Reprinted from Continental and Oceanic Rifts
Geodynamics Series, Volume 8

GRAVITY ANOMALIES OVER OCEANIC RIFTS

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The earliest measurements of the gravity field over mid-ocean ridges were obtained by Vening Meinesz (1948) using a pendulum apparatus. Continuous profiling using surface-ship gravimeters (Talwani et al., 1961; 1965) showed that free-air gravity anomalies over ridges were generally small and that ridge crests were approximately in isostatic equilibrium. Talwani et al. (1965) by considering ridge crests as a mass excess, suggested they were compensated by an underlying mass deficiency. The form of the compensation could not be deduced from gravity data alone, but was generally considered to extend to depths greater than 30 km below the ridge axis.

Mid-ocean ridges are now recognized as the site of creation of oceanic lithosphere which subsequently evolves by cooling and solidification of hot mantle material. Simple cooling models have been constructed (for example, Sclater and Francheteau, 1970) which generally explain the thickness of the plates, the shape of their upper surface, heat flow and other geophysical parameters such as the dispersion of surface waves. In these models the excess elevation of ridges is assumed to be compensated at the base of the lithosphere.

Lambeck (1972) showed that the simple cooling models are associated with a small amplitude free-air gravity anomaly high over the ridge crest and flanking lows. The amplitude and wavelength of the gravity anomaly depends on the thermal structure of the ridge crest and hence the half-spreading rate. Cochran and Talwani (1977) showed that a small amplitude gravity anomaly high is observed over most mid-ocean ridges, although they could not distinguish a strong dependence on the spreading rate.

These previous studies were mainly concerned with the manner mid-ocean ridges were supported and thus were limited to the interpretation of the relatively long-wavelength gravity field (wavelength $\lambda \gtrsim 800$ km). A number of prominent features also occur in the short-wavelength gravity field ($\lambda \lesssim 250$ km) over ridge crests (Talwani et al., 1965; Van Andel and Bowin, 1968; Woodside, 1972; Cochran, 1979; Collette et al., 1980). The most striking of these are large amplitude free-air gravity anomaly lows (up to -75 mgal) over oceanic rifts.

The purpose of this paper is to review the current knowledge of gravity anomalies over oceanic rifts. A brief description is given of gravity anomalies over active and extinct oceanic rifts in the Indian, Pacific and Atlantic oceans and some of the significant results of the interpretation of the gravity field over oceanic rifts summarized.

Gravity Anomalies

Free-air gravity anomaly and bathymetry profiles of active oceanic rifts in the Pacific, Indian, and Atlantic oceans are summarized in Figs. 1 and 2. The sources and locations of each profile are given in Table 1.

Figure 1 shows that active oceanic rifts are associated with large amplitude (up to -70 mgal) short-wavelength ($\lambda \sim 40$ to 60 km) free-air gravity anomaly lows which are flanked by gravity anomaly highs. The oceanic rift of the relatively slow-spreading Mid-Atlantic and Sheba ridges (half-spreading rate 1.2 to 1.5 cm/yr) correlate with a gravity low of amplitude 50 to 70 mgal and wavelength about 80 km. The oceanic rift of the faster spreading Gorda ridge (2.9 cm/yr) correlates with a gravity low of amplitude 30 mgal and wavelength about 40 km. These differences in the shape of the gravity anomaly cannot be simply explained by differences in topography. The rift valley on profiles 1 and 3 (Fig. 1), for example, has a similar width and depth but is associated with different shape gravity anomalies.

Fig. 2 shows that rift valley 2 in a segment of the Mid-Atlantic ridge investigated as part of project FAMOUS (ARCYANA, 1975) is associated with a gravity anomaly low of amplitude 50 to 60 mgal and wavelength of about 60 km. The gravity anomaly minimum occurs over the rift valley terrace between each outer wall. The minimum does not, therefore, occur over the inner floor, which Needham and Francheteau (1974) consider to be the locus for crust of zero age.

