ON ISOSTASY AT THE EQUATORIAL MARGIN OF BRAZIL

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ABSTRACT The cross-spectral method of studying isostasy was applied to profiles of free-air gravity anomalies and topography from the equatorial continental margin of Brazil. The experimental isostatic response function, or admittance, deduced in this study is compared with theoretical admittance functions predicted by a local and regional compensation mechanism. The results show that the most appropriate isostatic mechanism for this area is one in which the topography is not locally compensated, but regionally compensated by flexure of the lithosphere. Therefore, the topographic loads may be partially supported by the flexural elastic rigidity of the lithosphere. This study suggests that the effective elastic thickness of Eastern equatorial margin of Brazil is 5-10 km. Subsurface load may be present in the Pará-Maranhão region.

RESUMO SOBRE A ISOSTASIA NA MARGEM EQUATORIAL BRASILEIRA. A partir da aplicação de técnicas de análise espectral cruzada a perfis de batimetria e de anomalias gravimétricas ar-livre, dispostos perpendicularmente à margem equatorial brasileira, obtiveram-se funções experimentais da resposta isostática do embasamento. Essas funções, também denominadas funções de admitância experimentais, não estão associadas a nenhum mecanismo isostático predefinido. Funções de admitância teóricas podem ser obtidas a partir de modelos isostáticos conhecidos. A comparação entre funções de admitância experimentais e teóricas mostra que, na margem equatorial brasileira, o relevo batimétrico (carga superficial) é compensado regionalmente conforme o modelo isostático flexural. A espessura elástica equivalente média da litosfera para esta margem é estimada em 5-10 km. Na região da Bacia Pará-Maranhão, esse resultado pode estar subestimado devido à provável presença de carga subsuperficiais (contrastes de densidade intra ou subcrustais).

INTRODUÇÃO The concept of isostasy is based on the mass conservation principle in which some excess mass must be compensated by a deficiency of mass. The concept of isostasy describes the manner in which some topographic feature on Earth's surface is compensated at depth (Karner & Watts 1982).

Many isostatic models are presented in the literature and they attempt to describe the manner that the compensation is achieved. The rheological behaviour of the lithosphere determines the mechanism of compensation and this behaviour is implicit in any isostatic model.

The concept of isostasy has evolved since a large number of studies (Heiskanen & Vening Meinesz 1958, Walcott 1970a,b, Watts & Cochran 1974, Banks *et al.* 1977, Banks & Swain 1978, McNutt & Menard 1978, Watts 1978, Watts *et al.* 1980, Karner & Watts *op. cit.*) which recognized that the crust has a finite strength and can support vertical shear stress.

McKenzie & Bowin (1976) applied the methods of Dorman & Lewis (1970) to obtain the response of the gravity field to a unit line bathymetric load on the sea floor from profiles of bathymetry and gravity. The advantage of this method is that a function of isostatic response of the basement, derived from observational data, can be independently obtained from a particular isostatic model. Once the experimental admittance function has been determined, it can be interpreted in terms of known isostatic models and some mechanical properties of the lithosphere may be inferred.

The isostatic response to the sedimentary deposition may explain a number of features of the stratigraphic evolution of marginal basins (Karner & Watts *op. cit.*).

The object of this paper is to extract information on the isostatic mechanism operating at the equatorial margin of Brazil by analyzing the relationship between the gravity anomalies and the bathymetry across margins. We use the same approach of Karner & Watts (*op. cit.*), who recognized

that the effective elastic thickness of the lithosphere in continental margins appears to increase with time.

GEOLOGICAL SETTING The equatorial margin of Brazil is a transform margin which resulted from the Early Cretaceous phase of shearing stress during the separation between the African and South American continental margins (Miura & Barbosa 1972, Gorini & Bryan 1976, Asmus & Guazelli 1981, Figueiredo *et al.* 1982, Zalan & Warme 1985, Szatmari *et al.* 1987).

