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Gravity anomalies, crustal structure and flexure of the lithosphere at the Baltimore Canyon Trough

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Seismic and gravity anomaly data have been used to examine the long-term mechanical properties of the lithosphere beneath the Baltimore Canyon Trough, a 15 km deep sedimentary basin in the rifted U.S. Atlantic margin. Seismic data constrain the crustal structure at the time of rifting as well as the amount of sedimentation and erosion that has occurred. The gravity effect of the initial crustal structure, sedimentary loading and erosional unloading have been computed and compared to the observed free-air gravity anomaly. The best fitting model is one in which the elastic thickness of the lithosphere (T_e) changes at the East Coast magnetic anomaly. Landward of the anomaly, T_e is significantly smaller than previously reported values for continental or oceanic lithosphere of the same thermal age. Seaward of the anomaly, however, T_e follows the depth to the 450°C oceanic isotherm which also describes the response of oceanic lithosphere to seamount and oceanic island loads. These results are interpreted as indicating that the shelf part of the trough is underlain by continental lithosphere which, following rifting, did not acquire any significant long-term strength.

1. Introduction

One of the most well known features of the marine gravity field is the free-air gravity anomaly "edge effect" associated with Atlantic-type continental margins. Typically [1,2], this anomaly consists of a gravity high of about +25 to +75 mgal near the continental shelf break in slope, a low of about −25 to −75 mgal over the rise and, at some margins, a small high seaward of the low.

Early studies [1] attributed the edge effect to the juxtaposition of thick continental crust with thin oceanic crust. Talwani and Eldholm [2] showed that the edge effect could not be fully accounted for if a simple Airy-type isostatic correction for changes in crustal structure and sediment distribution was applied to the gravity anomaly, suggesting that large density differences may exist in the basement.

Walcott [3] and Cochran [4] showed that the fit between observed and computed gravity anomalies is improved if elastic plate, rather than local, models of isostasy were used for the sediment correction. These workers showed that the edge effect over the Mississippi and Amazon river deltas could be explained by the superposition of two effects: one due to the initial margin configuration and the other due to sediment loading. Walcott [3] and Cochran [4] found that the amplitude and wavelength of the calculated gravity anomalies were sensitive to the value of the elastic thickness, T_e, assumed and that the best fit to the observed data for these deltas was in the range 20–30 km.

It has now been established from gravity anomaly studies over seamount and oceanic island loads that T_e for the oceanic lithosphere increases with age at the time of loading [5]. These studies suggest that T_e follows the depth to the 300–600°C oceanic isotherm based on cooling plate models. The results of oceanic flexure studies are generally in accord [6] with the results of experimental rock mechanics, particularly with regard to the yield strength envelope curves constructed by Goetze and Evans [7] based on Byerlee's laws for frictional sliding and the ductile flow laws for olivine.

By way of contrast, there is controversy at present concerning the long-term mechanical properties of continental lithosphere. T_e for continental lithosphere beneath the Ganges and Ap-

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palachian basins is in the range 90–100 km [8,9], which is larger than the values of 40–50 km expected for oceanic lithosphere of the same thermal age. These data suggest that continental lithosphere is stronger than oceanic lithosphere of the same thermal age. T_e beneath the North Sea basin, however, is about 5 km [10] which is smaller than the values expected for oceanic lithosphere of the same thermal age, suggesting continental lithosphere may be weaker.

One of the best examples of loads on continental lithosphere are the sediments which accumulate at Atlantic-type margins. Karner and Watts [11] analyzed the edge effect anomaly over the margins of the northeastern U.S., South Africa and Australia and concluded that the lithosphere underlying the major basins in these margins, irrespective of whether it was oceanic or continental, responds to sediment loads in a similar way as oceanic lithosphere. Their model did not directly incorporate seismic constraints on the initial crustal structure, however.

The purpose of this paper is to use seismic reflection and refraction and gravity anomaly data to quantitatively evaluate the contribution of sediment loading to the edge effect at a rifted Atlantic-type continental margin. We focus the study on the U.S. margin off New Jersey since large amounts of seismic reflection and refraction data are now available from the region. Furthermore, the edge effect is well developed at this margin [12] where it consists of a narrow gravity high flanked by two broad lows. The main objective of the paper is to use the seismic and gravity anomaly data to constrain the long-term mechanical properties of the lithosphere underlying a rift-type basin.

2. Geological and geophysical setting

The New Jersey margin (Fig. 1) has been the site of extensive geological investigations during the past decade [13–15]. The margin is associated by up to 15 km of gently dipping Mesozoic to Tertiary sediments which form a deep basin, the Baltimore Canyon Trough. Basement dips gently seaward beneath the coastal plain towards a hinge-zone beyond which it rapidly increases in depth. Onshore, the Late Precambrian to Early Paleozoic basement contains a number of extensional-type features which formed in response to rifting in Triassic to Early Jurassic time. Similar extensional-type features have been identified in the basement offshore [16].

Seismic reflection profiles, commercial and COST well data, suggest that the first marine transgression began to onlap basement during the Triassic-Jurassic transition [15]. In near shore regions, sands and shales were deposited while in the offshore predominately carbonates accumulated. Along portions of the margin, reefs developed, acting as a barrier to sediment infill. Reef growth probably continued until early Cretaceous times when it became buried by prograding sands and shales. The margin continued to grow during the Late Cretaceous and Paleogene with sediments accumulating on both shelf and slope and rise regions. Rapidly prograding clastic wedges in the Miocene accompanied a return to near shore conditions, a state that probably persisted through to the Neogene.

The deep structure of the trough (Fig. 2) is now reasonably well known following completion of the Large Aperture Seismic Experiment (LASE) in 1982 [17]. The seismic structure consists of an upper sedimentary layer with P-wave velocities in the range 1.7–6.4 km/s, a "crustal" layer with velocities of 6.6(?)–7.5 km/s and a mantle with velocities of 8.0–8.2 km/s. If the prominent reflector on USGS seismic reflection profile 25 at depths of about 10–12 km beneath the shelf [15] marks the first major onlap of basement following rifting, then the thickness of the crustal layer beneath the trough ranges from 18 km at Expanding Spread Profile (ESP) 1 to 8 km at ESP 5. These thicknesses would include Triassic sediments which according to Poag [15] form a number of half-grabens in the basement beneath the trough. Velocities in the lower part of the crustal layer (7.1–7.5 km/s) are more or less continuous across the trough. The LASE study group [17] considered this layer as either oceanic or continental in type. However, the preferred model published by LASE [17, fig. 2] shows oceanic crust extending landward of the shelf break, almost to the hinge zone.

Several prominent belts of gravity and magnetic anomalies are associated with the margin offshore New Jersey. Free-air gravity anomaly maps [12] show a narrow (50–60 km) high (up to

