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Explaining the thick crust in Paraná basin, Brazil, with satellite GOCE gravity observations

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A B S T R A C T

Seismologic observations in the last decades have shown that the crustal thickness in Paraná basin locally is over 40 km thick, which is a greater value than expected by the simple isostatic model considering the topographic load. The goal of this work is to explain this apparent discrepancy by modeling the internal crustal density anomalies through the gravity field. We use the latest Earth Gravity Model derived from the observations of the GOCE satellite mission, to retrieve the gravity anomaly and correct it for topographic effects, thus obtaining the Bouguer field. We then model the gravity effect of known stratigraphic units and of the seismological crustal thickness. The large Paraná basin comprises over 3500 m of Paleozoic sedimentary sequence with density between 2400 and 2600 kg/m^3. During the Early Cretaceous the same basin was affected by a large amount of igneous activity with a volume of over 0.1 Mkm^3. The flood basalt volcanism is known as the Serra Geral Formation, and has a maximum thickness of 1500 m. The stratigraphic units of the basin are topped by post-volcanic deposits of the Bauru Group, of about 300 m thickness, located in the northern part of the basin. The density and thickness of the sedimentary sequence are constrained by sonic logs of drill-holes and exploration seismic. We use the crustal thickness estimated from the newest seismological results for South America to calculate its gravity effect. Further we model the isostatic crustal thickness variation, allowing the comparison between a seismological Moho, an isostatic Moho, and a gravity-based Moho. We find that there is a clear positive Bouguer residual anomaly located in the northern and southern part of the Paraná basin, indicating the presence of a hidden mass, not considered up to now. We propose a model that explains this mass as magmatic rock, probably gabbro in lower crust, with density contrast of 200 kg/m^3 and thickness of more than 10 km, thus demonstrating that the flood basalt layer constitutes only a part of the melted material, the rest being emplaced into the lower crust. The presence of the magmatic material in the crust presumably has altered the thermal state, consequently changing the maturation process of the hydrocarbons in the pre-volcanic and post-volcanic rocks of the Paraná basin.

1. Introduction

The Paraná basin in Brazil is an intracratonic basin located on the stable South American platform. The cratons that surround it are the São Francisco craton to the north, the Guaporé block northwest, southward continuation of the Amazon craton. To the west the Paraná basin is bordered by Quaternary basins, the Pantanal to the north, and the Chaco-Parana to the south (Fig. 1). The Tocantins province is a mountainous area placed between the Paraná basin and the São Francisco craton, and was formed during the last orogenic cycle of Proterozoic age. The Tocantins region is the only site of present intraplate tectonic activity of the stable South American platform, with low seismicity (Assumpção et al., 2004a). During Paleozoic the basin formed in alternating marine and continental environment, forming the pre-volcanic sequence (Gama et al., 1982; Melfi et al., 1987). Starting in Early Cretaceous, intense volcanic activity formed the Serra Geral Formation, mainly composed of tholeiitic basalts and rare rhyodacytes and rhyolites (<3%; Piccirillo et al., 1987). Due to its great dimensions the basin is classified as a Large Igneous Province (LIP) (Bryan and Ernst, 2008). In the Late Cretaceous the northern part of the basin subsided and the post-volcanic rocks of the Bauru Group were deposited.

The crust–mantle interface (Moho) has been studied using earthquake seismology techniques resulting in increasing

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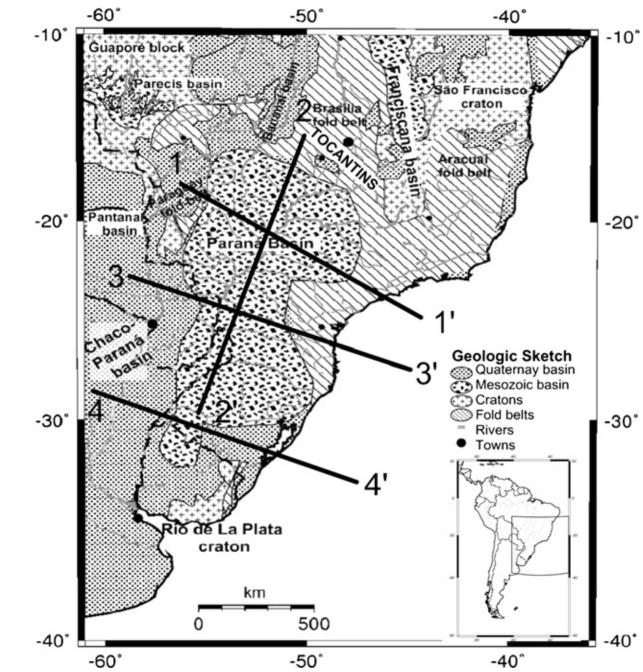


Fig. 1. Schematic map of the main geologic units. Modified after Trompette (1994), and location of three transverse and one longitudinal profile crossing the Paraná basin.

resolution in time (Assumpção et al., 1997, 2002, 2004a,b, 2006, 2013a,b; An and Assumpção, 2006; Feng et al., 2004, 2007; Lloyd et al., 2010). The authors are in general agreement that the Moho is located between 40 and 46 km.

This is interesting when we consider the Bouguer gravity field, which has moderately negative gravity values, but less than we would expect for a thickened crust underlying such a thick sediment package. In the past, the presence of a high density intrusion (Molina et al., 1987; Vidotti et al., 1998), a deep rift (Milani and Ramos, 1998), or a magmatic intrusion similar to the one found in Amazon basin (Mooney et al., 1983, 2001) were proposed. This would make the Paraná basin a rift basin, which though is incompatible with the thickened crust (Durrheim and Mooney, 1994). The increased vp/vs ratio greater than 1.83 would be compatible with crustal underplating (Christensen, 1996) and such increased value was found in some of the broadband stations placed in the Paraná basin (Julià et al., 2008). The fact that the increased vp/vs ratio is not reported (Snoke and James, 1997; Feng et al., 2004; An and Assumpção, 2006) for all stations could indicate that the underplating is discontinuous. An alternative source for the lack of negative gravity values could be in principle looked for in the mantle, in correlation with increased upper mantle velocities indicating an increase of lithospheric thickness below the Amazon craton and the São Francisco craton (Sial et al., 1999) leading to the assumption of a cratonic root below the Paraná basin (e.g. Cordani et al., 1984), but in such case the wavelength of the gravity signal should be greater.