The geology of the equatorial coastal area of Brazil is dominated by a series of Mesozoic sedimentary basin: Foz do Amazonas, Pará-Maranhão, Barreirinhas, Ceará, and Potiguar.

The most important features of the region beyond the steep continental slope are the Amazon Cone, the Ceará Rise, the Demerara and Ceará Abyssal Plains (Kumar 1978), and numerous fracture zones, including Saint Paul, Romanche, Fernando de Noronha and the North Brazilian Ridge (Cochran 1973). Also several *guyots* lie in the area between the continental slope and the North Brazilian Ridge.

Based on an extrapolated average rate of sea-floor spreading (*e.g.* Sibuet & Mascle 1978, Rabinowitz & LaBrecque 1978), oceanic accretion initiated in the equatorial margin around Late Aptian-Early Albian times; by Late Albian-Early Cenomanian times the end of the break-up was achieved (Mascle & Blarez 1987).

The Brazilian equatorial margin stratigraphy consists of clastic non-marine rift sediments and marine post-rift sediments. The rift sediments are very poorly known. Most of the drilled sections recovered Upper Cretaceous and Tertiary sediments deposited during the thermal subsidence phase of the basins in the continental margin (Cainelli & Morais 1986).

The accurate sediment thickness of the equatorial marginal basins is not very well known, but it is seismically estimated to vary from 7 to 15 km in the depocenters. The maximum thickness of sediments is located at the Amazon Cone.

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GRAVITY AND BATHYMETRY OF THE EQUATORIAL MARGIN OF BRAZIL

Fourteen gravity/topography profiles were employed in this study (Fig. 1). The gravity/bathymetry data were obtained by the Geosource Incorporated and Lamont-Doherty Geological Observatory of the Columbia University. The Geosource and Lamont-Doherty collected data with an average distance of 100 and 2,000 m, respectively. A more complete description of the data is reported in Mello (1987) and Bender (1987).

These profiles are separated into two groups: the first one (profiles A to H) lies in front of Pará and Maranhão states (Fig. 2) and the second one encompasses profiles lying from Piauí to Rio Grande do Norte states (Fig. 3). They are separated because of the general characteristics of the gravity/bathymetry profiles. The free-air gravity anomalies profiles crossing Pará-Maranhão shelf exhibit some gravity anomalies that are visually not related to the variation of the elevation of the sea-floor surface (topography). The general description of gravity/bathymetry profiles is discussed in the following items:

Figure 1 - Location of free-air gravity anomaly and topography profiles of the equatorial margin of Brazil used in this study

Pará-Maranhão region The gravity and topography profiles from this region are shown in figure 2. The light profiles (A-H) of this region, which are perpendicular to the isobaths, have lengths varying from 185 to 238 km and they are separated from each other by about 38 km.

The free-air gravity anomalies are characterized by high amplitude values, which are around 80-90 mGal. The region has a particularly wide continental shelf with a broad positive anomaly over the shelf. The bathymetry profiles of this region look very similar, but they differ from the rest of the Equatorial margin because of a wider continental shelf and a steeper slope of the Pará-Maranhão region.

Piauí-Potiguar region The figure 3 shows the six free-air and topography profiles (I-N) selected from this region. They are separated by approximately 80 km and they have about 310 km in length.

In general, the free-air gravity anomalies have amplitude less than 60 mGal (peak to peak). However, profiles L and N which run across *guyots* have free-air gravity anomalies over 100 mGal and 250 mGal respectively.