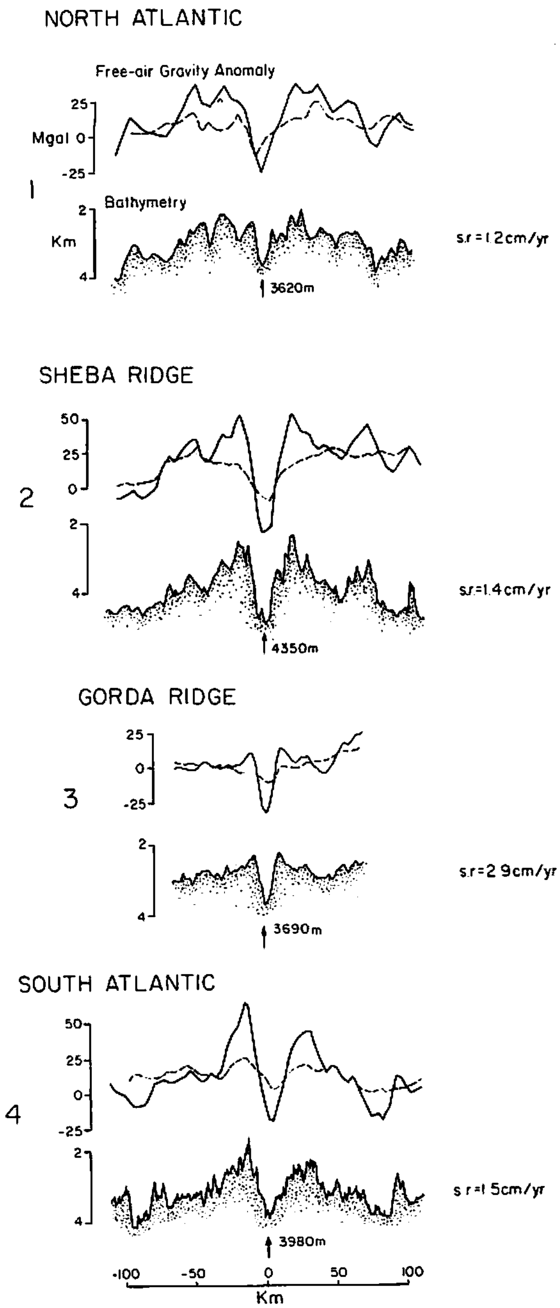

Figure 1. Selected free-air gravity anomaly, isostatic anomaly (based on Airy model of isostasy with T = 30 km), and bathymetry profiles of active oceanic rifts in the Atlantic (Profiles 1 and 4), Indian (Profile 2) and Pacific (Profile 3) oceans. Each profile has been projected normal to the local trend of the ridge crest. The half-spreading rates (s.r.), based on magnetic anomaly lineations over the ridge crest, are shown to the right of each bathymetry. The location of the profiles and sources of data are given in Table 1.

Isostasy

There is generally a strong correlation between short-wavelength free-air gravity anomalies and bathymetry over oceanic rifts and ridge flanks (for example, Figs. 1 and 2). This correlation provides information on the nature of the isostatic mechanism at ridge crests and the origin of oceanic rifts. In addition, since ridge flank topography is probably created at the ridge axis and then transported away from the ridge by sea-floor spreading (Needham and Francheteau, 1974), the isostatic mechanism at ridge flanks provides constraints on the physical processes occurring at the ridge axis.

Fig. 1 and 2b show isostatic gravity anomalies based on the Airy model (assumed depth of compensation T_c = 30 km) for each ridge crest profile. Although the isostatic anomalies over ridge crests are generally smaller than the free-air gravity anomalies, there are large departures from Airy isostasy particularly over the rifts. The oceanic rifts on each of the profiles are undercompensated, according to the Airy model.

The problem with using the Airy model at mid-ocean ridge crests, however, is that it predicts large variations in crustal thickness between the rim mountains and oceanic rifts. Seismic studies in the Atlantic (Keen and Tramontini, 1970) and Pacific oceans (Goslin et al., 1972; Stoffa et al., 1980) show that oceanic crust is of normal thickness at least to within about 10 km from the ridge axis. Thus the rim mountains do not appear to be supported by thickened crust, as predicted by the Airy model.