In this work we study the density structure of the Paraná basin lithosphere, using the new data of the geodetic satellite GOCE. The GOCE mission (Gravity field and steady-state Ocean Circulation Explorer) is a dedicated ESA project (Living Planet Project) to measure the Earth gravity field with unprecedented accuracy and spatial resolution (Rummel et al., 2002, 2011; Pail et al., 2011). The spacecraft was launched on 17 March 2009, and has acquired full gradient observations at satellite height continuously, following a six-month calibration phase (Floberghagen et al., 2011). The mission is expected to terminate in fall 2013. Next to the onboard observations of the gradient, the gravity is deduced from the exact positioning of the satellite (orbit solution is near to 2 cm precision, Bock et al., 2011) and the orbit predictions through a global gravity field model, which is improved during the lifetime of the satellite. The complete global gravity potential model is a spherical harmonic expansion, built by combining the dynamics of the satellite orbits and the gradient observations (Migliaccio et al., 2011; Pail et al., 2011). The earth gravity model then allows to compute all derivatives of the gravity potential, including gradients and gravity at any height above the earth surface (Braitenberg et al., 2010, 2011a, in press; Alvarez et al., 2012; Sampietro et al., in press; Tenze et al., in press). The new gravity satellite observations, combined with the seismologic results discussed above, allow the previous works on the gravity in this area (Molina et al., 1987; Vidotti et al., 1998) to be extended. Our work consists in integrating the seismologic constraints on crustal thickness (Laske et al., 2000; Feng et al., 2007; Lloyd et al., 2010; Assumpção et al., 2013a,b) and the knowledge on the surface structures acquired from drilling and seismic investigations carried out for exploration purposes by consortium Paulipetro and Petrobras (Gama et al., 1982; Zalán et al., 1986, 1987, isopach maps reported in Melfi et al., 1987). The wells in the Paraná basin give unique information on the interrelation and distribution of the sedimentary rocks filling the basin, and were used to reconstruct the geometry of the sedimentary units. The constraints are used to define geometries and densities, and reduce the gravity values for these known structures. Assuming that these investigations define the correct geometry, a gravity residual points to density anomalies not contained in the previously published crustal model, and located either in the crust or mantle, according to the involved wavelengths of the residual gravity signal.

2. The gravity fields derived from GOCE

We use the presently available gravity model built with the observations of the satellite GOCE (Pail et al., 2011), with data coverage from 1 November 2009 to 17 April 2011. This field is preferable to the higher-resolution older global model EGM2008 (Pavlis et al., 2012). The great difference between the two datasets is, that the GOCE model is based on satellite data only, the model EGM2008 is based on both terrestrial and satellite data. The GOCE model has therefore homogeneous precision over the whole area, whereas the EGM2008 model contains possible errors or non-accessible information on the terrestrial data quality. Resolution of EGM2008 is 9 km (Pavlis et al., 2012), and resolution of GOCE is 80 km (Pail et al., 2011). The two models have been published as Stokes coefficients of the spherical harmonic expansion, with a maximum degree and order $N = 250$ for GOCE and $N = 2159$ for EGM2008. The Stokes coefficients, a table of numbers, are fed to the spherical harmonic synthesis software, for example the one published together with the EGM2008 Stokes coefficients by Pavlis et al. (2012) and available freely (http://icgem.gfz-potsdam.de/ICGEM/potato/Service.html). The software permits to calculate different physical quantities derived from the gravity potential, as geoidal undulations, gravity anomaly field, gravity gradient components at any location on or above the earth surface. The comparison of the two datasets has revealed that the EGM2008 model has errors (Braitenberg et al., 2010, 2011b; Alvarez et al., 2012), for instance in the Parecis basin (Fig. 1), due to errors in the terrestrial data entering the model (Mariani, 2012). We calculate the free air gravity anomaly at the height of 6200 m above a spherical Earth of radius 6378136.3 m, with sampling 0.5° (30′); this field was obtained with the third edition of the TIM model (Pail et al., 2011), up to degree and order 250 of the spherical harmonic expansion of the gravitational potential field. The Bouguer field is obtained calculating the mass effect of the DTM (Digital Terrain Model) ETOPO1 (Amante and Eakins, 2008) with a resolution of 1′. The topography is taken from 30° to 70°W longitude and 40°S to 10°N latitude. The DTM is discretized by tesseroids (Uieda et al., 2011), with a density of 2670 kg/m^3 on land, and 1030 kg/m^3 in ocean. The DTM is given in ellipsoidal coordinates, therefore care must be taken to reproject them to geocentric coordinates prior to calculations to conform with the coordinates used in the spherical harmonic expansion.

In Fig. 2 the free air gravity anomaly and Bouguer anomaly according to GOCE are shown. Over the continental area the two fields are comparable, the difference being much greater over the ocean, where the Bouguer anomaly increases considerably to more than 180 mGal (1 mGal = 10^{-5} m/s^2). The continental margin on the oceanic side is marked in the gravity anomaly by a string of linear positive gravity values.

3. Seismologic models of crustal thickness variation

In recent years different authors have carried out seismologic investigations that have produced a Moho crustal thickness model for the South American plate, as Assumpção et al. (2013b), Lloyd et al. (2010), Feng et al. (2007) and CRUST2.0 (Laske et al., 2000). The results of local investigations of limited extent were seen to have been superseded by the above plate-wide models for our purpose (Berrocal et al., 2004; França and Assumpção, 2004).

The CRUST2.0 is a model based on the compilation of existing data (Laske et al., 2000) with a resolution of 2° by 2°, which brakes the crust into seven layers, comprising two layers of sediments (soft and hard sediments), and three crustal layers: upper, middle and lower crust. The other three levels are topography, ice and water depth. The model uses infill data based on geological criteria where observations are missing.

The Moho model according to Feng et al. (2007) is based on the inversion of waveforms and surface wave group velocities. Next to the 3D Moho model it has produced a model of the S-wave velocities in the upper mantle. The solution uses the CRUST2.0 model as starting model and constraint, where the seismologic observations are scarce. This happens specially along the coast and offshore in the Atlantic, due to the fact that the seismic events are located along the western coast of South America below the Andes.

The crustal thickness model according to Lloyd et al. (2010) combines the point evaluation obtained from receiver functions on 20 broadband stations, group velocities and waveforms of Rayleigh waves and S-wave travel times. Also here the inversion was constrained by the CRUST2.0 model, and the coastal region was not covered by the seismic rays, and therefore reproduces the CRUST2.0 model. The data entering the inversion were 6600 waveforms from Feng et al. (2004) and 1700 waveforms from Feng et al. (2007) and Van der Lee et al. (2001) and 225 point values for crustal thickness from other authors (Assumpção et al., 2002; Feng et al., 2004; Yuan et al., 2002) and from previous works by the same authors. Furthermore some results for seismic refraction investigations were included (personal communication by S. Lloyd).

The latest model is the crustal thickness according to Assumpção et al. (2013b), that includes data from active source experiments (deep seismic reflection surveys) and receiver functions, whereas offshore the seismologic information is combined with the crustal thicknesses derived from the Bouguer gravity values according to the works of Mohriak et al. (2000) and Zalán et al. (2011). With respect to the above cited studies, the observational database has been expanded including the database used in previous works (Feng et al., 2007; Lloyd et al., 2010; Tassara and Echaurren, 2012; Pavão et al., 2012) and further data recovered from congress-proceedings and unedited monographies. When for the same station multiple results were available, the weighted average was built, weighing the uncertainties.

