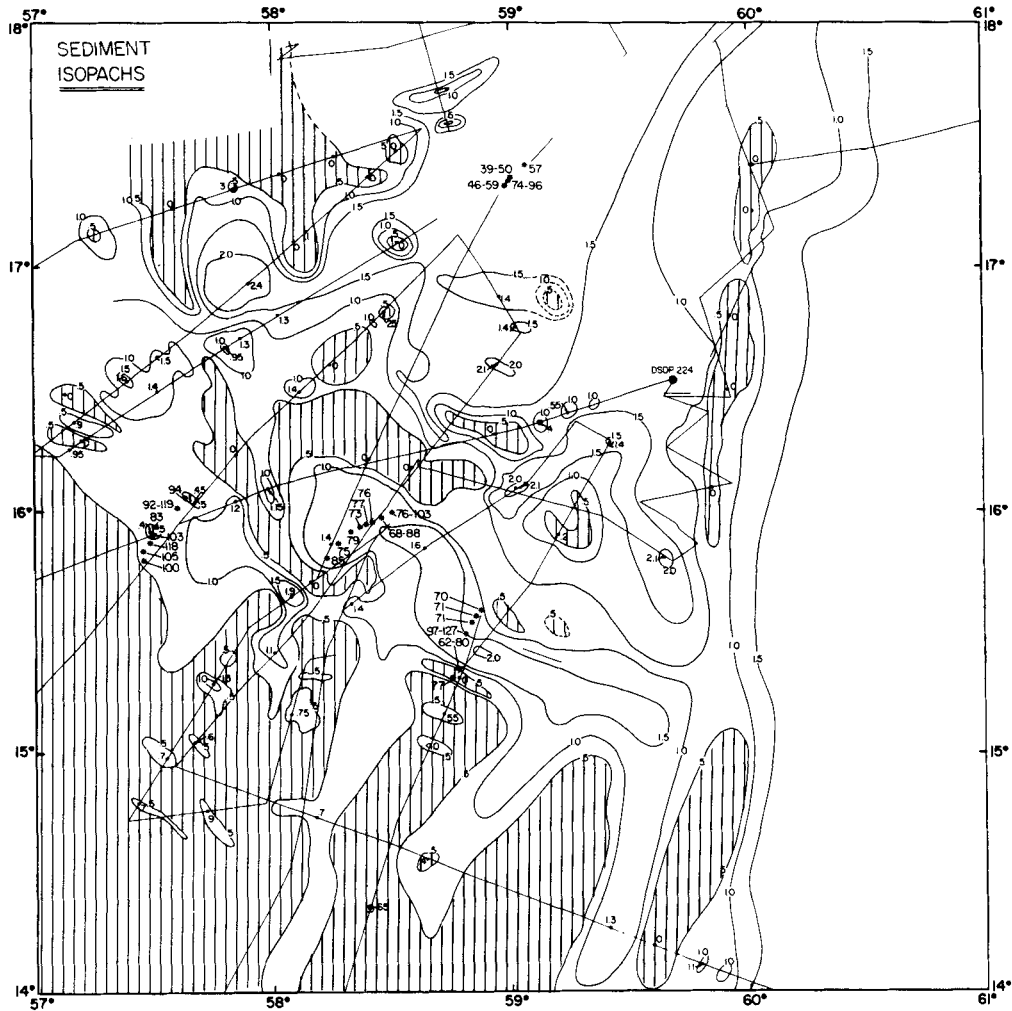
Figure 10. Sediment isopachs in seconds of two-way travel time from northern part of study area. Area covered by sediment less than 0.5 s thick is stippled. Light lines show location of seismic reflection and refraction data used to construct the map. Stars show locations of heat flow measurements. Numbers next to stars are heat flow values in mW m^{-2}.

Table 2. Sediment parameters used to calculate unloaded basement depths.

Depth (m)	Velocity (km s^{-1})	Density (g cm^{-3})
0–600	1.73	1.85
600–1750	2.6	2.2
>1750	3.2	2.3

HEAT FLOW

The second observable quantity which depends on the exact nature of the rifting mechanism and temperature is the surface heat flow. This is a particularly sensitive parameter for locations, such as the Sheba Ridge, where the rifting has occurred recently. Three heat flow

Table 3. Heat flow stations.

Station	Latitude	Longitude	Depth m	# Probes	Eval.	Env.	Gradient m°C/m	Heat Flow mW/m²
Western "Magnetic Quiet Zone"								
66 B	15°47.7'N	57°27.5'E	4111	5	10	A	105	100
C	49.7	27.4	4109	5	10	A	110	105
D	52.2	28.9	4111	5	10	B	124	118
E	54.3	29.7	4109	5	10	B	108	103
F	56.1	30.7	4107	5	10	B	87	83
G*	16 00.7	36.1	4087	5	9	B	97-125	92-119
H	03.6	37.8	4065	5	9	B	99	94
Owen Basin								
67 A	17°25.0'N	59°04.3'E	3658	5	9	A	56	57
E*	21.9	00.5	3665	2	7	A	38-49	39-50
F*	20.9	58°59.9	3665	2	7	A	73-94	74-96
G*	20.0	59.2	3667	2	7	A	45-58	46-59
Eastern "Magnetic Quiet Zone"								
68 A	15°48.3'N	58°13.7'E	4035	5	10	B	89	85
B	51.9	16.6	3939	5	10	A	79	75
C	54.5	19.8	3928	5	10	A	83	79
D	55.9	22.1	3939	5	10	A	77	73
E	56.4	23.7	3930	5	10	A	81	77
F	56.8	25.1	3939	5	10	A	80	76
G*	57.8	27.1	3964	5	9	A	72-93	68-88
H*	59.4	30.2	3995	5	7	A	80-104	76-103
69 A	15°18.5'N	58°45.5'E	4150	3	10	B	81	77
B*	20.4	46.6	4150	3	8	B	65-84	62-80
C*	30.7	49.5	4146	5	9	A	102-132	97-127
E	32.0	50.5	4144	4	10	A	75	71
F	33.8	51.3	4136	4	10	A	75	71
G	35.3	52.2	4127	4	10	A	74	70