Cochran (1979) considered both the Airy model and plate or flexure model of isostasy at ridge crests. Isostatic anomalies based on the flexure model (assumed effective elastic thickness T_e = 7 to 13 km) showed only small departures from isostasy at oceanic rifts. However, this model, like the Airy model, also predicts significant crustal thickening beneath the rim mountains.

Although the nature of the isostatic mechanism at oceanic rifts is therefore not well understood, some insight into the physical processes at the ridge axis have come from studies of the isostatic mechanism of ridge flank topography. Talwani et al. (1972) showed that for wavelengths less than about 150 km the correlation between free-air gravity anomalies and topography could be explained by a simple model in which the topography of the ridge flank was uncompensated. McKenzie and Bowin (1976) showed, using linear transfer function techniques, that the correlation between gravity anomalies and topography could be best explained if ridge flank topography was compensated by the flexure model with T_e = 10 km. The analysis of McKenzie and Bowin (1976), included ridge flank topography as well as features which formed at a ridge crest, such as fracture zone topography and the aseismic Walvis ridge.

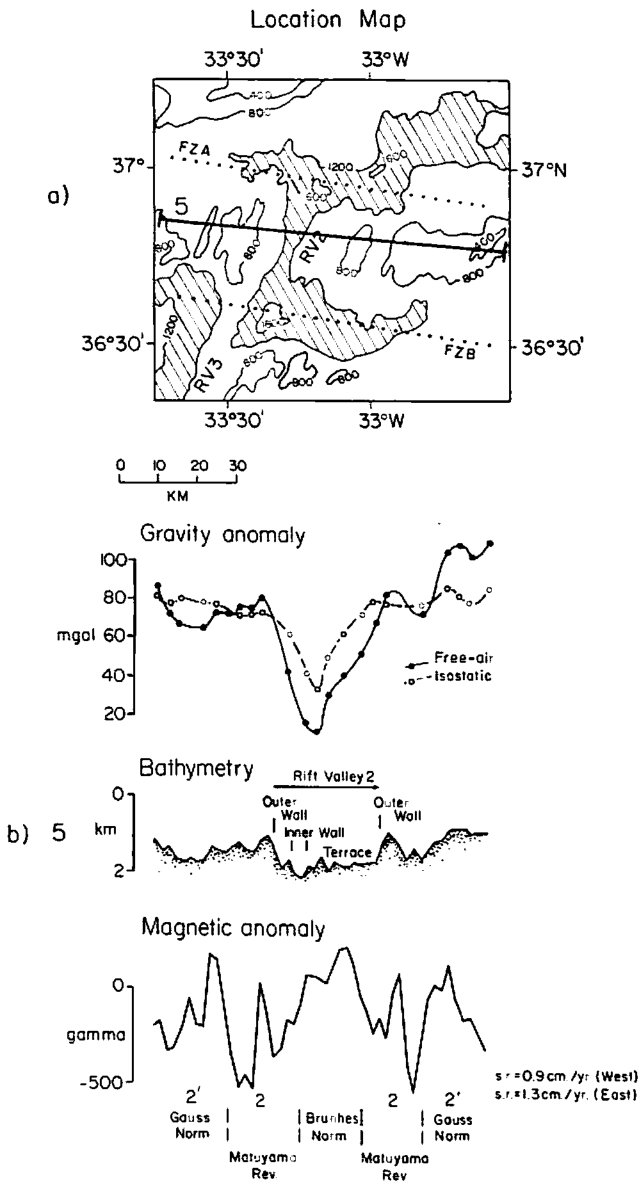

Figure 2. (a). Summary bathymetry map of the FAMOUS area (ARCYANA, 1975) of Mid-Atlantic ridge crest at 37° North (Ramberg et al., 1977). The contour interval is 400 fthm (732 meters). Heavy dots indicate fracture zones A and B. The continuous line shows the location of Profile 5 in Fig. 2b. RV2 = rift valley 2 and RV3 = rift valley 3 (Ramberg et al., 1977).
(b). Gravity anomaly, bathymetry and magnetic anomaly profile of rift valley 2 in the FAMOUS area of the Mid-Atlantic ridge crest at 37° North. The physiographic provinces on the bathymetry profile are based on terminology proposed by Needham and Francheteau (1974). The magnetic anomaly identifications and approximate boundaries between magnetic epochs are shown below the magnetic anomaly profile. Note the minimum in the free-air gravity anomaly and isostatic anomaly occurs not over the inner floor, which is considered the locus of crust of zero age, but over the terrace approximately equi-distant between the Outer walls. The half-spreading rates shown are based on Needham and Francheteau (1974) and assume zero age crust at the inner floor and their identification of the Brunhes-Matuyama boundary.

Recently, McNutt (1979) used similar transfer function techniques and analyzed free-air gravity anomaly and bathymetry data from the SURVEYOR survey of the Juan de Fuca and Gorda ridge crests. This study, which was based on gravity anomaly and bathymetry maps rather than individual profiles, showed that the gravity anomalies could be best explained if ridge flank topography was compensated by the flexure model with T_e = 2 to 6 km (Fig. 3).