In Fig. 3 the crustal thickness for the most recent models according to Assumpção et al. (2013b), Lloyd et al. (2010) and CRUST2.0 is shown. The models show marked differences, which are ascribed to the different inversion techniques (Lloyd et al., 2010), due to the smallness of the teleseismic signals, and due to the presence of the basalt layer in the Paraná, that disturbs the propagation of the seismic waves. The Moho of Lloyd et al. (2010) presents the greatest amplitude of oscillation, with a more superficial Moho in Colombia, Venezuela, Bolivia, Paraguay, Uruguay, Argentina, partly in the Borborema-province of Brazil (northwest of Paraná basin), and in the central-western sector of Paraná. The Moho of Assumpção et al. (2013b) has a smaller oscillation amplitude, and shows a more shallow Moho in Paraguay (near Asunción), west of Paraná. The more recent Moho of Assumpção et al. (2013b) is flatter and deep in the Paraná basin, without the segmentation in the northern and southern basin. All models have problems along the coast and offshore due to the lacking of seismological ray paths, and consequent usage of gravimetric or isostatic considerations. We therefore do not adopt the seismological Moho along the coast and offshore, to prevent a circular analysis.

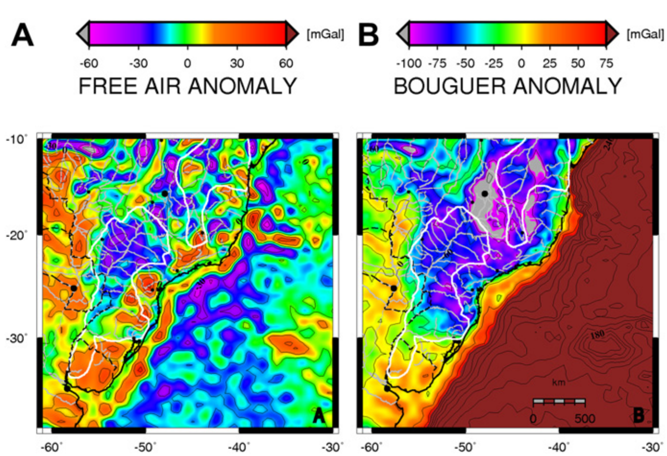


Fig. 2. Gravity observations in the study area. (A) Free air gravity anomaly and (B) Bouguer anomaly according to GOCE.

4. Gravity modeling

4.1. Upper crustal model

The main known layers covering the basement in Paraná are three, the pre-volcanic rocks, the basalt layer, and post-volcanic layer. The first level belongs to the pre-volcanic Paleozoic rocks (Gama et al., 1982, map reported in Melfi et al., 1987) where linear density changes from 2400 to 2600 kg/m^3 at 3500 m (Fig. 4A), and isopachs reach 3500 m (Silva and Vianna, 1982). The maximum thickness coincides with the depocentre of the basin. These isopachs rely on wells distributed throughout the Paraná basin, the location of which is seen in Gama et al. (1982). The second layer is composed of the basalt of the Serra Geral Formation with a constant density of 2850 kg/m^3 (Marques et al., 1984) and a thickness of about 1500 m (Fig. 4B). The position of the trunk of the drainage tree is aligned with these isopachs, as are the pre-volcanic sediment-isopachs. The top layer is located only in the northern part of the basin, and corresponds to Coniacian to Maastrichtian (89.3–65.5 Ma, Late Cretaceous) rocks of Bauru Group (Fernandes

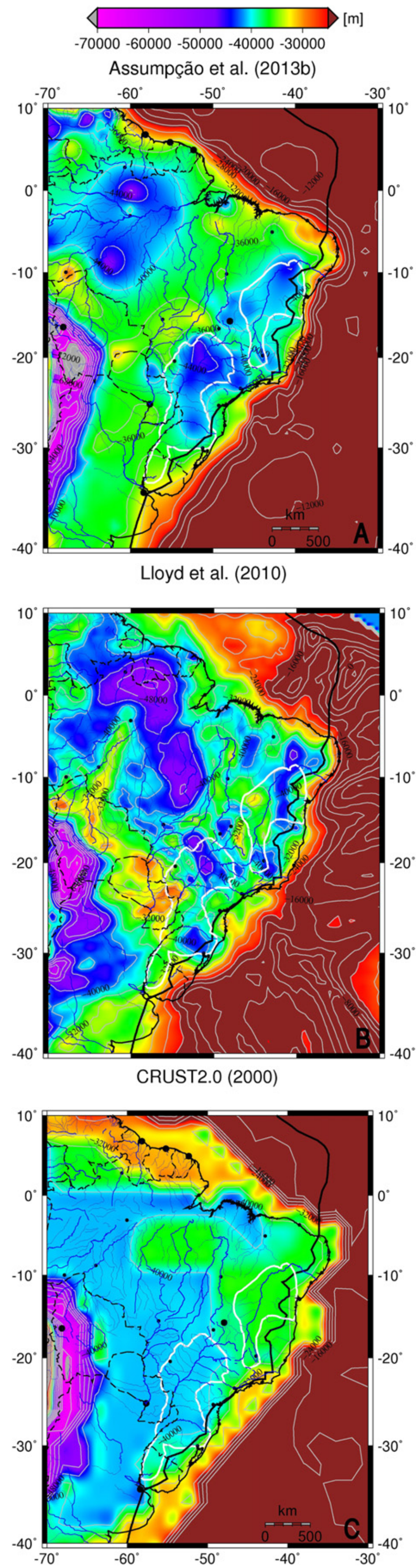


Fig. 3. Seismologic crustal thickness models for South America. (A) Model Assumpção et al. (2013b). (B) Model Lloyd et al. (2010). (C) Model CRUST2.0 (Laske et al., 2000).

et al., 2009, Fig. 4C). The latter is a small sedimentary basin of about 250 m thickness, with a constant density of 2200 kg/m^3 (Silva and Vianna, 1982).

The next step of investigation is to calculate the gravity field corresponding to the crustal model of known densities and geometries, comprising the crustal thickness and sediment layers. We define a reference crustal model, with an upper crust 20 km thick, and a lower crust reaching the normal crustal thickness of 35 km, with standard densities of 2670 and 2900 kg/m^3, respectively. The mantle has the density of 3200 kg/m^3. The reference model corresponds to a standard crustal model (IASP91, Kennett, 1991; Kennett and Engdahl, 1990). The reference model and the densities of our model are shown in Fig. 5.

The gravity effect of the three layers was calculated by discretizing the known geometries of the units into layers of varying density. Each layer is 100 m thin, with laterally varying density, for which the gravity field is calculated with a spectral field estimation (software Lithoflex: link www.lithoflex.org, Braitenberg et al., 2007). The pre-volcanic rocks give a negative contribution (Fig. 6A) of up to −20 mGal. The Serra Geral Formation contributes to a value of 11 mGal (Fig. 6B). The Bauru Group produces a smaller negative of up to −4 mGal (Fig. 6C). The composite gravity contribution of the sedimentary and volcanic rocks (Fig. 6D) reveals two sectors divided by an axis oriented NNE–SSW: the western sector has a more negative gravity effect (up to −20 mGal), the eastern one is less negative with values up to −9 mGal. This means that the positive mass of the Serra Geral Formation is smaller than the mass deficit imposed by the lighter sedimentary rocks. In terms of mass balance, this implies that with respect to the standard crustal column, the sedimentary rocks of the Paraná basin add up to a mass deficit, notwithstanding the presence of the basalt layer.