Stations 66, 68, 69   K = 0.951 Watts / m°C
67   = 1.016 Watts / m°C

*Instrument tilt between 8° and 40°.

profiles in the northern magnetic quiet zone and one in the Owen Basin were obtained during *Vema* cruise 3617 in 1980 (Table 3). No previous measurements have been published for this region. The locations of the heat flow values are indicated on the sediment isopach map (Fig. 10). Temperature measurements were made with the Lamont–Doherty Geological Observatory digital heat flow instrument, employing five thermistors mounted on a 5.5 m spear with a sixth thermistor placed on the core head to measure bottom water temperatures. Temperatures, water pressure and instrument tilt are recorded every 30 s in digital form and stored on magnetic tape in the instrument as well as transmitted acoustically to the surface using a 12 kHz pinger. Instrument tilt is indicated if the angle is greater than 8° or 40°. Thermal conductivity values were determined from piston cores taken during cruise V3617 using the needle probe technique (Von Herzen & Maxwell 1959) and corrected for *in situ* conditions (Radcliffe 1960).

All thermal gradients were linear (Fig. 11) within the measurement error of the thermistors (approximately ± 0.005°C). The average standard deviation of the temperature gradients fits is approximately a few thousandths of a degree Celsius per metre, a few per cent of the measured gradient. The quality of each heat flow measurement is evaluated using a zero (bad) to 10 (excellent) scale (Langseth & Taylor 1967). Only those greater or equal to six are listed in Table 3. The local sedimentary environment near each measurement has been evaluated using the catagories suggested by Sclater, Crowe & Anderson (1976). For this cruise, stations were in only two (A and B) of the categories. Type A environments are on thick and relatively uniform sedimented regions with all basement relief covered and no outcrops within 10 km. Type B environments are similar to A but have outcrops within 10 km of the station. Most of the heat flow values in Table 3 are type A regions.

Heat flow values in the quiet zone tend to decrease slightly northward towards Sharbithat Ridge and also tend to decrease substantially towards the Owen fracture zone. The western-
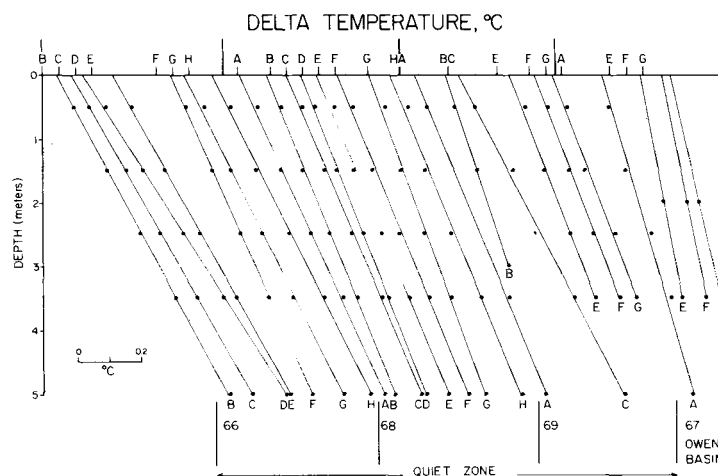
DELTA TEMPERATURE, °C

Figure 11. Temperature–depth profiles. Stations 66, 68 and 69 are in the magnetic quiet zone and station 67 is in the Owen Basin (see Fig. 10 and Table 3 for locations). In this figure, the depths of the individual temperature readings have not been corrected for the angle of penetration of the spear.

most measurements (station 66) were taken very close to profile V35-3 (Fig. 3) located in the western zone. Measured values range from 83 to 118 mW m^{-2}. There is an average of a half-second of sediment cover over very rough basement with some nearby outcrops. The two heat flow profiles (stations 68 and 69 in the eastern zone, near seismic lines V36-3 and V36-1 respectively, have significantly lower heat flow averaging about 75 mW m^{-2}. The easternmost station is located about 100 km from the Owen fracture zone over the narrow graben like feature with sediment thickness greater than 2 s which was described previously. Station 68, about 70 km further to the west, is on top of relatively smooth basement with somewhat more than 1 s of sediment cover. The average difference in heat flow between station 66 and the two stations to the east is about 20 mW m^{-2}. It might be expected that the thermal blanketing effect of the sediments may account for some of the difference in the average heat flow between stations 66 and 68–69. Assuming that all of the sediment has been deposited in the last 25 Myr, the correction, using the technique of Langseth, Hobert & Horai (1980), will be at most a 5 per cent increase in the heat flow values for the western stations and 15 per cent increase for the eastern stations. These corrections are insufficient to close the measured gap.

Station 67 is located just north of the end of seismic profile V36-3 (Fig. 7) in the Owen Basin. The measurements are 130 km north of the Sharbithat Ridge Complex axis and these measurements were taken to establish a heat flux for the Owen Basin to compare with the quiet zone. The basement is relatively smooth with at least about 1.5 s of sediment. The heat flow measurements (with the exception of 67F) are all much less than heat flow in the magnetic quiet zone, but higher than theoretically expected for old oceanic lithosphere (Parsons & Sclater 1977). However, the values are near world-wide heat flow of old regions of about 52 mW m^{-2} (Sclater, Jaupart & Galson 1980).