Origin of the Rift Valley

Studies of the gravity field over ridge crests therefore provide substantial evidence for the existance of a continuous layer of strength at the ridge axis. The effective elastic thickness T_e of the ridge axis, as inferred from ridge flank gravity anomaly and bathymetry data, is in the range 2 to 13 km (McKenzie and Bowin, 1976; Cochran, 1979; McNutt, 1979). These estimates for the effective elastic thickness, which are significantly less than values associated with seamount loads in the plate interiors (20 to 30 km; Watts, 1978), can be generally interpreted as due to the origin of ridge flank topography in the axial region of the ridge crest, where the lithosphere is hot and weak.

The presence of a continuous layer of strength at the ridge axis has also been inferred from previous studies based on thermal and petrologic models (Turcotte, 1974; Tapponnier and Francheteau, 1978). Turcotte (1974), based on evidence for the depth to the 700°C isotherm beneath southwest Iceland, argued for a strong mechanical layer at the ridge axis. Tapponnier and Francheteau (1978) based on assumptions of the depth to the solidus temperature beneath the ridge axis, suggested the thickness of the mechanical layer was about 4 km.

Lonsdale and Speiss (1979) and Johnson and Vogt (1973) have mapped large topographic features near or at the ridge axis which require a strong mechanical layer to support them. Two small seamounts, which rise about 1.4 km above the ridge flank, occur either side of the East Pacific rise crest at lat. 8° 46'N on crust 0.6 m.y. old. Large volcanoes are generally absent in the axial region of the rise crest. High central volcanoes (up to a few hundred meters relief) sometimes occur, however, in the axial region of the Mid-Atlantic ridge (Johnson and

TABLE 1. Summary of Gravity Anomaly and Bathymetry Data

Profile Number	Ridge Crest	Ship	Cruise	Position of Ridge Crest Crossing		Azimuth	Reference for Location of Profile
1	North Atlantic	Snellius	SN07	28.0°N	44.0°W	90°	Cochran (1979), Fig. 6
2	Sheba	Vema	V3503	14.5°N	56.3°E	32°	Cochran (in press)
3	Gorda	Surveyor	IDOE	42.5°N	126.8°W	90°	IDOE 1971 Surveyor Seamap*, Line 36
4	South Atlantic	Vema	V2412	34.0°S	5.0°W	90°	Cochran (1979), Fig. 6
5	North Atlantic FAMOUS	Vema	V3203	36.8°N	33.3°W	113°	This paper, Fig. 2a
6	Labrador Sea	Vema	V2911	57.0°N	44.0°W	25°	Kristoffersen and Talwani (1977), Fig. 1
7	Coral Sea	Vema	V3313	13.8°S	152.1°E	26°	Weissel and Watts (1979), Fig. 1

* Data available from National Geophysical and Solar Terrestrial Data Center (NGSDC).