4.2. *Gravity of crustal thickness variations*

With the aim of calculating the gravity residual of the crustal model including crustal thickness variations and sedimentary layers, we proceed to estimate the gravity effect of the Moho model. We have shown that in the most recent literature apart from CRUST2.0, two refined models exist, the model according to Lloyd et al. (2010) and the model by Assumpção et al. (2013b). The model of Feng et al. (2007) has been replaced by the newer Assumpção model, having been published by the same authors using the same methodology, and extending the dataset. We therefore must proceed to consider these two models in our further modeling, and discuss the results for both. The CRUST2.0 model does not need to be considered independently, because the other studies are a refinement of this global crustal model. We use a constant density contrast across the crust–mantle interface, but allow for two end member values, −500 and −300 kg/m^3, representing the density contrast between Peridotite and Granite or Gabbro. The first corresponds to a light lower crust, the second to a denser crust. The final residuals we obtain will be interpreted as deviations from the assumption of a constant density contrast along the crust–mantle interface. We present the Bouguer fields reduced by the effect of deviations of the crustal thickness model and of the sedimentary layers from the reference model. The model of Lloyd et al. (2010) results in two localized positive anomaly residuals, one in the northern (a in Fig. 7A and Fig. 7A), another in the south-central part of the basin (b in Fig. 7A and Fig. 7A). The amplitude of the residuals is exactly proportional to the density contrast used. The model of Assumpção et al. (2013b) results in a more extended positive residual centered on the basin (Fig. 8A and B). The existence of the residual shows that the existing crustal model does not reproduce the gravity observations correctly, and that there is a

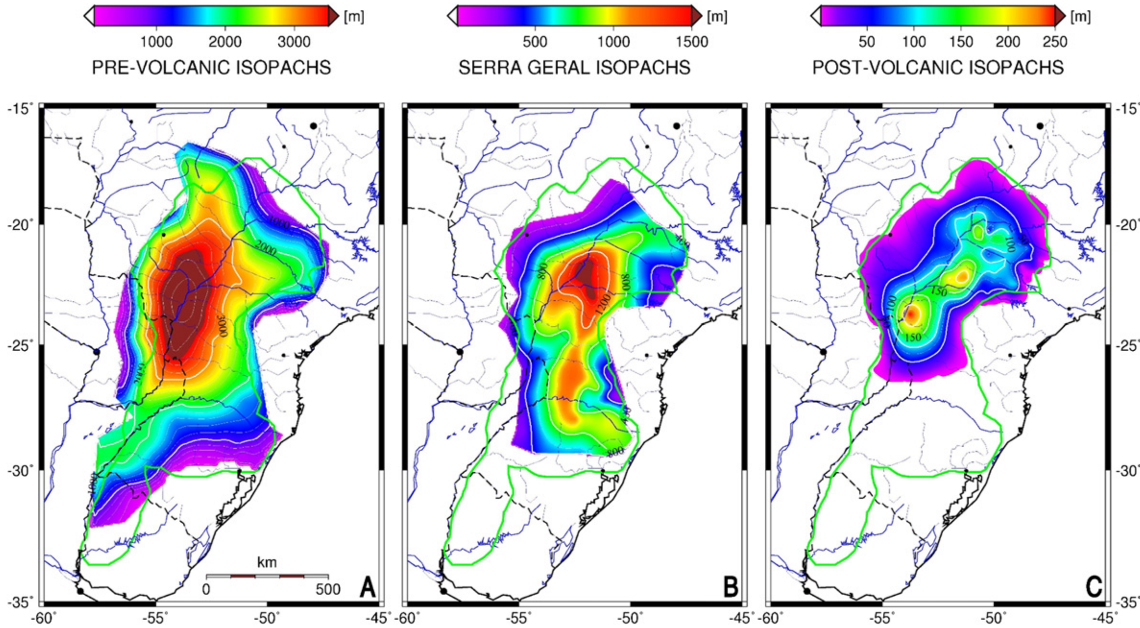


Fig. 4. Isopachs of the layers constituting the Paraná basin; the green line defines the limit of the Paraná basin. (A) Thickness of pre-volcanic rocks. (B) Isopachs of the volcanic Serra Geral formation. (C) Isopachs of the post-volcanic rocks, the Bauru Group. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

hidden mass in the crustal column. The fact that the seismologic Moho is deeper than expected from gravity is independent of the specific seismologic Moho model, and is not altered by the density contrast used across the crust–mantle interface.

4.3. Moho from gravity inversion

The gravity values corrected for the sedimentary rocks can be also used to calculate the gravity Moho, under the above assumption of a constant density contrast across the Moho interface. Therefore we use the GOCE gravity data, and apply the iterative inversion method applied successfully in different parts of the world (e.g. Zadro and Braitenberg, 1997; Ebbing et al., 2001; Braitenberg et al., 2006) and included in the Lithoflex software (Braitenberg et al., 2007). The inversion depends linearly on the density contrast across the interface, and we choose the lower value used in the previous chapter (-300 kg/m^3). The higher

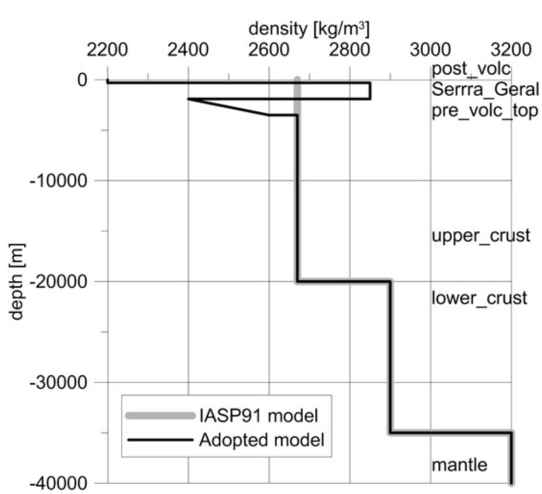


Fig. 5. Reference model (IASP91, Kennett, 1991; Kennett and Engdahl, 1990) and the densities used in the forward modeling of the crust.

density contrast produces a proportionally smaller amplitude of the Moho deviations from the standard reference depth of the Moho (35 km). The gravity Moho (Fig. 9) is deeper beneath the fold belts and shallower below the central axis of the Paraná basin, which does not comply to the seismologic results of a deep Moho below the Paraná. The discrepancy is another way of showing that there must be a dense mass at crustal level, that is located above the deep seismologic Moho, and produces the positive gravity signal that results in an apparent uplift of the Moho below the basin from the inversion process.