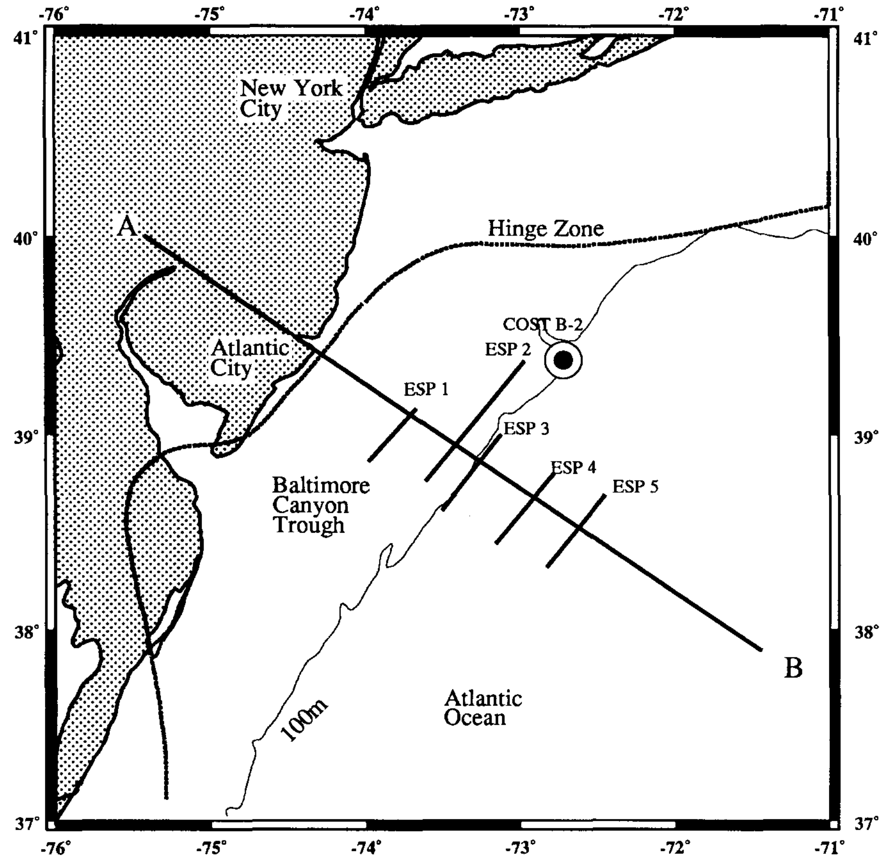

Fig. 1. Location map of study area showing position of the LASE Expanding Spread Profiles (ESP) 1 to 5 [17]. The "hinge zone" (thick dashed line) marks the zone where continental basement rapidly increases in depth. The 100 m contour approximately marks the position of the shelf break in slope. Profile *AB* (thick solid line) is shown in Fig. 2.

30 mgal) over the outer shelf and wide (100 km) flanking lows (up to −75 mgal) over the inner shelf, slope and rise. This edge effect anomaly can be explained by a model [18] in which the major change in crustal thickness at the margin occurs at the hinge zone break rather than at the shelf break as an Airy-type model would predict. The crustal structure based on gravity modelling is therefore in accord with the LASE results. The models cannot, however, constrain whether the trough is underlain by oceanic or thinned continental crust. Magnetic anomaly maps [19] show a broad (150 km) high centered in the region of the shelf break and a flanking low over the slope and rise. The East Coast Magnetic Anomaly, as this anomaly is known, has been interpreted [14] as caused in part by volcanic rocks within the sedimentary column and in part by the juxtaposition of highly magnetized oceanic rocks with weakly magnetized continental rocks.

If the Baltimore Canyon Trough shelf is underlain by continental crust [14,15], then the results of LASE [17], together with the gravity modeling studies, suggest that it has been significantly thinned as a result of rifting. The tectonic subsidence at the COST B-2 well, which was drilled in the outer shelf (Fig. 1), can be explained by a model [20–22] in which the crust and lithosphere beneath the trough was thinned by a factor of 3 or more at the time of initial rifting.

3. Data analysis

In the case of uniform extension, Mckenzie [23] has shown that there are two main contributors to vertical motions: subsidence due to crustal thin-

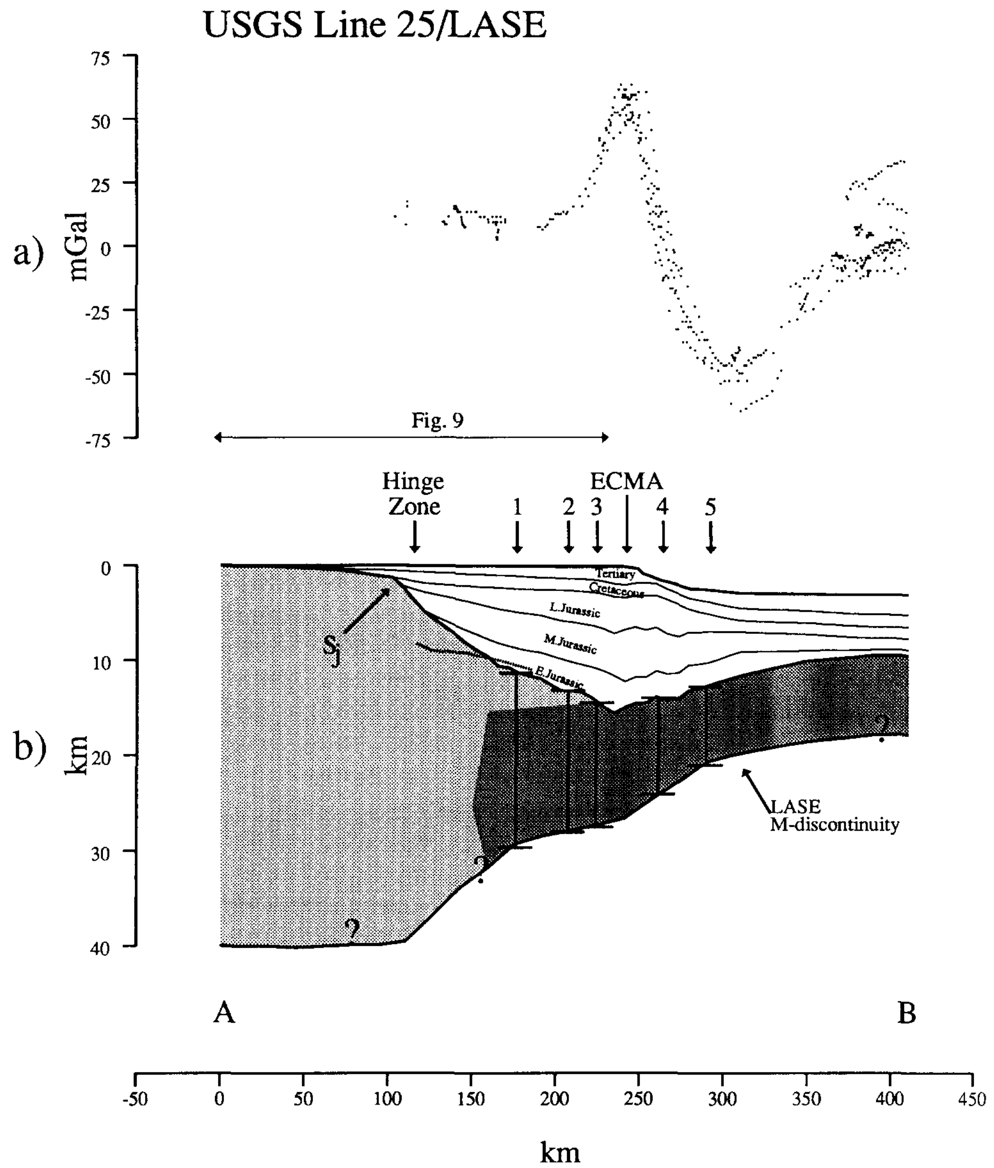

Fig. 2. Free-air gravity anomaly and deep crustal structure along profile *AB* (Fig. 1). (a) Free-air gravity anomaly. Solid dots show the location of gravity measurements that have been projected orthogonally onto the profile. Only gravity measurements currently in LDGO's global marine gravity data bank and within a "window" of 30 km either side of the profile have been plotted. (b) Deep crustal structure based on interpretation of USGS seismic reflection profile 25 [15] and the LASE [17] results. The heavy line showing the base of the Jurassic [15] is in reasonable agreement with the top of the syn-rift sediments (heavy dashed line) inferred by LASE [17]. The vertical bars show the thickness of the crust at LASE ESP 1 to 5. According to LASE, the lightly shaded region is continental crust and darkly shaded region is oceanic. The heavy line at the base of the crust is the M-discontinuity based on LASE [17, fig. 2]. Note that the thickness of the crust is not seismically constrained landward of ESP 1 or seaward of ESP 5. *ECMA* = East Coast Magnetic Anomaly. S_j indicates the approximate position of the Jurassic pinch-out.

ning and uplift caused by lithospheric heating. The magnitude of these motions is determined from the amount of extension, as well as from the initial crustal and lithospheric thickness assumed.

Seismic refraction studies of crustal structure constrain the amount of extension so they should, in principle, reveal the amount of tectonic subsidence and uplift that has occurred within a rift-type basin. In particular, if the crustal thinning and lithospheric heating occurs at the time of rifting, then seismic data may be used to "restore" the configuration that a basin would have had in the absence of the disturbing effects of sedimentary loading and erosional unloading.