ADMITTANCE FUNCTION The admittance function theory developed by Dorman & Lewis (1970) assumes that

Figure 2 - Location of free air gravity anomaly and topography profiles of Pará-Maranhão Basin. Bathymetry is from Cochran (1973) and isobaths is in metres

the observed gravity anomalies are caused by topography and its compensation, and attempts to determine a function (admittance) that, when convolved with the topographic profile, produces the gravity response of both the topography and the isostatic compensation (McKenzie & Bowin 1976). This concept may be mathematically expressed in the Fourier (frequency) domain:

$$Z(K) = G(K)/B(K) \quad (1)$$

where: $K = 2\pi/\lambda$

K = wavenumber

λ = wavelength

Z = admittance function

G = free-air gravity anomaly function

B = bathymetry function

Figure 3 - Location of free-air gravity anomaly and topography profiles off Piauí up to Rio Grande do Norte State. Bathymetry is from Moody et. al. (1979)

The upper case symbols refer to the Fourier Transform of the corresponding space domain functions designated by lower case symbols.

EXPERIMENTAL ISOSTATIC ADMITTANCE FUNCTION The calculation of the experimental admittance function (Dorman & Lewis 1970) is done in the frequency domain by using the Fast Fourier Transform (FFT) subroutines (Cooley & Tukey 1965).

In the presence of noise, a better statistical estimation of the admittance (Z) can be obtained from the cross spectrum of gravity and bathymetry (C_r), and the power spectrum of the bathymetry (E_b). Using the approach of McKenzie & Bowin (*op. cit.*) the complex admittance is as follows:

$$Z(K_n) = C_r(K_n)/E_b(K_n) \quad (2)$$

The cross-spectrum $C_r(K_n)$ is given by:

$$C_r(K_n) = \frac{1}{N} \sum_{r=1}^N G_r(K_n) B_r^*(K_n) \quad (3)$$

The power spectrum of bathymetry E_b is:

$$E_b(K_n) = \frac{1}{N} \sum_{r=1}^N B_r(K_n) B_r^*(K_n) \quad (4)$$

and the power spectrum of the gravity is:

$$E_g(K_n) = \frac{1}{N} \sum_{r=1}^N G_r(K_n) G_r^*(K_n) \quad (5)$$

where K_n is a particular wavenumber. N is the number of profiles used. The asterisk (*) denotes the complex conjugate and G and B are the Fourier Transform of the free-air anomaly and bathymetry, respectively.

The spectral smoothing is performed by taking the average of various individual spectra from N independent determinations. Some statistical parameters are required to control the quality of the admittance determination. An unbiased estimate of the coherence (Ψ^2) is given by (Munk & Cartwright 1966):

$$\Psi^2(K_n) = \left[\frac{N \cdot \frac{C_r(K_n) \cdot C_r^*(K_n)}{E_g(K_n) \cdot E_b(K_n)} - 1}{(N-1)} \right] \quad (6)$$

The coherence (Ψ^2) is a measure of the fraction of energy in the gravity at a particular wavenumber (K_n) which can be assigned to bathymetry. The standard deviation of an admittance estimation is:

$$\sigma(K_n) = [[\Psi^2]/(2N)]^{1/2} |Z(K_n)| \quad (7)$$

Before processing, all bathymetry and gravity profiles were linearly interpolated to regular spacing points and reflected to avoid the Gibb's phenomenon. The mean of each profile was removed and the profile was truncated to a constant length using a rectangular window. The interpolation was carried out using 128 points.

THEORETICAL ISOSTATIC ADMITTANCE FUNCTION The three most commonly cited isostatic models in the literature are Airy, Pratt, and Flexure models.

In the Airy model, the density of the crust is assumed constant but the crust thickens (root) with increasing

topographic altitude and thins (anti-root) below topographic depressions. In the Pratt model, the topography is the consequence of the crustal density variation, so that mountains may be interpreted as resulting from low density crusts. The Airy and Pratt models are local isostatic models and implying that the crusts has no stiffness and can not support any vertical shear stress. In this case, the point of compensation is either directly below or directly above the point load.

The Flexure model is a regional isostatic mechanism that distributes deformation laterally around the topographic feature as well as vertically beneath it, implying that shear stresses can be transmitted horizontally.

In the case of continental margins, the more appropriate isostatic models are the Airy and Flexure models (Kerner & Watts *op. cit.*).