5. Characteristic profiles across the model

The results of our modeling are best discussed along four characteristic profiles that cut the basin (location of profiles see Fig. 1), as shown in Fig. 10A–D. The gravity and seismologic Moho, an isostatic Moho and the DTM are graphed along the profiles. The isostatic Moho has been calculated for a local compensation Airy model (elastic thickness $T_e = 0$ km) and a regional compensation model ($T_e = 30$ km), for reference using the Lithoflex software. The isostatic model is the thin plate lithospheric flexure model, which represents the bending of a homogeneous elastic thin plate overlying an inviscid mantle and loaded by the topographic and intracrustal load (e.g. Watts, 2001). The calculations are accomplished with the Lithoflex software, which uses the flexure formulation of Braitenberg et al. (2002). Standard density values are used: crust 2670 kg/m^3, water 1030 kg/m^3, mantle 3200 kg/m^3, and we reduce topographic load to equivalent topography (see Braitenberg et al., 2002). In the frame of this study the exact value of the elastic thickness is not essential, as the goal of the isostatic calculations is only to obtain an indication of the expected variation of the isostatic Moho. The T_e value could also be greater, producing a slightly smoother Moho. The T_e of 30 km is what Tassara et al. (2007) had found for the Paraná basin. The Airy case is an end member value, showing the maximum Moho oscillation in the frame of the isostatic model. Density of crust and mantle were the same as the standard reference model defined above, other values are also kept to standard, as Young modulus (10^{11} N/m^2) and

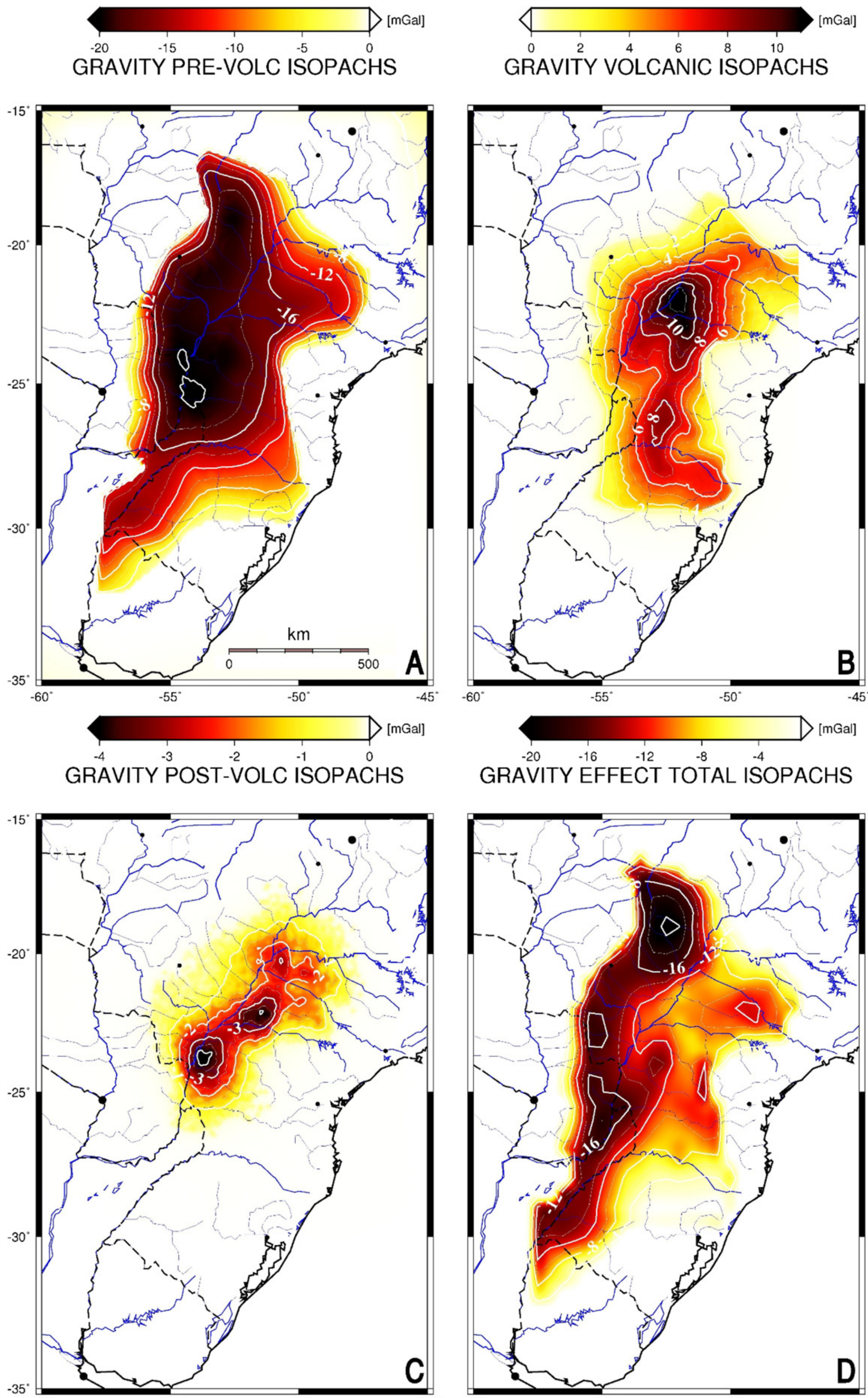


Fig. 6. Gravity effect of the sedimentary and volcanic infill of the Paraná basin. (A) Pre-volcanic rocks. (B) Basalt of Serra Geral Formation. (C) Post-volcanic Bauru Group. (D) Total effects of the Paraná layers.

Poisson ratio (0.25) (Watts, 2001; Braitenberg et al., 2002). The load entering the isostatic model includes all density variations respect to the standard density column, in our case sedimentary sequence and ocean water, and is expressed in terms of equivalent topography, as defined in Braitenberg et al. (2002). We have added the isostatic Moho to the seismologic and gravity Moho as a reference value, that tells us what the crustal thickness would be in case of isostatic equilibrium. The profiles 1–1′, 3–3′, and 4–4′ cut the basin in NW–SE direction, the profile 2–2′ cuts the basin along its longitudinal axis along the Paraná river (on Fig. 1). The gravity Moho is shown for the density value −300 kg/m^3. If gravity Moho, seismologic and isostatic Moho are all coinciding, the hypothesis of a homogenous crust is adequate. If the seismologic Moho is deeper than the gravity and isostatic Moho, it means that there is a densification in the crust, that has not been accounted for in our model. If the seismologic Moho is shallower than the gravity and isostatic Moho, it means that there is a density reduction in the crust, e.g. sediments, we have not yet modeled. The best agreement between the seismologic and gravity and isostatic Moho is found in the southern part of the basin, along profile 4–4′. CRUST2.0 is flat, and gives only a rough indication of where the Moho is. Moving to the north the disagreement between gravity–isostatic and seismologic Moho is systematic and increases, with a greater depth for the seismologic Moho compared to the gravity Moho. Along profile 1–1′, the northernmost profile, the gravity Moho and isostatic Moho are concave upwards below the Paraná basin. The seismologic Moho of Assumpção et al. (2013b) and of Lloyd et al. (2010) remain deep and do not show the Moho shallowing. The observations along the profiles show that the seismologic Moho is deeper than the gravity and isostatic Moho in the northern Paraná basin, particularly below the maximum thickness of the Serra Geral Formation. In the southern part of the basin the seismologic and gravity and isostatic Moho are in quite good agreement. This shows that the intracrustal densities must be changing in the basin from south to north. The discrepancy cannot be due to the inhomogeneities introduced by the basin, as we have corrected for them, and must be located below the basin. The longitudinal profile 2–2′ running from north to south, starting in the craton, crossing the fold belt, and entering the Paraná basin, shows that the higher topography of the fold belt produces a Moho deepening, both for gravity and the seismologic Moho; the segment in the basin would require a shallower Moho both from the isostatic and gravity modeling, whereas the seismologic Moho is even deeper than the Moho below the fold belt.