The process of rifting involves changes in crustal thickness as well as changes in density due to replacement of cold lithosphere by hot asthenosphere. Extended crust that may have been relatively high shortly after rifting subsides as the underlying mantle cools and increases its overall

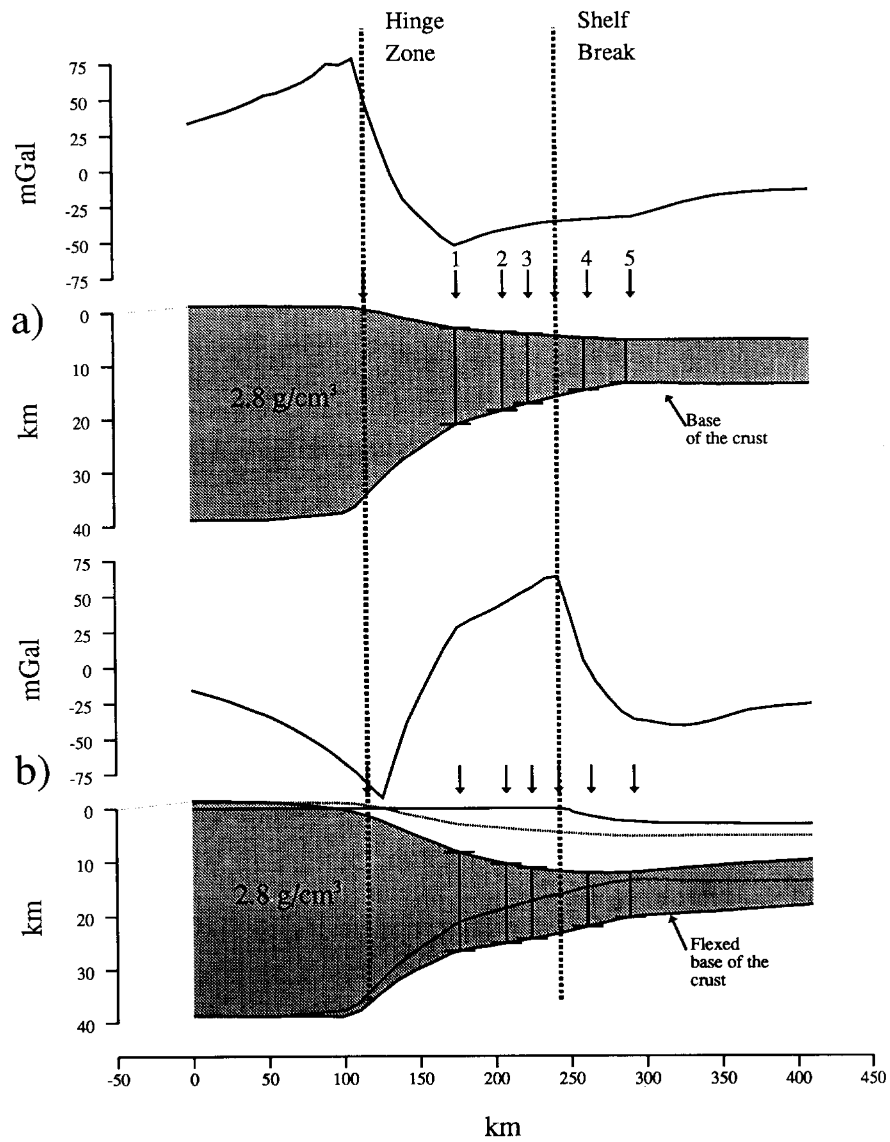

Fig. 3. Calculated profiles of gravity effect and crustal structure based on the thickness of crust measured seismically. Vertical bars are as defined in Fig. 2. The position of the hinge zone and shelf break is from Fig. 2. (a) Crustal structure and gravity effect obtained by isostatically balancing the thinned crust beneath the BCT with an unstretched reference column. (b) Flexed crustal structure and gravity effect assuming sediments infilled the margin to the present bathymetry and loaded an elastic plate with $T_e = 15$ km. The heavy dashed line in (b) shows the initial crustal structure shown in (a).

density. In Mckenzie's [23] model, an isostatic system is assumed in which a column of extended lithosphere is in balance with unstretched lithosphere. This assumption is supported by the nearly zero average free-air gravity anomaly observed over many starved rift-type basins (e.g. Bay of Biscay margin).

A convenient reference section is the "normal" thickness of the crust and lithosphere at sea level. Unfortunately, there are no reliable estimates of crustal and lithospheric thickness beneath the U.S. Atlantic coastal region. The only determinations are those reported by Pakiser and Steinhard [24] based on "regionalized" seismic refraction lines which suggest a crustal thickness in the range 30–40 km for the coastal plain. We will assume therefore, following Cochran [25], a reference section consisting of a crustal and lithospheric thickness of 31.2 and 125 km respectively and a 0°C density of crust and mantle rocks of 2.80 and 3.33 g/cm^3 respectively. Such a section is in balance with a mid-oceanic ridge with a 5 km crust, 1333°C mantle and a water depth of 2.5 km. By balancing an extended column against the reference column it is possible to compute the depth of an extended region, given its crustal thickness, and "restore" the configuration the basin would have in the absence of the modifying effects of sedimentation and erosion.

Fig. 3a shows the restored crust and upper mantle structure for the margin off New Jersey based on the results of the LASE. The restored crustal structure at the ESP sites (vertical bars) is shown together with the structure in flanking regions implied in the geological cross-section published by LASE. As expected, the regions of extended crust are associated with the largest water depths while regions of thick crust correlate with elevated regions.

The gravity effect of the restored margin was computed using a 2-D line integral method [26]. The resulting gravity anomaly edge effect (Fig. 3a) shows a characteristic high just landward of the transition between unstretched and stretched crust and a broad low over the extended region.

As was pointed out earlier, the restored crustal structure and the associated gravity anomaly edge effect would be modified during basin evolution by sedimentary loading and erosional unloading. The present-day relief of the margin consists of an alluvial plain at or near sea level, a gently dipping continental shelf and a slope and rise. The location of the shelf break, some 250 km seaward of the hinge zone (Fig. 2), is indicative of the extensive progradation of the margin while the presence of up to 15 km of sediment beneath the shelf points to aggradation. The total amount of sediment "load" that has been added is obtained by subtracting the restored water depth from the present-day depth. Similarly, since eastern New Jersey is presently near sea-level, the amount of erosion is provided by the height of the restored section above sea level.

Fig 3b shows the flexure and the gravity effect that would result if sediments of uniform density 2.6 g/cm^3 were loaded on the restored margin, assuming it behaved as an elastic plate with a thickness, T_e, of 15 km. The region of greatest applied load is associated with the largest flexure (up to 6 km). The total sediment thickness, which is obtained by adding the sediment load to the flexure, amounts to some 15 km. This thickness is in reasonable agreement with observed thicknesses underlying the trough. The gravity effect also reaches its largest values (up to +75 mgal) over the region of the greatest load. Flanking the high are two lows, produced by the displacement of relatively dense mantle material by low density crustal rocks.

The gravity effects shown in Fig. 3 only represent components of the expected anomaly. The observed gravity anomaly is the sum of all the processes that control margin development which include not only the effects of rifting and sedimentary loading, but also erosional unloading. In order to estimate the latter component, the gravity effect of the uplifted region of the margin landward of the hinge zone and the thickened (compensating) crust beneath the region was also computed.