The development of cross-spectral techniques for analyzing the relationship between gravity and topography has led to the representation of isostatic processes in terms of theoretical mathematical filters; such as the admittance functions (Karner *op. cit.*).

The construction of theoretical admittance functions is fully discussed by Karner (*op. cit.*) and in this paper we present a summary of this theory.

The Green's equivalent layer theorem states that the gravity effect of an unknown mass distribution below a given horizontal plane may be equivalent to the gravity effect of a surface density distribution on that plane (Grant & West 1965). By the repeated use of the Green's equivalent layer theorem it is possible to construct theoretical admittance functions (Karner & Watts *op. cit.*).

The gravity effect of a unit line topographic surface $h(x)$ with fixed density ρ , at a depth $z = d$ from the reference plane ($z = 0$), in the frequency domain, is given by Parker (1972):

$$G(K) = 2\pi G \rho \exp(-Kd) \sum_{n=1}^{\infty} K^{(n-1)} F[h^n(x)]/n! \quad (8)$$

G = gravitational constant

n = is the power of the topographic function

F = is the symbol of Fourier transform

z is positive downwards

This equation is non-linear but we will use the linear component of the equation ($n=1$); this approximation is acceptable for $h(x) \gg d$ (Parker *op. cit.*).

The gravity effect of the bathymetric surface is:

$$G(K) = 2\pi G \Delta \rho_1 \exp(-Kd) B(K) \quad (9)$$

where $\Delta \rho_1$ is the density contrast between the water and the topographic load; d is the mean elevation of the topography and $B(K)$ the Fourier Transform of the bathymetric surface.

The gravity effect of the root caused by the emplacement of the topographic load is:

$$G_r = 2\pi G \Delta \rho_2 \exp[-K(d+t)] R(K) \quad (10)$$

where $\Delta \rho_2$ is the density contrast between the crust material; t is the distance between the mean surface bathymetry and the mean depth of its respective root; $R(K)$ is the Fourier Transform of the root $r(x)$.

The resultant gravity effect of the interaction of the bathymetric surface and the surface of compensation will be the destructive interference among their respective gravity effects [$G_b(K)$ and $G_r(K)$]. The general gravity effect for these two interfaces is:

$$G_f(K) = 2\pi G \Delta \rho_1 \exp(-Kd) B(K) \times \left[1 - \frac{\Delta \rho_2 R(K)}{\Delta \rho_1 B(K)} \exp(-Kd) \right] \quad (11)$$

The admittance function is defined as the ratio between the gravity anomaly and the bathymetry (McKenzie & Bowin *op. cit.*):

$$Z(K) = G(K)/B(K) \quad (12)$$

The general admittance function equation is:

$$Z(K) = 2\pi G \Delta \rho_1 \exp(-Kd) \left[1 - \frac{\Delta \rho_2 R(K)}{\Delta \rho_1 B(K)} \exp(-Kd) \right] \quad (13)$$

The ratio between the root and the topography is given by the isostatic models. In the flexural isostatic models the deflexion $r(x)$ is determined by the solution of a fourth-order elastic differential equation (Banks *et al.* 1977, Hetenyi 1946, Turcotte & Schubert 1982):

$$D \frac{\delta^4}{\delta x^4} r(x) + P \frac{\delta^2}{\delta x^2} r(x) + \Delta \rho_3 g r(x) = \Delta \rho_1 e b(x) \quad (14)$$

where D is the flexural rigidity:

$$D = \frac{E T_e^3}{12(1-\nu^2)} \quad (15)$$

E is the Young's modulus. T_e is the elastic thickness of the plate; ν is the Poisson's ratio; P is in-plane stress which is considered here equal to zero; and $\Delta \rho_3$ is the density contrast between the mantle material and the material infilling the deflection (sediments).