Vogt, 1973). The absence of large volcanoes in the axial region of the East Pacific rise may be due to the lesser strength of the lithosphere at fast-spreading ridges (Lonsdale, 1977). Thicker, stronger, lithosphere at slow-spreading ridges may explain the presence of high central volcanoes in the axial region of the Mid-Atlantic ridge.

A number of studies have now been carried out of the nature of the isostatic mechanism at seamounts and oceanic islands in the interiors

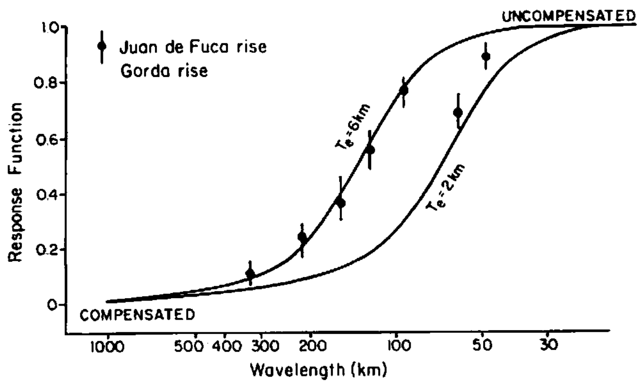

Figure 3. Isostatic response function obtained by McNutt (1980) from gravity anomaly and bathymetry data in the SURVEYOR area of the Juan de Fuca and Gorda rise crests. This function represents the variation in isostatic compensation of ridge crest topography as a function of wavelength. The critical wavelengths for which isostasy is important at ridge crest topography is in the range of about 40 to 300 km. For shorter wavelengths features are uncompensated but for longer wavelengths features are fully compensated. The solid line is a theoretical isostatic response function based on the flexure model for an assumed effective elastic thickness of T_e = 6 km and T_e = 2 km and a crustal structure based on a 2 km thick layer of density 2.6 g/cm^3 and a 3 km thick layer of density 2.9 g/cm^3 overlying a mantle of density 3.4 g/cm^3.

of the plates (summarized in Watts et al., 1980). These studies suggest that the effective elastic thickness T_e of oceanic lithosphere is a strong function of the age of the lithosphere at the time it is loaded and that it does not change appreciably with time. Watts (1978) showed that T_e corresponds closely to the 300°C to 600°C oceanic isotherms, based on simple cooling models.

These studies of the plate interiors have implications for the gravity field over oceanic rifts. First, estimates of T_e for ridge crests (2 to 13 km) may generally constrain the average depth to the 300°C to 600°C oceanic isotherms at the ridge axis. Second, gravity anomalies associated with oceanic rifts like those associated with seamounts and oceanic islands should be supported for long periods of geological time.

Figure 4 shows free-air gravity anomaly and bathymetry profiles over two oceanic rifts in the Labrador sea (Atlantic) and Coral Sea (Pacific) which ceased spreading 40 to 56 m.y.B.P. The amplitude and wavelength of the gravity anomaly associated with these extinct rifts are similar to the gravity anomalies associated with the active rifts (Fig. 1). Thus gravity anomalies over oceanic rifts do not appear to change appreciably with time, in agreement with results of flexure studies in the plate interiors.

A number of models have now been proposed to explain the existence of oceanic rifts. Deffeyes (1970) and Anderson and Noltimier (1973) consider that oceanic rifts occur because of an imbalance between the width of lava intrusion and the distance required to reach the mean spreading rate. Sleep (1969) and Sleep and Rosendahl (1979) suggest a viscodynamic explanation in which oceanic rifts form as a result of viscous head loss of the ascending hot mantle material. Collette et al. (1980), in contrast, consider that oceanic rifts originate due to a viscous lag in the mantle underlying the ridge axis.