Fig. 4 shows the "sum" anomaly together with the individual contributions from rifting, sedimentation and erosion. Since sedimentation involves the addition of material to the crust, it was added to the rifting effect. Erosion involves the removal of mass and so was subtracted. The sum anomaly shows a gravity high centered over the self break and flanking lows over the hinge zone and slope and rise regions. The principal contribution to the high comes from the effect of sedimentary loading while the flanking lows are determined mainly by

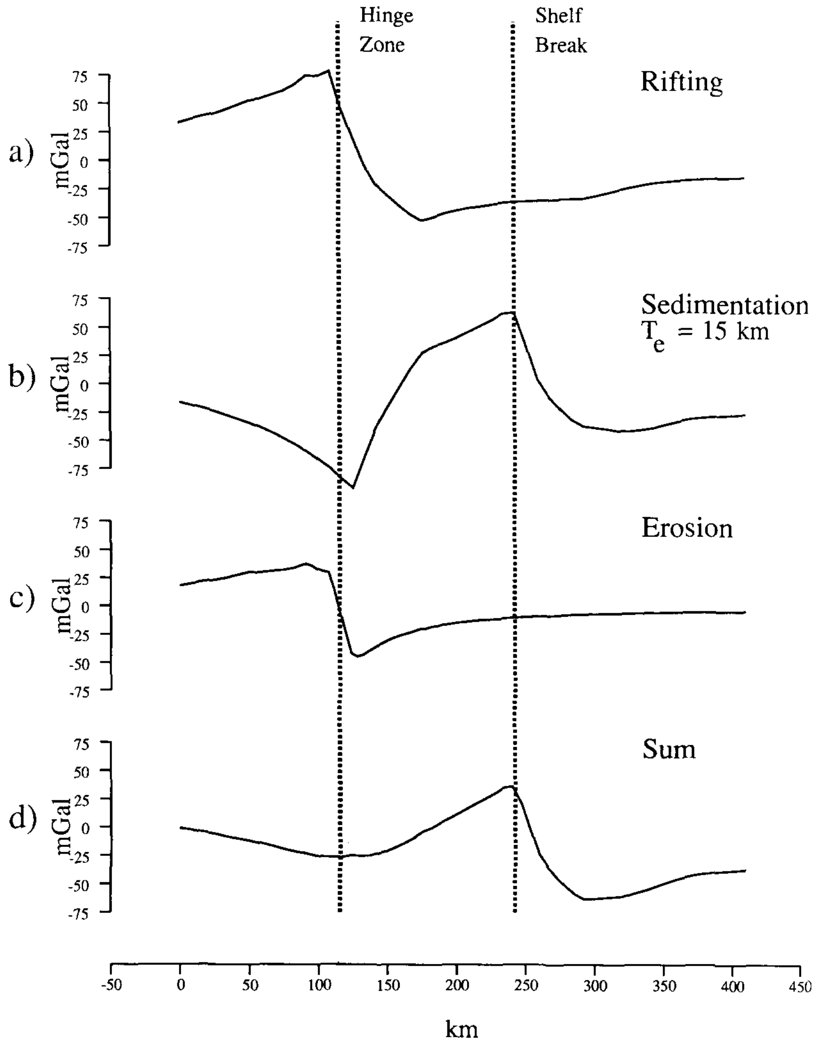

Fig. 4. Gravity effects associated with the restored margin and its modification by sediment loading and erosional unloading. (a) Gravity effect of initial crustal structure (Fig. 3a). (b) Gravity effect of sediment loading an elastic plate with $T_e = 15$ km. (c) Gravity effect of erosional unloading assuming the material above sea-level and the compensating crustal "root" is removed. (d) Sum anomaly obtained by adding (a) and (b) and subtracting (c). The position of the hinge zone and the shelf break is based on Fig. 2.

the downward flexure of basement and the base of the crust. The landward high associated with the restored margin (Fig. 2a) is more than compensated by the large amplitude lows that result from flexure.

We compare in Fig. 5 the computed gravity anomaly assuming $T_e = 15$ km to observed free-air gravity anomalies over the trough. Also shown in Fig. 5 are computed curves for $T_e = 5$ and 25 km. The figure shows that the amplitude and wavelength of the gravity anomaly is a strong function of the value of T_e that is assumed. The best overall fit to the observed anomaly is for $T_e = 15$ km. Significant discrepancies exist, however, between this profile and the observations especially in flanking regions of the slope and rise and the middle shelf.

Fig. 5 suggests that it should be possible to use seismic data, together with the observed gravity anomaly, to constrain T_e of the lithosphere underlying a rift-type basin. In this case, T_e refers to the long-term thermal and mechanical properties of the lithosphere, particularly with regard to its response to loading following rifting. It is im-

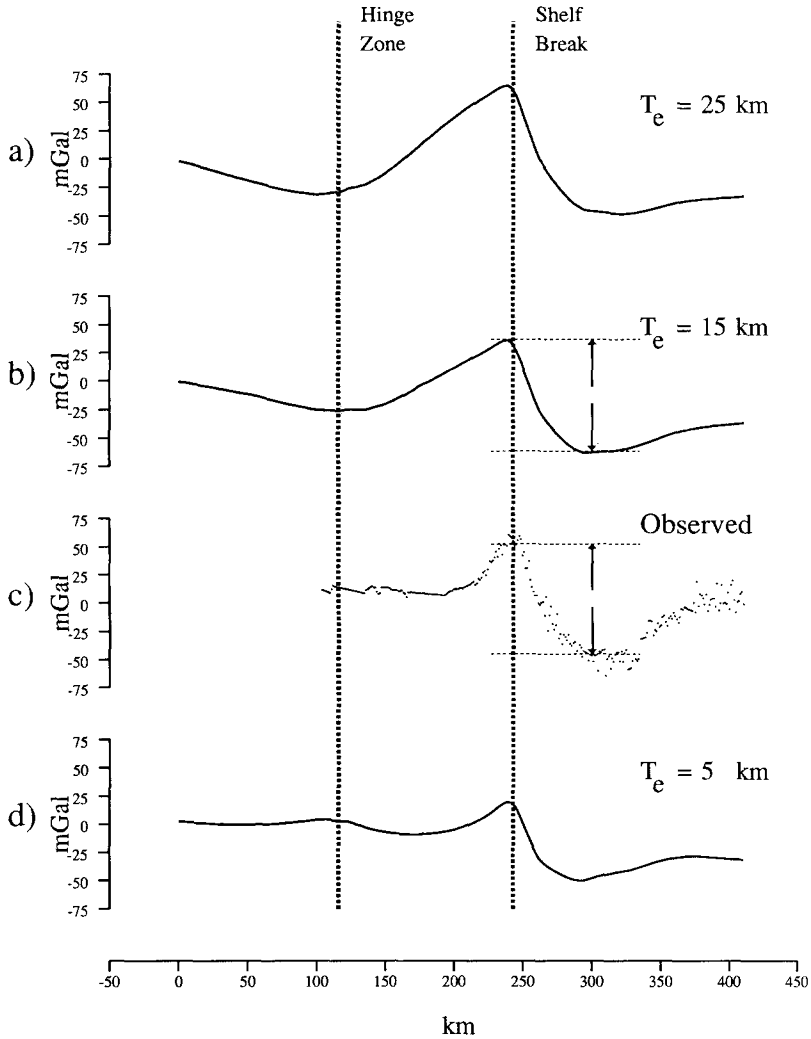

Fig. 5. Comparison of observed and calculated gravity anomalies for different values of T_e. (a) $T_e = 25$ km. (b) $T_e = 15$ km. (c) Observed gravity anomaly based on Fig. 2. (d) $T_e = 5$ km. The arrows highlight the similarity in peak to trough amplitude between the observed and the calculated anomaly for $T_e = 15$ km.

portant to point out that, even though a local-type isostatic balance was assumed to compute the initial configuration of the margin, T_e is independent of crustal structure and depends only on the lithosphere's intrinsic strength.

The main difficulty with the calculated profiles in Fig. 5 is that they assume the variations in crustal structure measured seismically were produced by extension at the time of rifting and that no subsequent modification of the crust, such as underplating [27], occurred following rifting. Furthermore, the models assume that the sediments loaded a lithosphere of uniform T_e. It is possible that changes in crustal thickness or T_e following rifting could explain some or all the differences between the observed and calculated gravity anomalies in Fig. 5.

4. Flexural backstripping

An alternative approach, which allows for changes in both crustal thickness and T_e following rifting, is to use the backstripping technique [29]. In this technique, an attempt is made to correct the geological record for the disturbing effects of sedimentation and erosion and to isolate the form of the tectonic subsidence and uplift in a basin through time. The usual procedure is to first restore the thickness and depth of a sedimentary layer, taking into account compaction, changes in the sediment infill and sea-level changes. The restored layer is then unloaded from the basement using an Airy-type model.