6. Underplating below flood basalt

Our analysis so far has shown that there is a difference between the northern and southern Paraná basin, in that the seismologic Moho below the northern sector is significantly deeper than expected from the gravity and isostatic considerations. In order to explain this apparent discrepancy, we must introduce a surplus

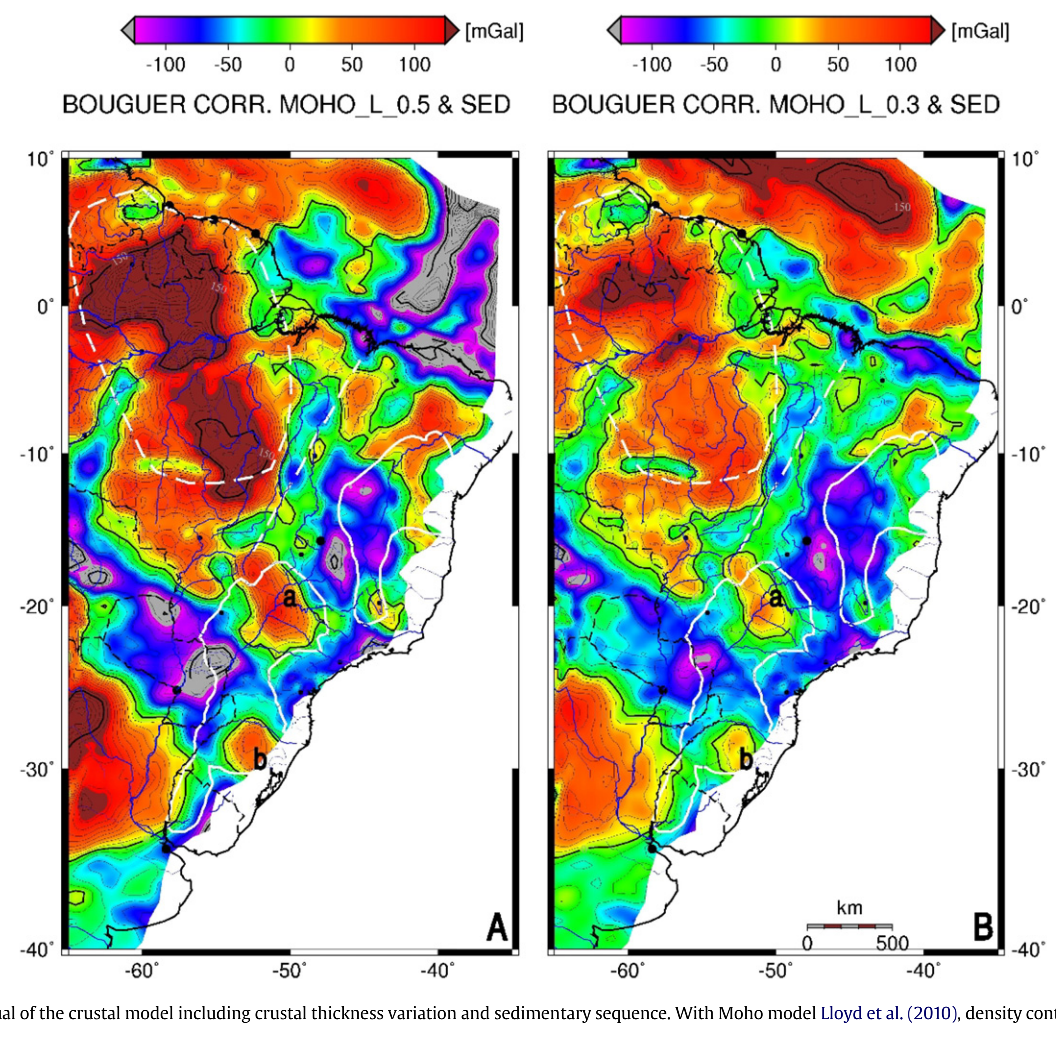


Fig. 7. Gravity residual of the crustal model including crustal thickness variation and sedimentary sequence. With Moho model Lloyd et al. (2010), density contrast: (A) −500 kg/m^3; (B) −300 kg/m^3.