DURATION OF RIFTING

Seafloor spreading at the East Sheba Ridge began about 10 Myr BP. A number of types of data suggest that the rifting event began about 25 Myr BP. The uplift and westward tilting of

3

the high ridge of the Owen fracture zone began during the late Oligocene or earliest Miocene time (DSDP sites 223 and 224, Whitmarsh, Weser & Ross *et al.* 1974a, b). Large-scale faulting resulting in the uplift of the Arabian plateau began in Late Oligocene (Beydoun 1970), and in the earliest Miocene the seas transgressed westward from Ras Fartak (Beydoun 1982). Thus we will assume a 15 Myr period of extension (starting at 25 Myr BP near the Oligocene–Miocene boundary), followed by 10 Myr of cooling.

INITITAL THERMAL STRUCTURE

The age of the Owen Basin has not been determined. Whitmarsh (1979) reported the presence of ENE-trending magnetic anomalies near 19°N in the Owen Basin which have the appearance of seafloor spreading magnetic anomalies. It has not, however, been possible to identify a specific sequence. Two DSDP sites (223 and 224; see Fig. 2 for locations) drilled in shallow areas near the Owen fracture zone reached basement which in both locations consists of early Tertiary (57 and 51.5 Myr respectively) mafic volcanic rocks (Whitmarsh, Weser & Ross *et al.* 1974a, b). These rocks are similar in age to the early Tertiary seafloor to the east in the Arabian Sea and presumably reflect events related to spreading on the Carlsberg Ridge rather than the creation of the Owen Basin.

The Owen Basin is probably not related to the Oman Basin (Fig. 1), to the north. Structural trends in the Oman Basin, as indicated by the Samail ophiolites (Pallister 1981; Tilton, Hopson & Wright 1981), are different from those observed in the Owen Basin (Whitmarsh 1979). A more likely explanation is that the Owen Basin was formed by the breakup of Gondwanaland. Thus, it would be expected to be of a similar age to the southern Somali Basin and Mozambique Basin, where Late Jurassic and Early Cretaceous magnetic anomaly sequences have been identified (Simpson *et al.* 1979; Rabinowitz, Coffin & Falvey 1983).

Thus one would expect thick cold lithosphere to underly the Owen Basin. However, geophysical parameters from the Owen Basin away from the uplifted portions of the Owen fracture zone are not consistent with these conclusions and indicate a thinner lithosphere. Specifically the basement depths, about 4700 m after correction for the sediment loading, and the heat flow values (Table 3) are what would be expected from a region about 40–70 Myr old.

Thus two sets of conditions will be used as end members to represent the possible extremes of the initial temperature–depth relationship at the initiation of rifting 25 Myr BP. Model A is a thin lithosphere with the initial thermal structure of 15 Myr old oceanic lithosphere. Model B has a thicker lithosphere with the initial thermal structure of 125 Myr old oceanic lithosphere, consistent with a late Jurassic origin of the Owen Basin. Since the present data set only includes the present depths and heat flow and not the changes in these parameters through time we shall not attempt to distinguish between the two initial conditions.

Model calculations and results

Profile V35-3 (Fig. 7) was chosen to model because it is a typical profile across the region in which the various features of interest are present and well developed. The profile is also quite close to heat flow measurements (stations DHF 66), is far away from any possible complication effects of the Owen fracture zone, and is also not near East Sheba Ridge fracture zones.

It is assumed in the modelling that the amount of extension increases across the magnetic quiet zone south towards the East Sheba Ridge. The crustal stretching is the same for both
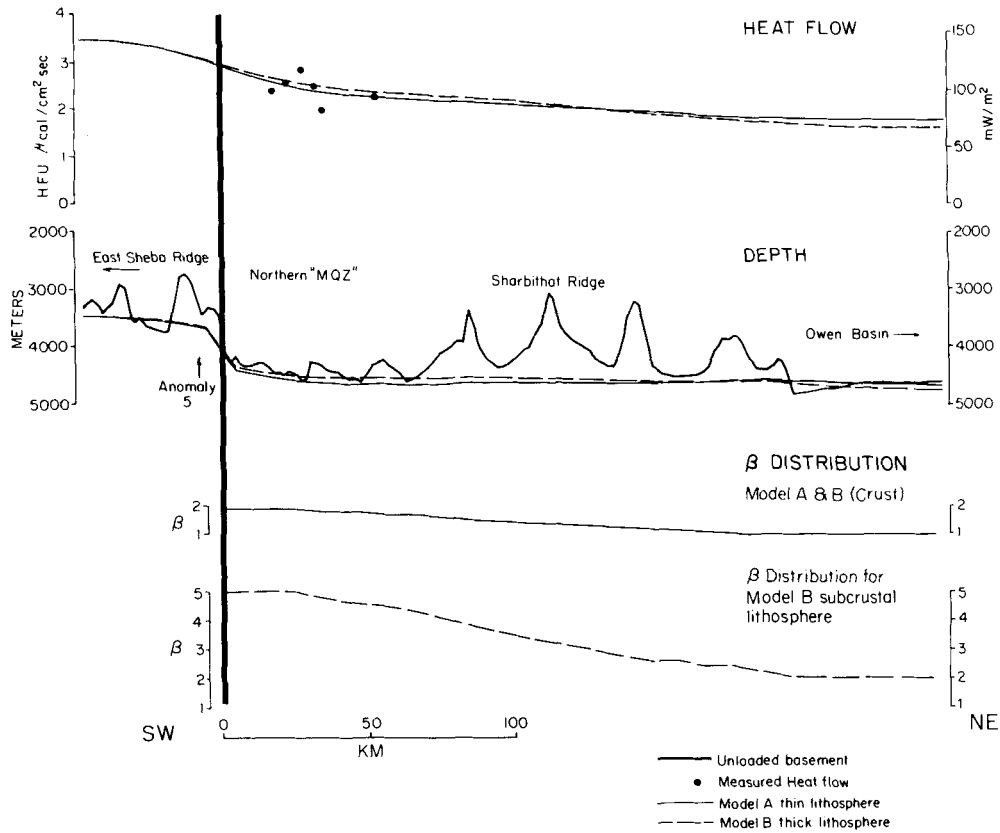
Figure 12. Comparison of observed and modelled depths and heat flow after 15 Myr of slow extension and 10 Myr of post-rift cooling. Unloaded basement depths were calculated by correcting observed basement for sediment loading assuming Airy type compensation. The model assumes that Sharbithat Ridge is a volcanic addition on top of the crust and matches a line through its base rather than the exact shape of ridge, which varies greatly from profile to profile. Both models A and B result in similar depths and heat flow. For model A, the thin lithosphere and crust are extended by same amount. For model B, with thick lithosphere, the crust is stretched by the same amount as for model A but the rest of the lithosphere is thinned much more in order to match the observed data.