Flexural backstripping follows a similar procedure except that sediments are flexurally unloaded from the basement. It can be shown that the flexure of an elastic plate, y, due to a harmonic load of the form $h \cos(kx)$ is also harmonic and is given by:

$$y = \frac{(r_s - r_w) hg \cos(kx)}{(r_m - r_{infill}) g + Dk^4}$$

where r_s, r_w, r_m, r_{infill} are the densities of the sediment, water, mantle and infilling material respectively, h is the load height, g is the acceleration due to gravity, $k = 2\pi/\lambda$ where λ is wavelength and D is the flexural rigidity of the plate. For wide loads, $k = 0$ and:

$$y = (r_s - r_w) h / (r_m - r_{infill})$$

Let us now replace the load h by S^*, which in the case of a sedimentary unit is the entire load acting on the plate, and assume $r_{infill} = r_w$. Then:

$$y = (r_s - r_w) S^* / (r_m - r_w)$$

Since $S^* = y + h$, it follows that:

$$h = (r_m - r_s) S^* / (r_m - r_w) \quad (1)$$

which is Airy backstripping [21] with zero water depth and no sea-level change. Flexural backstripping therefore amounts to first computing the flexure due to the total sediment thickness with $r_{infill} = r_w$ and then subtracting the resulting flexure from the observed sediment thickness.

In order to apply flexural backstripping techniques to the margin off New Jersey, the stratigraphic sequences identified by Poag [15] on depth converted seismic reflection profiles were digitised and used to construct profiles of sediment thickness across the margin. Each sequence was then flexurally backstripped taking into account an appropriate value of r_s and T_e and the result summed to obtain the total tectonic subsidence and uplift. For purposes of the initial calculation $r_s = 2.6$ g/cm^3 and:

$$T_e = 2.04 t^{1/2} \quad (2)$$

where t is the age (m.y.) since rifting and T_e is the elastic thickness (km) were assumed. The constant term in equation (2) was chosen so that T_e approximately corresponds to the depth of the 300°C oceanic isotherm based on the cooling plate model [28]. While it is convenient to express the variation of T_e with age in this way, it is important to point out that the temperature structure is not explicitly modelled in backstripping. Rather, each sedimentary sequence is backstripped for different values of the constant in equation (2), thereby simulating plates of varying long-term strength.

The results of flexural backstripping are compared to the restored configuration of the margin based on seismic refraction data in Fig. 6. The overall pattern of subsidence and uplift determined using the two methods is similar, with the greatest changes occurring in the vicinity of the hinge zone. The main differences occur in regions flanking the trough. The refraction data imply a broad regional uplift landward of the hinge zone whereas the reflection data suggest a narrow "rim" uplift. There are no reliable constraints on crustal thickness landward of the hinge zone, so it is likely that the backstripping approach is a better indicator of tectonic movements in this region. The rim develops early in margin evolution and is required in order to explain the absence of the Jurassic landward of the hinge zone. By way of contrast, the refraction data shows little or no uplift seaward of the shelf break whereas the reflection data suggests a broad uplift. Again, the constraints from seismic refraction data are poor in this region.

The uplifts revealed by flexural backstripping (Fig. 6) are similar in overall form to uplifts described from young rift-type basins. For example, the uplift landward of the hinge zone is strikingly similar to border uplifts flanking young rift-type basins such as the Red Sea [30] while the broad uplift is similar in form to the outer rises that have been claimed [31] to flank some Atlantic-type margins. Several models have been proposed to explain these uplifts. These include thermal bulges due to lateral flow of heat [32], flank uplifts due to some form of small-scale convection [30], outer rises due to detachment faulting and unloading of lower crustal rocks [33], and uplifts that result from underplating of stretched crust some time after the extraction of

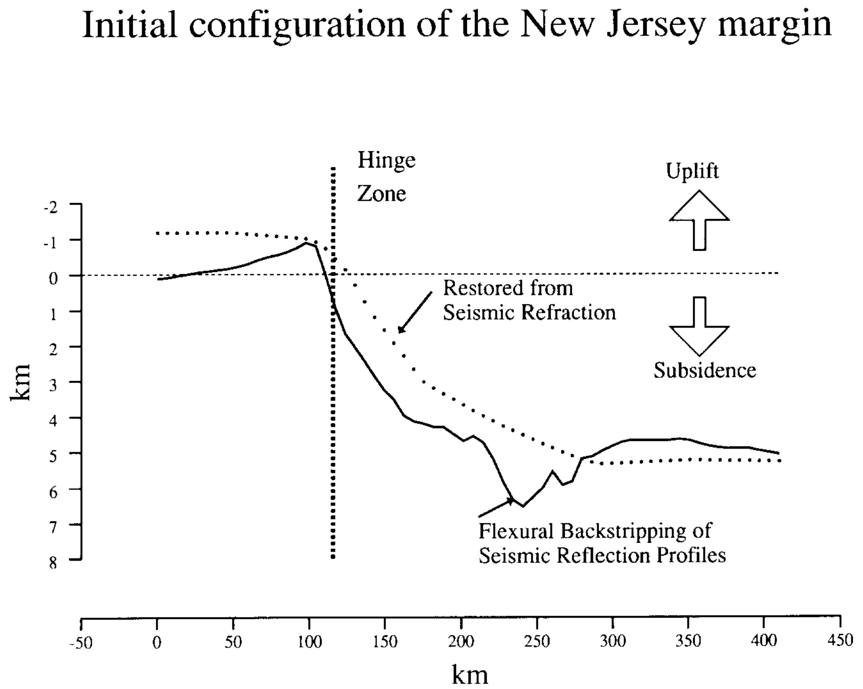

Fig. 6. Comparison of the initial water depth of the margin obtained by restoring the crustal structure based on refraction data (Fig. 3a) with that obtained by flexural backstripping of the reflection data. The water depth is the depth the margin would have been in the absence of sedimentary loading and erosional unloading. Note the relatively high flanking regions.

basaltic magmas to produce oceanic crust occurs [27]. Unfortunately, it is not possible to distinguish between these different models. The amplitude and wavelength of the uplifts depends critically on the value of T_e that is assumed. For example, if T_e follows a lower isotherm than assumed in Fig. 6 then the border uplifts are reduced in extent and it may not be necessary to invoke other tectonic mechanisms to explain them.

One of the largest discrepancies between the results of flexural backstripping and the restored margin occurs in the vicinity of the shelf break. The backstripped curve exceeds the restored margin depths by almost 2 km. This discrepancy is attributed to the assumption of a uniform density for the unloaded sediments. If the average density is higher than 2.6 g/cm^3 due, for example, to the presence of a carbonate reef complex at the shelf break [15] the subsidence due to sediment loading would locally increase, but the tectonic subsidence required would decrease. The flexural backstripping would, in this case, more closely agree with the results of restoring the margin using refraction data.

As was the case for the restored margin, the gravity anomalies associated with flexural backstripping can be computed and compared to the observed data. Since backstripping yields the depth the margin would have had in the absence of sedimentation and erosion, the corresponding crustal structure can be calculated by isostatic balancing. Similarly, the gravity edge effect and the individual gravity components due to sedimentation and erosion may be computed—the amount of flexure being determined by subtracting the backstripped curve from the isopach for each sedimentary unit.

Fig. 7 show the sum gravity anomaly from flexural backstripping, together with two other cases where T_e follows the depth to the 150 °C and 450 °C oceanic isotherms. These cases represent the gravity anomaly that would be expected for weaker and stronger plates respectively. The figure shows that the gravity anomaly is quite sensitive to the relationship that is assumed between T_e and age since rifting. The best overall fit to the peak to trough amplitude of the observed data is for the case of a plate in which T_e follows