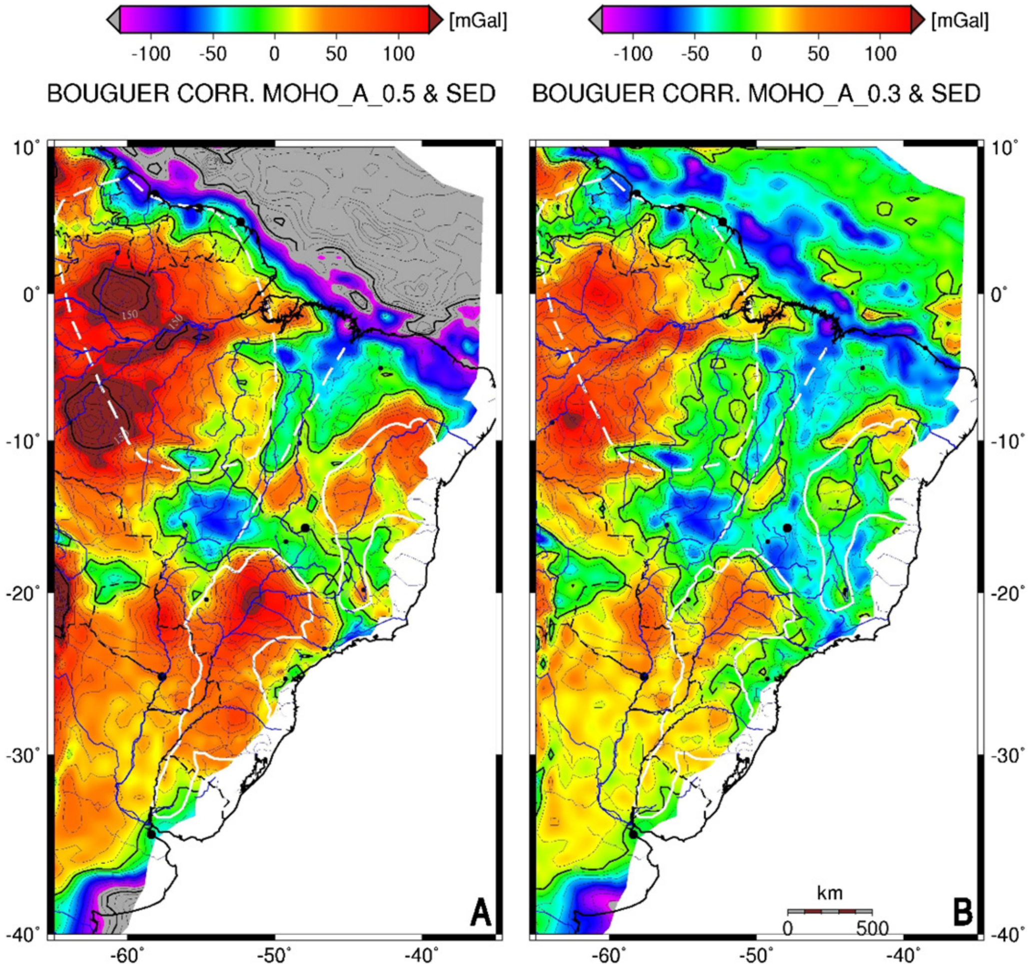


Fig. 8. Gravity residual of the crustal model including crustal thickness variation and sedimentary sequence. With Moho model Assumpção et al. (2013b), density contrast: (A) −500 kg/m^3; (B) −300 kg/m^3.

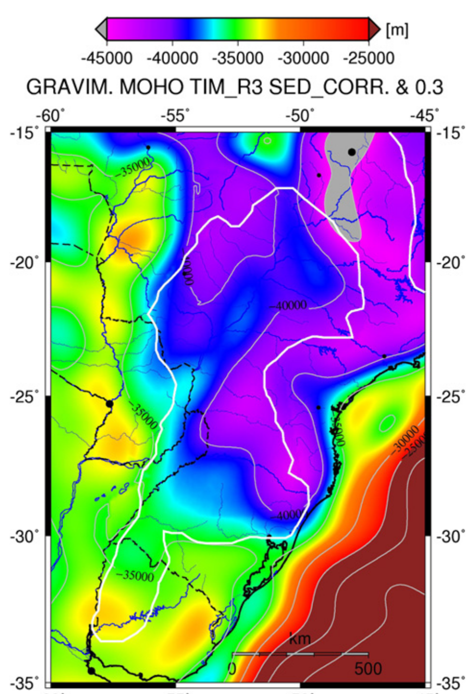


Fig. 9. Gravity Moho from a homogeneous crust−mantle interface model, density contrast −300 kg/m^3 GOCE gravity data.

mass at crustal level, below the known layers. The surplus mass has two effects: it is the source to the positive gravity residual, and it constitutes the extra-load which we need in the isostatic model to push the Moho downwards.

One possible mechanism to create an increased density crust, is to associate the hidden mass to the magmatism, in the sense that the melting basaltic material is emplaced partly as flood basalt at the top of the crust, but partly in the crust, with a density greater than 2900 kg/m^3, higher than the reference crust. The dense material was emplaced in mid-lower crust in form of sills, that enter into cracks or at the bottom of crust, increasing the total volume of the crust.

Considering the possible rock types, it is to be noted that depending on temperature and depth, there is a transition from basalt, to diabase to gabbro, the composition being the same. Basalt occurs at the surface, diabase at 2–3 km depth, and gabbro, with large crystals, in the lower crust. The densities as a function of depth for some varieties of basalts are shown in Fig. 11A, with values extracted from the tables of Christensen and Mooney (1995). The density surplus in the lower crust with respect to the standard reference density of lower crust (2900 kg/m^3) is of up to 150 kg/m^3. We consider also the velocity anomalies in the lower lithosphere, according to an update of the papers of Feng et al., (2004), 2007, and M. Assumpção (personal communication): the authors (see Feng et al., 2004) publish the percentage variation of S-wave velocity with respect to reference velocity at 30 km depth, which corresponds to the average S-wave velocities between the surface and