end member model calculations, reaching a maximum of $\beta = 1.9$ at the quiet zone/East Sheba Ridge boundary. The original 5 km oceanic crust in the quiet zone is thus thinned in the model to a minimum of only 2.6 km. The quiet zone crustal thickness was not recorded from sonobuoy refraction experiments. For model A, the thin lithosphere column was stretched uniformly. For model B considerably more subcrustal thinning was necessary to match the geophysical parameters in the quiet zone. Extension of the lower lithosphere by a greater amount than the crust is used to simulate additional heating from below. This model also requires a 100 per cent thinning of the Owen Basin's subcrustal lithosphere to match present-day depths and heat flow. Both models result in similar calculated heat flow and depths which are close to the observed values (Fig. 12).

Fig. 13 shows calculations for both models when only the subcrustal portion of the lithosphere is thinned by an amount similar to the calculations in Fig. 12. The heat flow values predicted by the two models are similar to each other and to the observed values, as one would expect since the thinning of the crust does not significantly change the temperature distribution. However, the depths in the quiet zone are 100–400 m too high
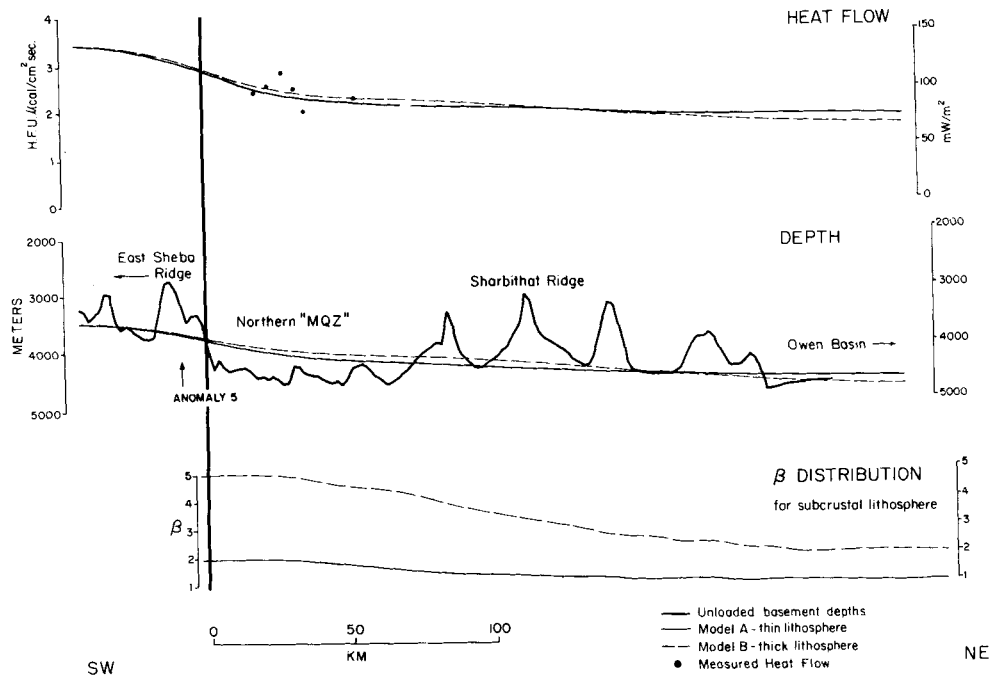
Figure 13. Comparison of observed and modelled depth and heat flow resulting from 15 Myr of slow extension and 10 Myr of post-rift cooling. Only the subcrustal portions of the lithosphere have been thinned, by amounts similar to subcrustal thinning shown in Fig. 12. Note that the sharp observed basement deepening is not reproduced and the predicted depths are too shallow in the quiet zone.

and also the sharp basement drop from the Sheba Ridge into the quiet zone is not reproduced. Thus crustal thinning (or equivalently an increase in the average crustal density) is a significant factor in matching the observed basement depths.

The amount of extension in the easternmost Gulf of Aden estimated from plate reconstructions can be used as a check on the amount of extension predicted from our modelling. The amount of total opening estimated in that manner (Cochran 1981, 1982b) prior to the initiation of seafloor spreading at the latitude of profile V35-3 is about 95 ± 35 km. The amount of extension for both models (Fig. 12) across the northern quiet zone is 56 km or 45 per cent of the original width. If the amount of extension in the southern quiet zone is similar this implies a total of 112 km of opening prior to 10 Myr BP. The two estimates are in reasonable agreement.