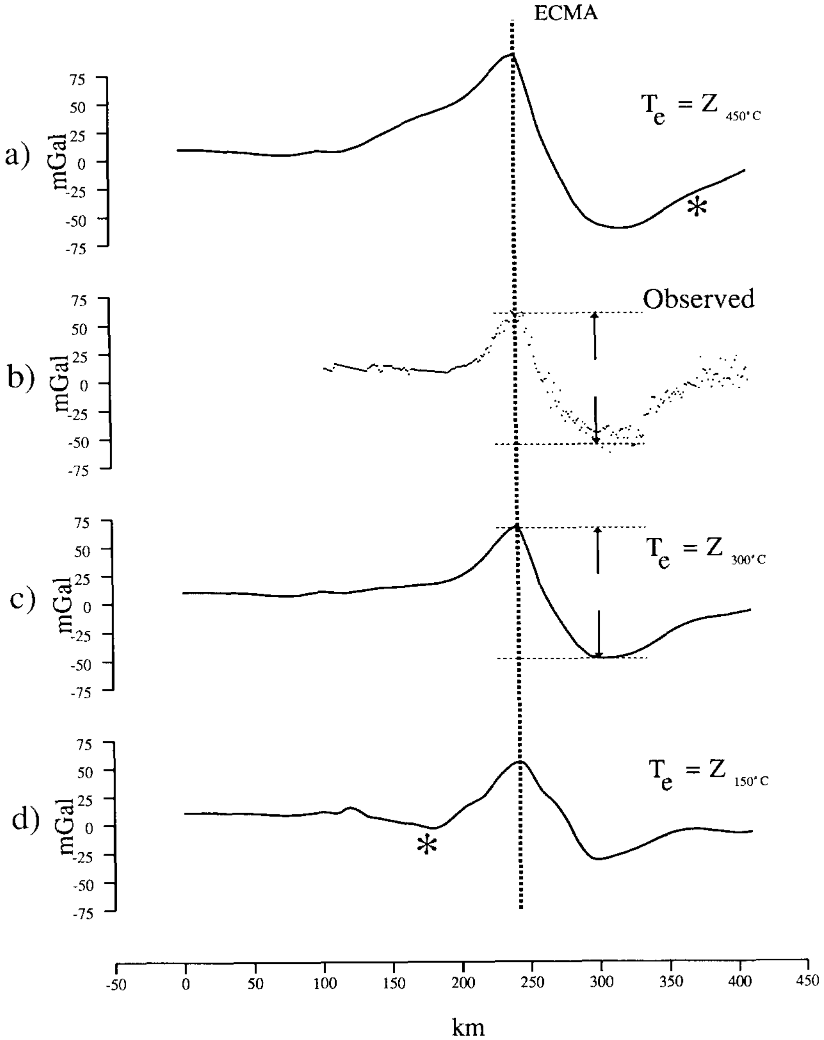

Fig. 7. Comparison of observed and calculated gravity anomalies for different values of T_e following rifting. (a) $T_e = Z_{450°C}$. (b) Observed gravity anomaly based on Fig. 2. (c) $T_e = Z_{300°C}$. (d) $T_e = Z_{150°C}$. The asterisks highlight the region of particularly close fit between the calculated for $T_e = Z_{450°C}$ and $T_e = Z_{150°C}$ and the observed and the arrows highlight the close fit of the peak to trough amplitude between the calculated for $T_e = Z_{300°C}$ and the observed.

the depth to the 300 °C isotherm. However, the fit to the wavelength, although an improvement on the case in Fig. 5, still shows discrepancies particularly in shelf and slope regions. In shelf regions, the calculated anomaly does not show the small minimum seen in the observed data. A better fit in this region would appear to be for a plate where T_e follows the depth to the 150 °C isotherm. Also, in slope regions the calculated anomaly does not show the steep gradient seen in the observed data. A better fit in this case would appear to be for a plate in which T_e follows the 450 °C isotherm.

As pointed out earlier, oceanic flexure studies show that T_e follows the depth to the 450 °C oceanic isotherm [5]. The good fit of the data seaward of the shelf break to this isotherm can therefore be attributed to the loading of slope sediments on relatively strong oceanic lithosphere. The fact that some seamount and oceanic island loads show low values of T_e has been attributed to either thermal rejuvenation of the lithosphere [34] as it passes over a mantle hotspot, some form of local reheating of the lithosphere during volcano emplacement [35] or emplacement of seamount and oceanic island loads over anomalous regions in the mantle such as French Polynesia [36]. It would be expected, however, that once these temperature perturbations are removed then T_e would eventually return to normal values for its age.

There is no evidence that T_e for oceanic lithosphere follows the depth to an isotherm as low as 150 °C. Even in the case of the Amazon Cone, where more than 10 km of sediments loaded oce-

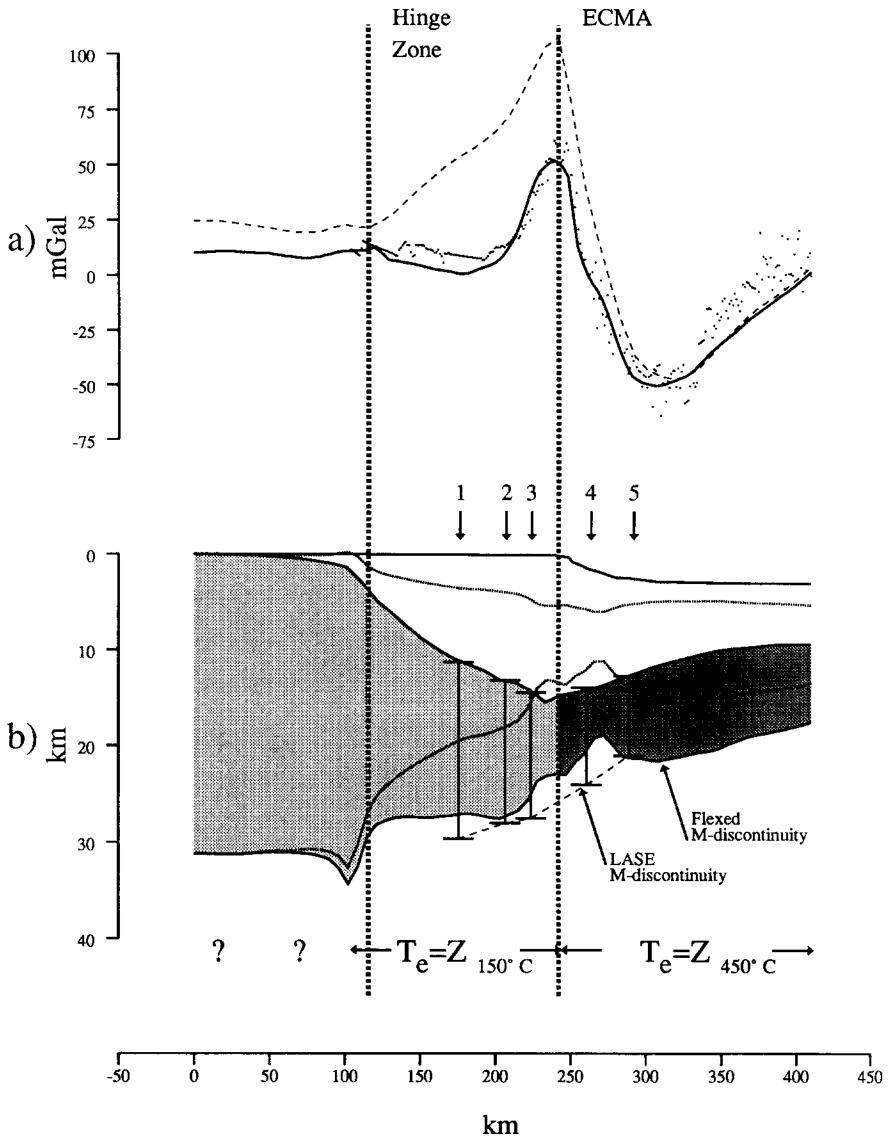

Fig. 8. Comparison of the best fitting model to the observed gravity anomaly and seismic data. (a) Gravity anomaly. The solid dots are the measured gravity values from Fig. 2. The thin dash line is the calculated anomaly assuming the sediments loaded an elastic plate in which T_e varied with age following rifting and is given by the depth to the 450 °c oceanic isotherm. The thick line is for T_e given by the depth to the 450 ° C isotherm seaward of the ECMA and the 150 ° C isotherm landward of the ECMA. (b) Crustal structure implied by the model in which T_e follows the 450 ° C isotherm seaward of the ECMA and the 150 ° C isotherm landward of the ECMA. The heavy dash lines show the initial configuration of the margin based on flexural backstripping. The upper thick line is the present water depth, the middle thick line is the flexed basement and the thick lower line is the base of the flexed crust. The thin dash line is the depth to the M-discontinuity based on LASE ESP 1 to 5 (Fig. 2).

anic lithosphere, T_e [4] appears to follow the depth to the 450 °C oceanic isotherm [6]. Thus, shelf sediments in the Baltimore Canyon Trough loaded a plate that is significantly weaker than normal oceanic lithosphere. Separating the shelf from the slope is the East Coast magnetic anomaly which some investigators [14] interpret as the result of the juxtaposition of continental and oceanic crust. One possibility therefore is that shelf sediments have loaded continental lithosphere which is very weak, at least in comparison to oceanic values and previously reported values from foreland basins on continental lithosphere.