The equation (14) may be solved by one-dimensional Fourier Transform if we set D as spatially constant flexural rigidity (Banks *et al. op. cit.*):

$$R(K) = \frac{\Delta \rho_1}{\Delta \rho_3} \left[1 + \frac{K D^4}{\Delta \rho_3 g} \right]^{-1} H(K) \quad (16)$$

where g is the gravitational acceleration. When the lithosphere has no strength ($D = 0$) and the density of the sediments approaches the crustal density the root is the same of the Airy model:

$$R(K) = \frac{\Delta \rho_1}{\Delta \rho_3} H(K) \quad (17)$$

The Airy Admittance may be obtained by substituting (17) in (13).

$$Z_A(K) = 2\pi G \Delta \rho_1 \exp(-Kd) [1 - \exp(-Kd)] \quad (18)$$

Similarly, the flexure admittance is determined by substituting (16) in (13):

$$Z_F(K) = 2\pi G \Delta \rho_2 \exp(-Kd) \times \left[1 - \frac{\Delta \rho_2}{\Delta \rho_3} \left[1 + \frac{K D^4}{\Delta \rho_3 g} \right]^{-1} \exp(-Kd) \right] \quad (19)$$

Following the approach of McKenzie & Bowin (1976), we estimated both the superficial sediment density ρ_2 and the mean depth of bathymetry (d) based on the experimental admittance functions. The estimation was obtained by linear regression between the $\log_{10}[Z(k)]$ and the wavenumbers (K_n). These linear regressions are represented by the dashed lines in the figures 4a and 5a on the experimental admittance functions. The estimated superficial sediment density ρ_2 and the mean depth of bathymetry are 2.38 g/cm³ and 1.58 km for

Figure 4 - a. Observed admittance function of Pará-Maranhão region. The dashed line represents a linear regression fit to estimate ρ_2 and d. b. Coherence of the admittance function. Note its high values for wavenumber < 0.2

Pará-Maranhão region and 2.30 g/cm³ and 2.60 km for Piauí-Potiguar region, respectively. Using these estimated values for ρ_2 and d and varying the depth of compensation and the effective elastic thickness (EET), we constructed theoretical admittance functions for Airy and Flexure Isostatic models. For the first one, we used the EET equal to zero, the estimated parameters ρ and d , and the following depth of the compensation: 10, 20, 30, 50, 100, and ∞ km (Figs. 6a and 7a). For the Flexure model, we used the ρ_2 and d values, the mean crust thickness of 18 km and the following EET values: 1, 5, 10, 20, 40, and ∞ km (Figs. 6b and 7b).

For both models, either an infinite depth of compensation or an infinite EET is equivalent to computing only the gravity effect of the bathymetry. In doing so, any wavelength of the bathymetry would not be compensated.

DISCUSSION Two experimental admittance functions for profiles A-H (Pará-Maranhão region) and I-N (Piauí-Potiguar region) are presented in figures 4a and 5a respectively. Note that the wavelengths in the admittance function from Pará-Maranhão region have coherence greater than 0.5 (Fig. 4b). This indicates that more than 50% of the energy of the gravity signal can be assigned to the gravity effect of the bathymetry. The coherence is particularly high

Figure 5 – a. Observed admittance function of Piauí-Potiguar region. The dashed line represents a linear regression fit to estimate p_2 and d . b. Coherence of the admittance function. Note its high values for wavenumber < 0.2

Figure 6 – Observed admittance values (solid circles) estimated for the Pará-Maranhão region. The vertical bars represent the standard deviation. The solid curves were constructed based on the Airy model (A) and the Flexure model (B). T_c = depth of compensation (km). T_e = effective elastic thickness (km)

Figure 7 – Observed admittance values (solid circles) estimated for the Piauí-Potiguar region. The vertical bars represent the standard deviation. The solid curves were constructed based on the Airy model (A) and the Flexure model (B). T_c = depth of compensation (km). T_e = effective elastic thickness (km)

Figure 8 – Plot of observed free-air gravity and bathymetry profiles of the Pará-Maranhão region. Also displayed is the predicted free air gravity anomaly (or filtered topography) and the “difference” profile which represents the deviation of the predicted gravity anomalies from the observed ones.