Lithospheric thickness and rifting

The calculated temperatures with depth at the southern end of the northern magnetic quiet zone just at the initiation of seafloor spreading are shown in Fig. 14. Since both of the 'end member' cases considered had to match the same elevations and heat flow following 10 Myr of cooling it is not surprising that their temperature structures at the time of initial seafloor spreading are similar. The temperature profile at the time of initiation of the seafloor spreading corresponds to the temperature profile for oceanic lithosphere between 10 and 14 Myr old. The old oceanic lithosphere of the Owen Basin was greatly thinned before seafloor spreading began. Similar amounts of lithospheric thinning have been deduced at other
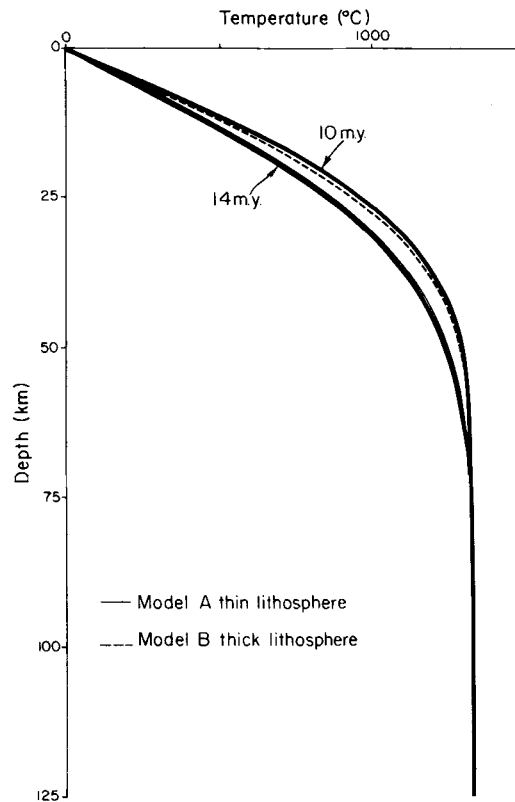
Figure 14. Comparison of temperature structure for models A and B (from Fig. 12) at East Sheba—northern magnetic quiet zone boundary after 15 Myr of slow extension compared with temperature structure for 10 and 14 Myr oceanic lithosphere.

continental margins where extensional models have been applied. Thus thin lithosphere is required before the generation of new oceanic lithosphere can commence.

New spreading centres are also formed on oceanic crust through the process of ridge crest jumps. Many such examples have been documented around the world, the most notable being the reorganization of the East Pacific Rise system (Sclater, Anderson & Bell 1971; Anderson & Sclater 1972; Cande, Herron & Hall 1982; Mammerickx & Klitgord 1982). Almost without exception these 'jumps' occur into lithosphere no more than 10 Ma (Table 4). The typical morphology associated with the boundary between the new crust and old at a ridge jump is much sharper than for Sheba Ridge and resembles that normally associated with a fracture zone (Anderson & Sclater 1972; Cande *et al.* 1982). Also the time required for the new spreading centre to develop appears to be very short compared with the 15 Myr required at Sheba Ridge or times approaching 50 Myr which have been reported at some continental margins (Jansa & Wade 1975; Talwani *et al.* 1979). Both observations reflect the already thin lithosphere near existing ridge crests into which 'ridge crest jumps' normally occur.

Both of the situations mentioned above, the rifting of old lithosphere to form a magnetic quiet zone and ridge crest jumps, share a common denominator, the need for a thin lithosphere. The features found in our study are (Fig. 1) are a direct result of the necessity substantially to thin the initially thick lithosphere before seafloor spreading can begin at

Table 4. Ridge crest jumps.

Abandoned Spreading Center	Estimated Date Abandoned (M.Y.B.P.)	Estimated age of oceanic lithosphere rifted (m.y.) during formation of new spreading center	Reference
Selkirk Rise	23	< 3	a
Roggeveen Rise	20	< 4	a
Gallego Rise	19	< 7	a
Mendoza Rise	18.5	< 7	a
Galapagos Rise	6.5	5-10*	a
Mathematician Seamounts (south)	11? to	0-few millions years	b
(north)	2.5	< 10	b
Antarctic-Pacific spreading center northward propagation between Tula & Humboldt fracture zones	52	10	c
Galapagos Spreading Center near 95°W (ridge crest jumps and propagating rifts)	< 3	< 2	d

References

[a] Mammerickx *et al.* (1975) and Mammerickx, Herron & Dorman (1980).
[b] Mammerickx & Klitgord (1982) and Klitgord & Mammerickx (1982).
[c] Cande *et al.* (1982).
[d] Hey, Duennebier & Morgan (1980).
*Spreading began on the East Pacific Rise 8.2 Myr BP, concurrent to slow spreading at the Galapagos Rise.

the East Sheba Ridge. The thickness of the lithosphere at the initiation of seafloor spreading is similar to the maximum thickness into which ridge crest jumps have occurred. This correspondence suggests that there is a maximum lithospheric thickness into which seafloor spreading can occur. This maximum thickness (about 40—45 km; Fig. 14) corresponds to thermal structure of about 10 Ma lithosphere.