In order to test this possibility, a simple model was constructed in which T_e varied both with age since rifting and position across the margin (Fig. 8). Landward of the East Coast magnetic anomaly, T_e was given by the depth to the 150 °C isotherm whereas seaward of the anomaly it followed the 450 °C isotherm. A finite difference modelling technique [6] was used to flexurally backstrip the trough with a variable T_e and the corresponding gravity anomalies computed in a similar way as in Fig. 5. The gravity effect of the model explains well the amplitude and wavelength of the observed data not only in the vicinity of the shelf break but also over flanking shelf and slope regions.

We caution, however, that the model in Fig. 8 is a simple one that involves a limited number of bodies and uniform density contrasts. The approach taken in this paper has been to compute the gravity effect of the different processes responsible for margin formation and compare it to observations. Thus, no attempt has been made to find by "trial and error" the number of bodies that will best fit the observed gravity anomaly. The significance of Fig. 8 is that it shows that flexural loading of a basement in which T_e varies across the margin can account for a significant component of the observed gravity anomaly, even though a large part of this loading appears to have involved relatively weak continental lithosphere.

5. Depth to the M-discontinuity, flexure and underplating

It has been known for some time [37] that Atlantic-type margins are associated with magmatic rocks, usually in the form of sills and dykes, flood basalts and alkaline igneous rocks. For example, basaltic lavas characterize the upper part of the Triassic section of eastern North America. According to Cox [38] an important first step in the generation of basaltic lavas is the ponding of magma at or near the base of the crust. It is not unreasonable to expect, therefore, that Atlantic-type margins may in some cases be underplated by magmatic rocks during or following rifting.

Underplating in the sense used by Cox [38] refers to igneous material that has been added within or below the crust. A necessary consequence of underplating at an Atlantic-type margin, therefore, would be to thicken the crust and depress the M-discontinuity below the depths that it would be expected to be on the basis of its rifting, sedimentary and erosion history.

White et al. [27] suggest that a 10–15 km thick lens of material with a seismic velocity of 7.3–7.4 km/s beneath the Rockall Plateau margin are underplated igneous rocks. The Rockall Plateau is located within the Thulean lava field which included Greenland, Iceland, the Faeroes and the coastal areas of northeastern Ireland and west Scotland. The seaward-dipping reflector sequences mapped on the slope off Hatton Bank [27] are probably oceanic analogues of these flood basalts, which formed by sea-floor spreading during the earliest stages of rifting. White et al. [27] suggested that the underplated material was left after extraction of these extrusive lavas.

In the case of the New Jersey margin, a test of whether additional material has been added to the crust or not is to compare the depth to the M-discontinuity to the depths expected for its thermal and mechanical evolution. Fig. 8 shows that there is a reasonably good agreement between the observed depths and the predicted depths based on flexural loading. Thus, the M-discontinuity is at its expected depth and there is no need at this margin to invoke thickening of the crust by underplating or any other mechanism.

The problems remains, however, of the origin of the 7.1–7.4 km/s layer that underlies the Baltimore Canyon Trough. The high-velocity layer may represent magmatic material in the form of dykes that have intruded the crust but have not substantially thickened it. The absence of highly reflective layers beneath the trough argues for dykes intrusions, although it may be that they could not be observed because of poor resolution of the seismic reflection data.

The most striking difference between the structure of the trough (Fig. 8) and other margins is the large thickness of sediment. Other differences occur, especially when the nature of the underlying crustal material is taken into account. At the Voring [39] and Hatton Bank [27] margins, a seaward dipping sequence of reflectors overly either thick oceanic crust or stretched continental crust. These reflectors have been interpreted as the result of voluminous extrusion of basaltic lavas at or near the ridge axis. By way of contrast, the Bay of Biscay margin [40] appears to be associated with an absence of magmatism. If the 7.1–7.4 km/s layer at the trough represents magmatic intrusions in an otherwise normal thickness crust, then this margin can be considered of intermediate type at least in regard to the degree to which it has been affected by magmatism.

6. Discussion

This study suggests that the lithosphere underlying the Baltimore Canyon Trough has been one of low strength compared to oceanic lithosphere through much of its post-rift development.

The relative strength of oceanic and continental lithosphere can, as shown by Molnar and Tapponier [41] and Vink et al. [42], be considered in terms of the yield strength envelope. The yield strength envelope for oceanic lithosphere is determined by a combination of Byerlee's rock friction laws and the flow laws for olivine. By way of contrast, the continental lithosphere contains more siliceous rocks which deform by ductile flow at lower temperatures than olivine. On this basis it has been suggested [41,42] that continental lithosphere should be weaker that oceanic lithosphere.

It is difficult, however, to apply the yield strength envelope model to the case of continental rifting because of uncertainties in the geotherm, rheology, and the initial thickness of the crust and lithosphere. Oxburgh [43] argued that even though the continental crust and lithosphere may be initially weak because of its high quartz content, it should eventually regain its strength as it cools and more material with a high olivine content is involved in supporting a load. Kusznir and Park [44] showed that the strength of extended continental crust and lithosphere would depend not only on the amount of thinning but also on the strain rate. Rapidly extending basins may appear less rigid than "normal" because the high geothermal gradient has insufficient time to equilibrate while slowly extending basins may appear more rigid. In both cases, however, the strength of rifted continental crust and lithosphere approaches the limit set by the deformation laws of quartz and olivine and so would still be expected to increase with age following rifting.

This study suggests that even after 200 m.y. following a rifting event, the continental lithosphere has not acquired any significant long-term strength. The best fit to the observed gravity anomaly is for a model in which T_e followed the depth to the 150°C. However, it is difficult to distinguish between this case and one in which the continental lithosphere beneath the trough had no strength.

If this is correct, then it is difficult to explain the large values of T_e that have been determined from studies of the gravity anomaly over foreland basins. The gravity anomaly "low" over the Ganges, Appalachian and Alps [9] basins have wavelengths of the order of 300–400 km, for example, and require values of T_e in the range of 60–100 km in order to explain them. These values are greater than expected for oceanic lithosphere of the same thermal age and would imply that the continental lithosphere is capable, at least in some areas, of considerable long-term strength. In order to explain the results of this study then, either (a) the continental lithosphere appears as a much weaker structure in its response to large-scale "side-driven" loads such as extension than it does for small-scale flexural loads, (b) the continental lithosphere shows considerable weakness in thermally reactivated regions due to the strong dependence of strength on the homologous temperature and the low melting temperature of granite, or (c) the continental lithosphere is susceptible to the effects of sediment blanketing [45] which raise temperatures by a sufficient amount to significantly weaken it.

Irrespective of the actual cause, Fig. 8 shows that the free-air gravity anomaly edge effect is a sensitive indicator of lithospheric strength at Atlantic-type margins. For example, if T_e of the crust and lithosphere beneath the Baltimore Canyon Trough followed the depth to the 450°C isotherm then much larger anomalies would be predicted over the margin than are actually observed. The differences exceed 50 mgal (Fig. 8) over the middle shelf suggesting that it should be possible to use gravity anomaly data to distinguish whether a rift-type basin in underlain by oceanic or stretched continental crust: the gravity anomaly should be relatively high and of long wavelength over basins underlain by oceanic crust and relatively low and short wavelength over basins underlain by stretched continental crust. We would interpret, therefore, the large amplitude (up to 90 mgal) and wide (100 km) gravity high over continental slope between Svalbard and Norway [2] as a result of sedimentary loading of relatively strong oceanic lithosphere and the low amplitude (up to 45 mgal) and narrow (50 km) high over the slope between Shetland and Norway [2] as the result of loading weaker continental crust. These interpretations are consistent with tectonic studies of the regions which show that between Svalbard and Norway oceanic crust extends as far as the Senja escarpment [46] in the vicinity of the shelf break and that stretched continental crust underlies most of the North Sea basin [10].