Figure 9 – Plot of observed free-air gravity and bathymetry profiles of the Piauí-Potiguar region. Also displayed is the predicted free-air gravity (or filtered topography) and “difference” profiles which represents the deviation of the predicted gravity anomalies from the observed ones

Figure 10 – Power spectra of gravity (a) and topography (b) obtained from an ensemble average of profiles of Piauí-Potiguar region

Figure 11 – Power spectra of gravity (a) and topography (b) obtained from an ensemble average of profiles of Pará-Maranhão region

Figure 12 – Synthetic stratigraphic section of Potiguar Basin

Figure 13 – Observed stratigraphic section of Potiguar Basin corresponding to the modelled section of figure 12

Figure 14 – Synthetic stratigraphic section of Potiguar Basin constructed with constant sea-level through time. The onlap pattern in the coastal plain area was caused by increasing elastic thickness of the lithosphere with time (EET from 0 to 42 km)

(nearly 0.85) in the range of 30-100 km. With respect to the admittance function of Piauí-Potiguar region the coherence is higher (nearly 1.0) for wavelengths greater than 30 km (Fig. 5b).

Pará-Maranhão Region One way of testing the validity of the experimental admittance function is convolving it with the bathymetry. The result is a predicted gravity profile which can be compared with the observed one. The difference among them in Pará-Maranhão region (Fig. 8) and in the Piauí-Potiguar region (Fig. 9) can be also expressed in a

statistical sense (root mean squared-RMS). Note that the experimental filter does not contain any information about the positive anomaly on the middle of the continental shelf because the cross-spectrum technique vanishes the gravity information that is not related to the bathymetry. Thus, the positive anomaly which can not be predicted by the experimental isostatic response function is responsible for the great RMS values.

Figure 6 displays the observed and theoretical admittance functions for the Airy model (continuous line in the figure 6a). The diagnostic waveband, which is the rollover band interval of the functions suggests that the best fit occurs for a depth of compensation deeper than 30 km, which would be an overestimation compared to the average crustal thickness of 20 km determined by using subsidence analysis (Chang & Kowsmann in Cainelli *et al.* 1986). So the Airy model fails to explain the isostatic state of the western part of the equatorial margin (Pará-Maranhão region).

Now, examining at the diagnostic waveband of the Flexure model (Fig. 6b), the best fit seems to be for an EET in the range of 5-10 km. It is important to mention the reasons of the mismatch between the theoretical admittance and the experimental admittance. The mismatch, which occurs for wavelengths less than 70 km, is related to the overestimation of the mean depth of bathymetry. On the other hand, for very long wavelengths the mismatch is associated with the large positive anomaly over the middle of the Ilha de Santana Platform.

Most of the power spectrum of the gravity signal (Figs. 10 and 11) is shared between two positive anomalies: one in the shelf break and the other in the platform. This is a special situation because, in general, the gravity signal in a passive margin is mainly controlled by the shelf break (Karner & Watts *op. cit.*). Although most of the gravity energy is in large wavelength (Fig. 11), the experimental admittance energy is relatively low for wavelength larger than 150 km. This means that there could be an "overcompensation" of the topographic features with wavelength between 150 and 300 km. The theoretical admittance function of the flexural isostatic model, which considers only the surface loads fails also to explain the experimental admittance function of this region. We believe that this "overcompensation" is caused by subsurface loads. Therefore, a further work including subsurface load in the theoretical admittance function is recommended.

The presence of subsurface loads in this region may have important geological implication on the interpretation of the formation and evolution of the western equatorial margin. Unfortunately, there is no deep crustal and basinal geophysical data available in this area to confirm the subsurface load hypothesis.