Structural changes near the Owen fracture zone

The models which we have discussed in the previous sections strictly apply only to the western zone. As the Owen fracture zone is approached, the structures observed on the Sheba Ridge and in the quiet zone change. The ridge flank depths increase and the ridge loses its characteristic 'cooling curve' shape. Instead it takes the form of a region of rough bathymetry with a relatively constant mean depth of about 3500 m (Fig. 7, profile V36-1). The change in the morphology of the ridge is accompanied by a decrease in the amplitude of the magnetic anomalies beginning at about 130 km from the Owen fracture zone. Also, as the Owen fracture zone is approached, the sharp topographic break between the ridge flank and the quiet zone disappears with the result that the ridge and trough morphology extends further north and the distinction between ridge flank and quiet zone becomes less clear. ·

The Australian—Antarctic Discordance on the Southeast Indian Ridge (120°E–128°E) is another section of the mid-ocean ridge characterized by greater than normal depth, loss of the characteristic mid-ocean ridge shape and poorly developed magnetic anomalies. These observations have been interpreted as indicating that the Discordance resulted from downward asthenospheric flow with low crystallization temperatures for the oceanic crust (Weissel & Hayes 1977; Anderson *et al.* 1980). Although there is not an exact analogy, in part because the Australian—Antarctic Discordance covers a much larger area, the obser-

vations on the East Sheba Ridge also suggest the presence of lowered upper mantle temperatures near the Owen fracture zone. This conclusion is supported by the decrease in heat flow values in the magnetic quiet zone toward the Owen fracture zone.

One mechanism which might be expected to produce lower upper mantle temperatures is lateral conductive heat flow across the Owen fracture zone. The lithosphere to the east in the Arabia Sea is about 50 Myr older than the adjacent Sheba Ridge lithosphere and thus there will be a significant temperature contrast across the Owen fracture zone. However, simple calculations show that this effect will only significantly decrease heat flow and increase depths within about 50 km west of the fracture zone, not enough to account for the values at stations 69 and 68 which are 100 and 170 km respectively to the west of the Owen fracture zone. Although the specific mechanism is not clear, it appears that the normal convective flow to the mid-ocean ridge is not as well developed at the eastern end of Sheba Ridge as further to the west resulting in the lower than normal asthenospheric and lithospheric temperatures.

Conclusion

Magnetic quiet zones are located throughout the Gulf of Aden to the north and south of the Sheba Ridge from Afar to the Owen fracture zone. This includes the easternmost region where oceanic lithosphere of the Owen and Somali Basins was rifted to form the East Sheba Ridge. In this area the boundary between the quiet zones and the old oceanic lithosphere is marked by ridge complexes, the Sharbithat and Error Ridges (to the north and south respectively) which have no significant magnetic anomaly signature and are associated with large negative gravity anomalies that cannot be explained by basement features. The quiet zones are characterized by low and uncorrelatable magnetic anomalies and deeper basement (by about 1 km) than the adjacent 10 Ma East Sheba Ridge crust. In the western zone the boundary between the East Sheba Ridge and northern quiet zone is marked by a sharp increase to the north in the basement depth.

The well-stratified sediments of the Owen Basin abut against the northern flank of the Sharbithat Ridge with no signs of disturbance. This suggests either that the Sharbithat Ridge was formed early in the rifting history prior to the deposition of most of the sediments or that it pre-dates the opening. On some profiles across the magnetic quiet zone and the southern flank of Sharbithat Ridge the lower half of the sediment cover appears to be disturbed and in some cases is upturned on to the ridge flank. This indicates possible tectonic activity before the deposition of the relatively flat-lying upper layers. Thus the major phase of tectonic activity responsible for the development of the magnetic quiet zone appears to have occurred during the early stages of development of Sheba Ridge.

The magnetic quiet zone regions appears to have been produced by the diffuse extension of old oceanic lithosphere. They formed during the opening of the Gulf of Aden and the formation of the magnetic quiet zones along the continental margins during a rifting event from 25 to 10 Myr BP. Using a variation of stretching models developed by McKenzie (1978), Royden, Sclater & von Herzen (1980), and Steckler (1981), which permits finite length rifting events, it is possible to match observed depths and heat flow values. The model results imply that during the rifting event the crust of the magnetic quiet zone was extended about 45 per cent. During extension the lithosphere was thinned to a thermal thickness similar to that of 10–14 Ma oceanic lithosphere. Examination of ridge crest jumps suggest the maximum age of lithosphere into which these jumps occur is about 10 Myr. These observations, as well as the large amounts of thinning deduced at rifted continental margins, indicate that lithosphere must be thinned to a thickness similar to that of 10–14 Myr or younger oceanic lithosphere before seafloor spreading is initiated.

Acknowledgments

Reference

Anderson, R. N. & Sclater, J. G., 1972. Topography and evolution of the East Pacific rise between 5°S and 20°S, *Earth planet. Sci. Lett.*, 14, 433–411.

Anderson, R. N., Spariosu, D. J., Weissel, J. K. & Hayes, D. E., 1980. The interrelation between variations in magnetic anomaly amplitudes and basalt magnetization and chemistry along the Southeast Indian Ridge, *J. geophys. Res.*, 85, 3883–3898.

Azzaroli, A. & Fas, V., 1964. Geological outlines of the northern end of the Horn of Africa, *22nd int. Geol. Congr. Rep.*, Part IV, 293–314.

Beydoun, Z. R., 1970. Southern Arabia and northern Somalia: comparative geology, *Phil. Trans. R. Soc. A*, 267, 267–292.

Beydoun, Z. R., 1982. The Gulf of Aden and Northwest Arabian Sea, in *The Ocean Basins and Margins: the Indian Ocean*, eds Nairn, A. E. M. & Stehli, F. G., Plenum Press, New York.

Bunce, E. T., Langseth, M. G., Chase, R. L., Ewing, M. 1967. Structure of the Western Somali Basin, *J. geophys. Res.*, 72, 2547–2555.

Cande, S. C., Herron, E. M. & Hall, B. R., 1982. The Early Cenozoic tectonic history of the Southwest Pacific, *Earth planet. Sci. Lett.*, 57, 63–74.

Cochran, J. R., 1981. The Gulf of Aden: structure and evolution of a young ocean basin and continental margin, *J. geophys. Res.*, 86, 263–288.