It has been argued [45,47] that flexure is a major control on the development of stratigraphic patterns at the edge of rift-type basins. Watts and Thorne [47], for example, argued that the best fit to onlap patterns at the edge of the Baltimore Canyon Trough was for a model in which T_e followed the depth to the 450°C isotherm. Their preferred model was based on a two-layer extensional model in which lithospheric thinning competes with the effects of flexure producing coastal plain emergence. Beaumont et al. [45] showed that stratigraphic patterns offshore Halifax can be explained by a one-layer model if T_e follows the 250°C isotherm instead of the 450°C isotherm. They did not need to invoke such large amounts of uplift in the coastal plain as Watts and Thorne [47] since flexure landward of the hinge zone was already reduced in their models by the use of a weaker lithosphere.

The results of this study suggest a limited role for flexure in contributing to the onlap patterns at the edge of rift-type basins. A similar conclusion has been drawn recently by Fowler and Mckenzie [48] on the basis of a low T_e for the North Sea basin [10] and SEASAT gravity anomaly data over the Rockall and Exmouth plateaus. White and Mckenzie [49] have shown that, in the case of the western margin of the North Sea basin, onlap patterns can be produced if the lithosphere has negligible flexural strength, provided that the crust and mantle are stretched by equal amounts and the mantle stretching occurred over a slightly larger area than the crustal stretching.

In order to test this hypothesis, a simple stratigraphic model was constructed (Fig. 9) and compared to observed patterns of onlap at the edge of the Baltimore Canyon Trough. The model is similar to the one used by White and Mckenzie [49] and includes a lithosphere of negligible strength, horizontal as well as vertical heat conduction [32], and a gaussian distribution for the crustal and mantle stretching. In contrast to previous models [45,47], subsidence of the coastal plain results from thermal contraction landward of the hinge zone rather than flexure due to sediment and water loading seaward of the hinge. Mantle stretching, which is assumed to extend over a broader area than the crustal stretching, causes sufficient uplift in the coastal plain for the Jurassic to pinch-out about 25 km landward of the hinge zone (Fig. 9). Continued cooling of the underlying lithosphere causes younger units to progressively onlap the basement. A comparison of the predicted pattern of onlap (Fig. 9) to the observed pattern (Fig. 2) shows good overall agreement. The main discrepancy is in the proportion of predicted syn-rift to post-rift sediments which can be attributed to the use of an instantaneous rather than a finite rifting model. Fig. 9 shows there is reasonable agreement between the predicted and observed depths to basement. The main discrepancies occur in the predicted depths either side of the hinge, due probably to neglect in the model of the effect of compaction and variable sediment supply and paleobathymetry.

Fig. 9 suggests that thermal contraction and uplift of the crust and lithosphere as it cools following a rifting event may be a major control on the development of stratigraphic patterns at the edge of a basin. It is not known, however, whether a model such as the one used in Fig. 9 is applicable to all rift-type basins. Flexure may still be important in basins formed on old parts of the craton which, because of their deep crustal roots, show unusually great strength or in basins where the rate of extension is slow. Irrespective of their

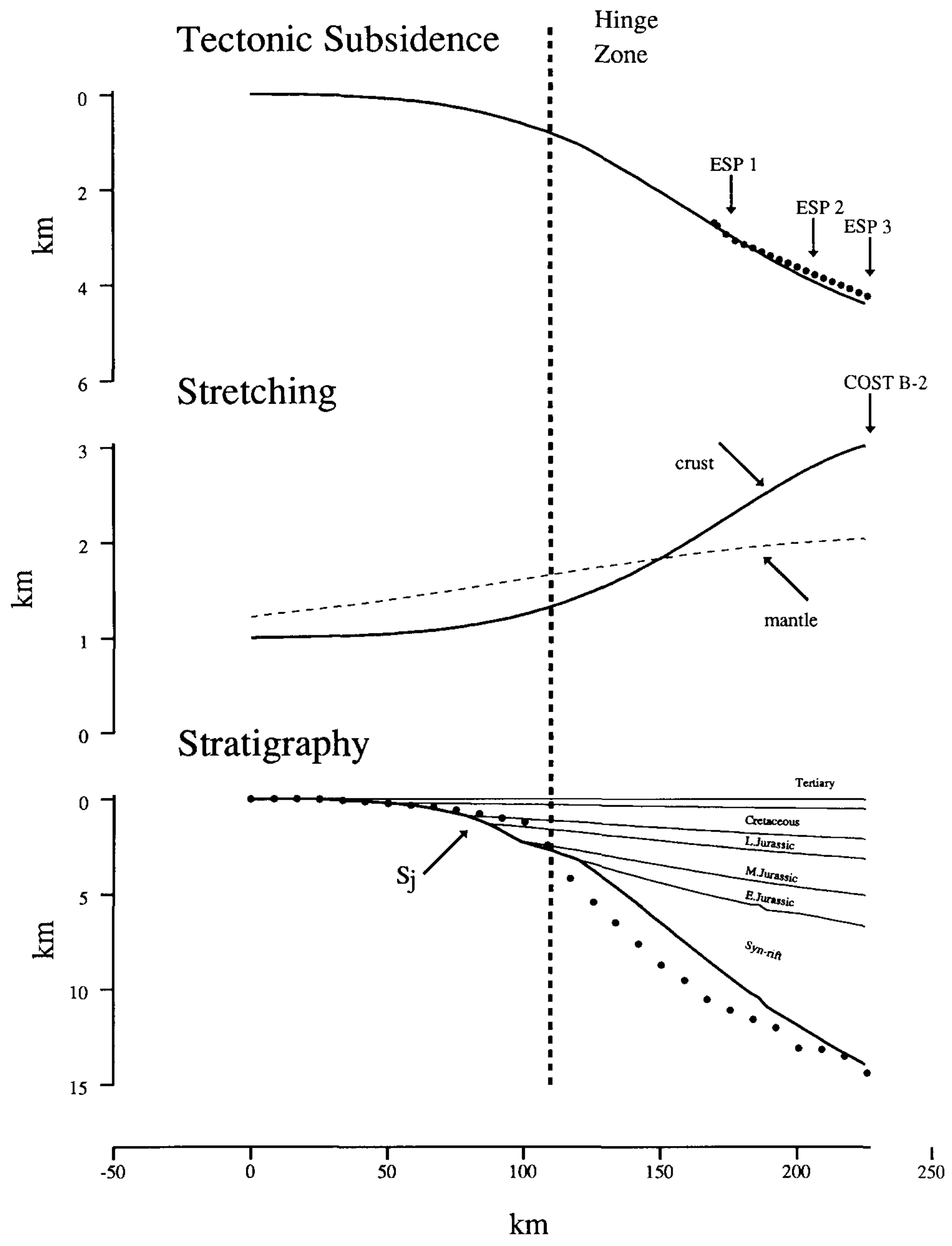

Fig. 9. Tectonic subsidence, stretching and stratigraphy in the vicinity of the hinge zone. The stratigraphy has been computed using a 2-D thermal model with lateral heat flow [32] and the distribution of crustal and mantle stretching shown. The model is based on $T_e = 0$ km. Other thermal parameters are based on Watts and Thorne [47]. The maximum value of the crustal stretching was based on backstripping studies at the COST B-2 well. The distribution of crust and mantle stretching was assumed to be approximated by a gaussian curve with standard deviation $\sigma = 65$ km and $\sigma = 140$ km respectively. The distribution of crustal stretching was chosen so as to best fit the tectonic subsidence inferred from the thickness of the crust at ESP 1, 2 and 3. S_j as defined in Fig. 2. The solid dots on the upper curves show the calculated tectonic subsidence based on the crustal thickness at ESP 1, 2 and 3. The solid dots on the lower curves show the observed depths to basement (Fig. 2).

relative importance, thermal contraction and uplift, like flexure, are examples of the tectonic control on the development of stratigraphic sequences. It is still not necessary, therefore, to appeal to other causes such as large eustatic sea-level changes [50] as the primary cause of onlap patterns at the edge of rift-type basins.

Finally, a weak zone would be expected to profoundly effect structural styles in the basement, especially if the margin later became involved in compressional-type deformation. A weak zone, for example, should show strongly inverted structures [51]. It should possible, therefore, in those ancient terrains, which are suspected to have originated as Atlantic-type margins to use the intensity of inverted structures to effectively map out the extent of the former stretched, weak, continental crust and thereby document the boundary separating unstretched from extensively heated and thinned continental crust.

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