These models for oceanic rifts therefore differ from that proposed by Vening Meinesz (1950) for

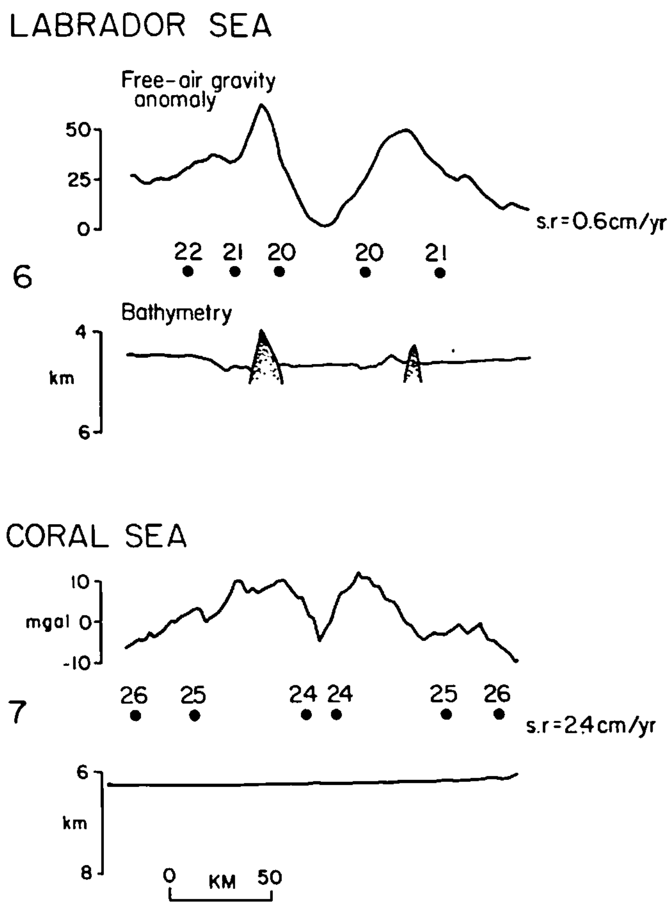

Figure 4. Selected free-air gravity anomaly and bathymetry profiles of the extinct ridge crests in the Labrador and Coral sea. The location of the profiles are given in Table 1. The magnetic anomaly identifications, based on Kristoffersen and Talwani (1977) for the Labrador sea and Weissel and Watts (1979) for the Coral sea, are shown above each bathymetry profile. The Labrador sea, which shows a major slowdown in half-spreading rate at anomaly 20 (about 46 m.y.B.P.), ceased spreading at anomaly 13 (about 40 m.y.B.P.). The Coral sea ceased spreading at anomaly 24 (about 56 m.y.B.P.). The topography of these extinct ridges is obscured by sediments but each spreading center is associated with a prominent free-air gravity anomaly low and a flanking gravity high.

continental rifts. In this model, steep normal faults formed by crustal extension produce a narrow wedge which subsides to form a rift valley. Vening Meinesz (1950) showed that the width of the rift would be expected to be less than about twice the elastic thickness of the crust. The elastic thickness at oceanic rifts, based on gravity and bathymetry studies as well as thermal and petrologic studies, is in the range 2 to 13 km suggesting, on the basis of Vening Meinesz's (1950) hypothesis, that the width of oceanic rifts should be less than about 4 to 26 km. This is in general agreement with the observed rifts in Fig. 1, which are typically less than 30 km in width. Le Pichon et al., (1973) rejected Vening Meinesz's (1950) hypothesis, however, since oceanic rifts are not a single down-dropped wedge between 60° fault scarps but are a series of low dipping (30°) blocks descending toward the axis of the rift.

Recently, Tapponnier and Francheteau (1978)