Cochran, J. R., 1982a. Reply. *J. geophys. Res.*, 87, 6765–6770.

Cochran, J. R., 1982b. The magnetic quiet zone in the eastern Gulf of Aden: Implications for the early development of the continental margin, *Geophys. J. R. astr. Soc.*, 68, 171–201.

Cochran, J. R., 1983. Effects of finite rifting times on the development of sedimentary basins, *Earth planet. Sci. Lett.*, 66, 289–302.

Coleman, R. G., Fleck, R. J., Hodge, C. E. & Ghent, E. D., 1975. The volcanic rocks of southwest Saudi Arabia and the opening of the Red Sea, in *Red Sea Research 1970–1975, Bull. Miner. Resour.*, 22, D1–D30, Directorate General of Mineral Resources, Jiddah. Saudi Arabia.

Coleman, R. G., Hadley, D. G., Fleck, R. G., Hedge, C. T. & Donato, M. M., 1979. The Miocene Tihama Asir Ophiolite and its bearing on the opening of the Red Sea, in *Evolution and Mineralization of the Arabian—Nubian Shield*, ed. Al-Shanti, A. M. S., *Bull. Inst. Appl. Geol.*, 3, 173–186, Institute of Applied Geology, King Abdulaziz University, Jiddah.

de Charpal, O., Guennoc, P., Montadert, L. & Roberts, D. G., 1978. Rifting crustal attenuation and subsidence in the Bay of Biscay, *Nature*, 275, 706–711.

General Bathymetric Chart of the Oceans (GEBCO), 5th edn, 1975. Canadian Hydrographic Service, Ottawa, Canada.

Girdler, R. W., Brown, C., Noy, D. J. M. & Styles, P., 1980. A geophysical survey of the westernmost Gulf of Aden, *Phil. Trans. R. Soc. A*, 298, 1–43.

Girdler, R. W. & Styles, P., 1978. Seafloor spreading in the western Gulf of Aden, *Nature*, 271, 615–617.

Girdler, R. W. & Styles, P., 1982. Comments on "The Gulf of Aden: Structure and evolution of a young ocean basin and continental margin" by J. R. Cochran, *J. geophys. Res.*, 87, 6761–6763.

Hey, R., Duennebier, F. K. & Morgan, W. J., 1980. Propagating rifts on mid-ocean ridges, *J. geophys. Res.*, 85, 3647–3658.

Jansa, L. F. & Wade, J. A., 1975. Geology of the continental margin off Nova Scotia and Newfoundland, in *Offshore Geology of Eastern Canada, Pap. geol. Surv. 74–30,* 2, 51–105.

Jarvis, G. T. & McKenzie, D. P., 1980. Sedimentary basin formation with finite extension rates, *Earth planet. Sci. Lett.,* 48, 42–52.

Karner, G. D. & Watts, A. B., 1982. On isostasy at Atlantic-type continental margins, *J. geophys. Res.,* 87, 2923–2948.

Keen, C. E. & Barrett, D. C., 1981. Thinned and subsided crust on the rifted margin of eastern Canada: crustal structure, thermal evolution and subsidence history, *Geophys. J. R. astr. Soc.,* 65, 443–465.

Klitgord, K. D. & Mammerickx, J., 1982. Northern East Pacific rise: magnetic anomaly and bathymetry framework, *J. geophys. Res.,* 87, 6725–6750.

LaBrecque, J. L., Kent, D. V. & Cande, S. C., 1977. Revised magnetic polarity time scale for Late Cretaceous and Cenozoic Time, *Geology,* 5, 330–335.

Langseth, M. G., Hobart, M. A. & Horai, K., 1980. Heat flow in the Bering Sea, *J. geophys. Res.,* 85, 3740–3750.

Langseth, M. G. & Taylor, P. I., 1967. Recent heat flow measurements in the Indian Ocean, *J. geophys. Res.,* 72, 6249–6260.

Laughton, A. S., 1966. The Gulf of Aden, *Phil. Trans. R. Soc. A,* 259, 150–171.

Laughton, A. S., Whitmarsh, R. B. & Jones, M. T., 1970. The evolution of the Gulf of Aden, *Phil. Trans. R. Soc. A,* 267, 227–266.

Mammerickx, J., Anderson, R. N., Menard, H. W. & Smith, S. M., 1975. Morphology and tectonic evolution of the east-central Pacific, *Bull. geol. Soc. Am.,* 86, 111–118.

Mammerickx, J., Herron, E. & Dorman, L., 1980. Evidence for two fossil spreading ridges in the southeast Pacific, *Bull. geol. Soc. Am.,* 91, 263–271.

Mammerickx, J. & Klitgord, K. D., 1982. Northern East Pacific Rise: evolution from 25 m.y.B.P. to the present, *J. geophys. Res.,* 87, 6751–6759.

Matthews, D. H., Williams, C. A. & Laughton, A. S., 1967. Mid-ocean ridge in the mouth of the Gulf of Aden, *Nature,* 215, 1052–1053.

McKenzie, D. P., 1978. Some remarks on the development of sedimentary basins, *Earth planet. Sci. Lett.,* 40, 25–32.

McKenzie, D. P. & Sclater, J. G., 1971. The evolution of the Indian Ocean since the late Cretaceous, *Geophys. J. R. astr. Soc.,* 25, 437–528.

Montadert, L., de Charpal, O., Roberts, D., Guennoc, P. & Sibuet, J. C., 1979. Northeast Atlantic passive continental margins: rifting and subsidence processes, in *Deep Drilling Results in the Atlantic Ocean: Continental Margins and Paleoenvironments,* eds Talwani, M., Hey, W. & Ryan, W. B. F., *Am. geophys. Un., M. Ewing Series,* 3, 154–186.