Piauí-Potiguar Region Following the same procedure as above, we tested the validity of the observed admittance function by means of a comparison between the observed gravimetry profiles and the predicted ones (Fig. 9). Most of the predicted profiles are very close to the observed ones.

Figure 7a displays theoretical curves which were obtained according to the Airy model. The crustal thickness required to explain the observed admittance function should be thicker than 25 km. However, as determined by tectonic subsidence analysis the average crustal thickness should not be greater than 18-20 km (Mello 1987). Therefore, the Airy model seems unlikely.

Finally, comparing the experimental admittance function with those functions of the Flexure model (Fig. 7b) we obtain a good fit where the EETs are in the range 5-10 km.

Note that profile N shows the greatest difference (RMS) between the predicted and observed gravity signals. The difference is related to tertiary *guyots* associated to volcanism along transform faults.

The lithosphere deflects if a load is applied on the Earth's surface. This deformation may be predicted by assuming that the whole lithosphere is replaced by an equivalent elastic thin plate with the aforementioned thickness EET. Nevertheless this does not imply that the lithosphere behaves elastically.

Following the rationale of Watts (*op. cit.*), which says that effective elastic thickness of the lithosphere increases with its thermal age, we understand the EET associated to our experimental admittance functions as a weighted average of EETs at the corresponding time of deposition of each discrete sedimentary loads since the formation of the marginal basins. The thick column of sediments deposited during the rift phase probably loaded a low rigidity lithosphere, hence these low rigidity values had an important "weight" in the average. Moreover, during the post-rift phase the sediments were deposited on a lithosphere with progressively greater flexural rigidity, increasing the average EET. The combination of both phases resulted in a low EET which is similar to that observed in mid-ocean ridges.

IMPLICATION OF THE EFFECTIVE ELASTIC THICKNESS FOR THE STRATIGRAPHIC RECORD

We are able to predict stratigraphic features during the evolution of the equatorial marginal basins by assuming that the flexural strength of the lithosphere increases with time and that there was flexural isostatic equilibrium during development of the sedimentary basins (Fig. 12). As observed by Karner & Watts (1982), the style of basement flexure changes with time resulting in the progressive widening of the basin and in onlap pattern of the coastal plain.

Figure 12 shows a synthetic stratigraphic section for the Potiguar Basin (Mello 1987). The corresponding observed stratigraphic section is displayed in figure 13. Among many stratigraphic features present in the section, the onlap formed by the Upper Cretaceous sedimentary rocks is focused. The discussion of all variables involved in the forward stratigraphy modelling is beyond the scope of this paper and it is fully described by Mello (*op. cit.*).

Figure 14 exhibits a synthetic stratigraphic section, in which the coastal onlap has been modelled by considering only the changes of the mechanical properties of the lithosphere. Therefore, no sea-level variation were considered. This figure suggests that event without sea-level rise the onlap pattern is achieved by increasing the EET of the lithosphere with time. Following Watts (1978) and Karner & Watts (1982), the manner in which the EET evolves is related to the depth evolution of an isotherm in the range between 300 and 600°C. The magnitude of the coastal plain onlap is also controlled by the amplitude of the sea-level changes. The stratigraphic effect of sea-level variations is added to the onlap pattern of figure 14. If we compare figures 12 and 13 with 14 we can see that more than 50% of the onlap magnitude of the Potiguar Upper Cretaceous section may be assigned to the flexural compensation of the sedimentary load.

CONCLUSIONS This study suggests that the character of the free-air gravity anomaly profiles across the equatorial margin of Brazil can be explained by the Flexural Isostatic model with an average effective elastic thickness (EET) of 5-10 km. Subsurface loads may be present in the Pará-Maranhão region.

We can infer that the observed onlaps in the equatorial marginal basins were formed by interacting sea-level variation and the increase of the basement deflection wavelength due to the increasing flexural strength with age.

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Mesa Redonda:

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Debate Geral:

"Definição de um Novo Modelo para o Curso de Especialização em Ensino de Geociências".

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