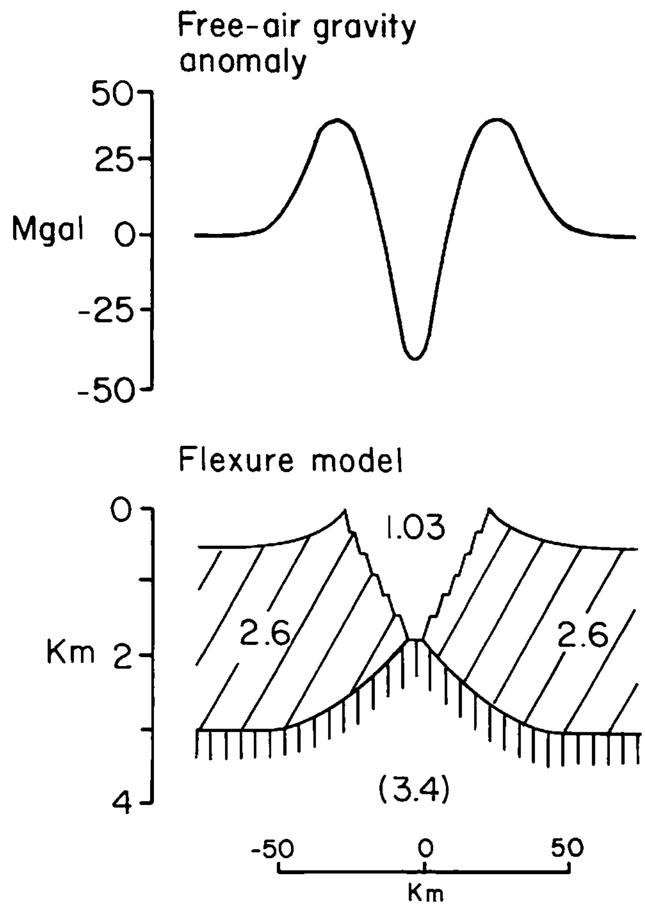

Figure 5. Computed free-air gravity anomaly based on the simple flexure model proposed by Tapponnier and Francheteau (1978) at oceanic rifts. In this model the mass deficit of the rift is balanced by a mass excess at depth. Since a layer of continuous strength is assumed at the ridge axis the balance is not achieved locally but regionally over a broad area. The flexure model shown has been computed assuming a net upward force, equal to the mass deficit, acts on a thin elastic plate with T_e = 4 km (Tapponnier and Francheteau, 1978). The gravity anomaly has been computed using the simple Bouguer slab approximation and the density distribution shown.

proposed a model for oceanic rifts which includes a thin elastic layer at the ridge axis. In this model the oceanic rift is a consequence of steady state necking or stretching of the lithosphere. The region of the rift is not stretched to zero thickness, however, since a uniform thickness of lithosphere is maintained at the rift. A result of the stretching is a restribution of mass from the peripheral regions into the axial region. Isostasy is not achieved locally by crustal thinning but regionally by a broad uparching of the rift. The wavelength of the uparching is wider than the oceanic rift itself and depends on the thickness of the elastic layer assumed.

Gravity data alone cannot, unfortunately, distinguish between these models for oceanic rifts. Sleep's (1969) model for slow-spreading ridges, in which the oceanic rift is caused by a loss of hydrostatic head, could explain the presence of large-amplitude negative gravity anomalies at rifts. This model cannot, however, explain the presence of small-amplitude gravity anomaly highs in flanking regions (Figs. 1 and 4). Tapponnier and Francheteau's (1978) model, which includes an elastic layer at the ridge axis, can explain the observed gravity anomaly patterns (Fig. 5). The main problem with this model, however, is in calculating the relative contribution of the gravity effect of the rift and the uparched crustal layers. Although the crustal structure at oceanic rifts is reasonably well known beneath the rim mountains, it is poorly known beneath the rift.

This review therefore suggests that there has been a significant increase in our understanding of the gravity field and tectonics of oceanic rifts during the past decade. Most information on the nature of the isostatic mechanism at oceanic rifts, however, has come not from studies of the rifts themselves but from geological features which form at or near rifts and are transported away from them by sea-floor spreading. We currently know very little about the nature of isostasy and the relative importance of viscous and elastic forces at rifts. There is accumulating evidence, however, for the existance of an elastic layer at both active and inactive rifts. Further studies using long seismic arrays, in combination with gravity data, appear to hold the most promise to address this problem during the next decade.

Acknowledgments. This work was supported by Office of Naval Research contract N00014-75-C-0210 Scope C and National Science Foundation grant OCE 79-18917. J. R. Cochran and B. J. Collette made some helpful comments on the manuscript. Lamont-Doherty Geological Observatory Contribution Number 3071.

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