Mutter, J. C., Talwani, M. & Stoffa, P. L., 1982. Origin of seaward-dipping reflectors in oceanic crust off the Norwegian margin by "subaerial seafloor spreading", *Geology,* 10, 353–357.

Pallister, J. S., 1981. Structure of the sheeted dike complex of the Somali ophiolite near Ilbra, Oman, *J. geophys. Res.,* 86, 2661–2672.

Parsons, B. & Sclater, J. G., 1977. An analysis of ocean floor bathymetry and heat flow with age, *J. geophys. Res.,* 82, 803–827.

Rabinowitz, P. D., Coffin, M. F. & Falvey, D., 1983. The separation of Madagascar and Africa, *Science,* 220, 67–69.

Radcliffe, E. H., 1960. The thermal conductivities of ocean sediments, *J. geophys. Res.,* 65, 1535–1541.

Royden, L. & Keen, C. E., 1980. Rifting processes and thermal evolution of the continental margin of eastern Canada determined from subsidence curves, *Earth planet. Sci. Lett.,* 51, 343–361.

Royden, L., Sclater, J. G. & von Herzen, R. P., 1980. Continental margin subsidence and heat flow: important parameters in formation of petroleum hydrocarbons, *Bull. Am. Ass. Petrol. Geol.,* 64, 173–187.

Sclater, J. G., Anderson, R. N. & Bell, M. L., 1971. Elevation of ridges and evolution of the central eastern Pacific, *J. geophys. Res.,* 76, 7888–7915.

Sclater, J. G., Crowe, J. & Anderson, R. N., 1976. On the reliability of oceanic heat averages, *J. geophys. Res.,* 81, 2997–3006.

Sclater, J. G., Jaupart, C. & Galson, D., 1980. The heat flow through oceanic and continental crust and the heat loss of the Earth, *Rev. Geophys. Space Phys.,* 18, 269–311.

Simpson, E. S. W., Sclater, J. G., Parsons, B., Norton, I. & Meinke, L., 1979. Mesozoic magnetic lineations in the Mozambique Basin, *Earth planet. Sci. Lett.,* 43, 260–264.

Sleep, N. S., 1971. Thermal effects of the formation of Atlantic continental margins by continental break-up, *Geophys. J. R. astr. Soc.,* 24, 325–350.

Somaliland Oil Exploration Co., Ltd, 1954. *A geological reconnaissance of the sedimentary deposits of the Protectorate of British Somaliland*, Crown Agents for the Colonies, London.

Steckler, M. S., 1981. The thermal and mechanical evolution of Atlantic-type continental margins, *PhD thesis*, Columbia University, New York.

Steckler, M. S., 1985. Two-dimensional considerations in the subsidence of continental margins, *J. geophys. Res.*, in press.

Steckler, M. S. & Watts, A. B., 1980. The Gulf of Lion: subsidence of a young continental margin, *Nature*, 287, 425–429.

Steckler, M. S. & Watts, A. B., 1981. Subsidence history and tectonic evolution of Atlantic-type continental margins, in *Dynamics of Passive Margins, Geodynamics Series*, 6, 184–196, *American Geophysical Union*, Washington, DC.

Styles, P. & Hall, S. A., 1980. A comparison of the seafloor spreading histories of the western Gulf of Aden and the Central Red Sea, in *Geodynamic Evolution of the Afro-Arabian Rift System*, pp. 587–606, Accademia Nazunali Dei Lincei, Rome.

Talwani, M. & Eldholm, O., 1973. The boundary between continental and oceanic basement at the margin of rifted continents, *Nature*, 241, 325–330.

Talwani, M., Mutter, J., Houtz, R. & König, M., 1979. The crustal structure and evolution of the area underlying the magnetic quiet zone on the margin south of Australia, in *Geological and Geophysical Investigations of Continental Margins*, eds Watkins, J. S., Montadert, L. & Dickerson, P. W., pp. 151–175, American Association for Petroleum Geology, Tulsa, Oklahoma.

Talwani, M. & Udintsev, G., 1976. Tectonic synthesis, in *Init. Rep. Deep Sea Drill. Proj.*, 38, 1213–1242, eds Talwani, M., Udintsov, G. *et al.*, US Government Printing Office, Washington, DC.

Tilton, G. R., Hopson, C. A. & Wright, J. E., 1981. Uranium-lead isotopic ages of the Somali Ophiolite, Oman, with applications to Tethyan Ocean Ridge tectonics, *J. geophys. Res.*, 86, 2763–2776.

Von Herzen, R. & Maxwell, A. E., 1959. The measurement of thermal conductivity of deep-sea sediments by a needle-probe method, *J. geophys. Res.*, 64, 1557–1563.

Watts, A. B. & Ryan, W. B. F., 1976. Flexure of the lithosphere and continental margin basins, *Tectonophys.*, 36, 24–44.

Weissel, J. K. & Hayes, D. E., 1977. The Australian Antarctic Discordance, new results and implications, *J. geophys. Res.*, 79, 2579–2587.

Whitmarsh, R. B., 1979. The Owen Basin off the south-east margin of Arabia and the evolution of the Owen Fracture Zone, *Geophys. J. R. astr. Soc.*, 58, 441–470.

Whitmarsh, Weser & Ross *et al.*, 1974a. Site 223, in *Init. Results Deep Sea drill. Proj.*, 23, 291–317, eds Supko, R. R. & Weser, O. E., US Government Printing Office, Washington, DC.

Whitmarsh, Weser & Ross *et al.*, 1974b. Site 224, in *Init. Results Deep Sea drill. Proj.*, 23, 383–395, eds Supko, R. R. & Weser, O. E., US Government Printing Office, Washington, DC.