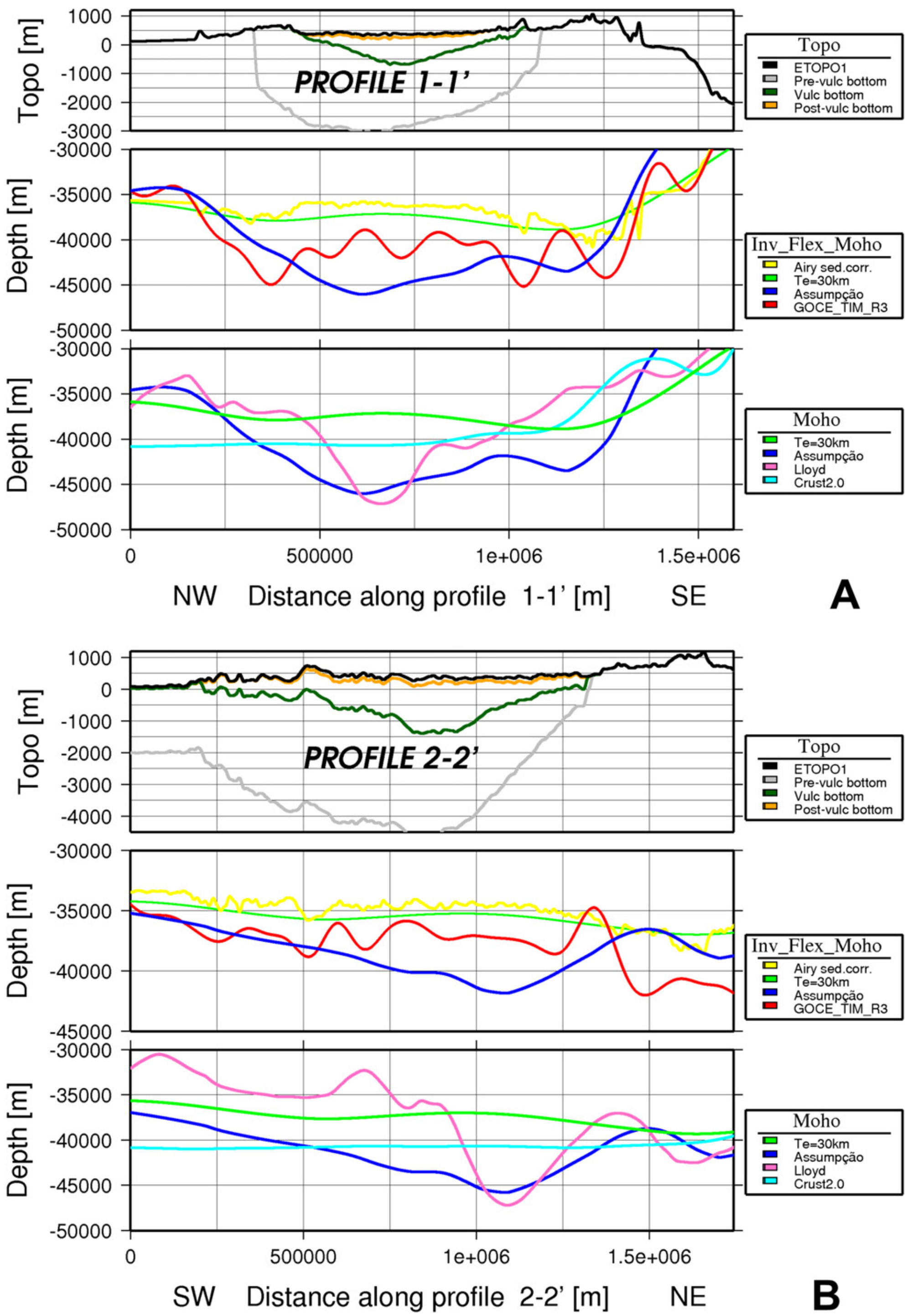
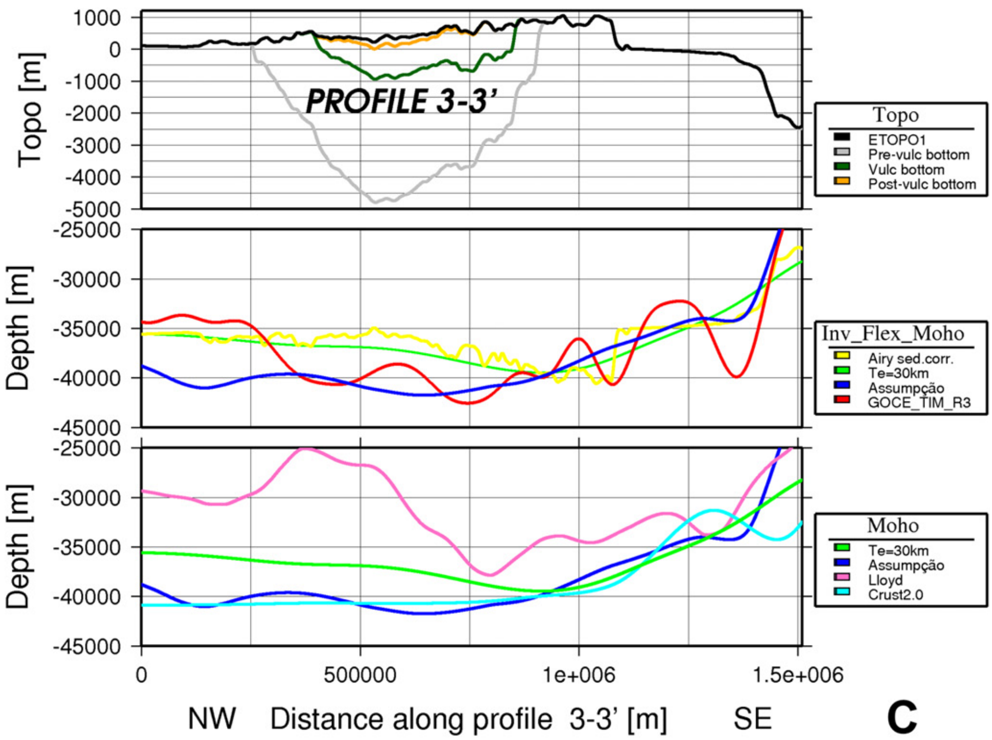


Fig. 10. 2D section along profiles illustrating DTM, seismologic Moho, gravity Moho and isostatic Moho. Fig. 1 shows the location of three transverse and one longitudinal profile crossing the basin. Beneath the Serra Geral Formation the seismologic Moho is consistently deeper than the gravity and isostatic Moho, pointing to increased density in the crust presumably related to the magmatic activity reaching the surface. (A) Profile 1–1′; northern transverse profile crossing the maximum crustal thickening in Paraná basin. (B) Profile 2–2′ longitudinal profile along Paraná basin, along axis of crustal thickening below basin. (C) Profile 3–3′ transversal profile at southern end of crustal thickening; (D) Profile 4–4′ transversal profile south of the Paraná basin, here the overthickened crust is no longer observed. The profile is further south than the presumably underplated crust.



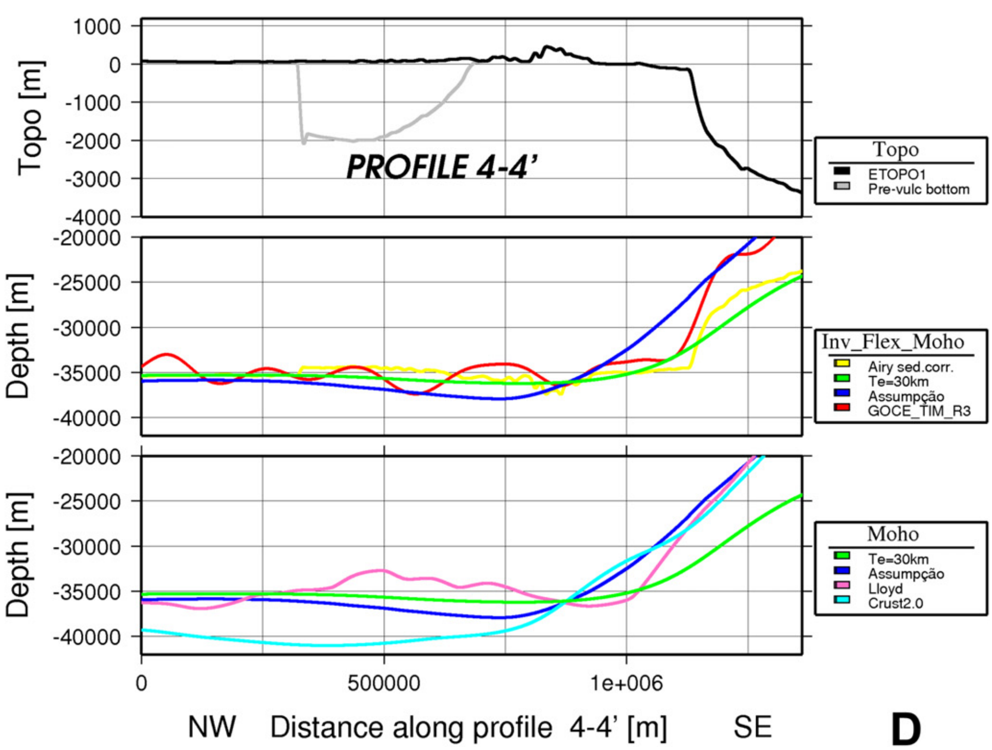


Fig. 10. (*continued*).

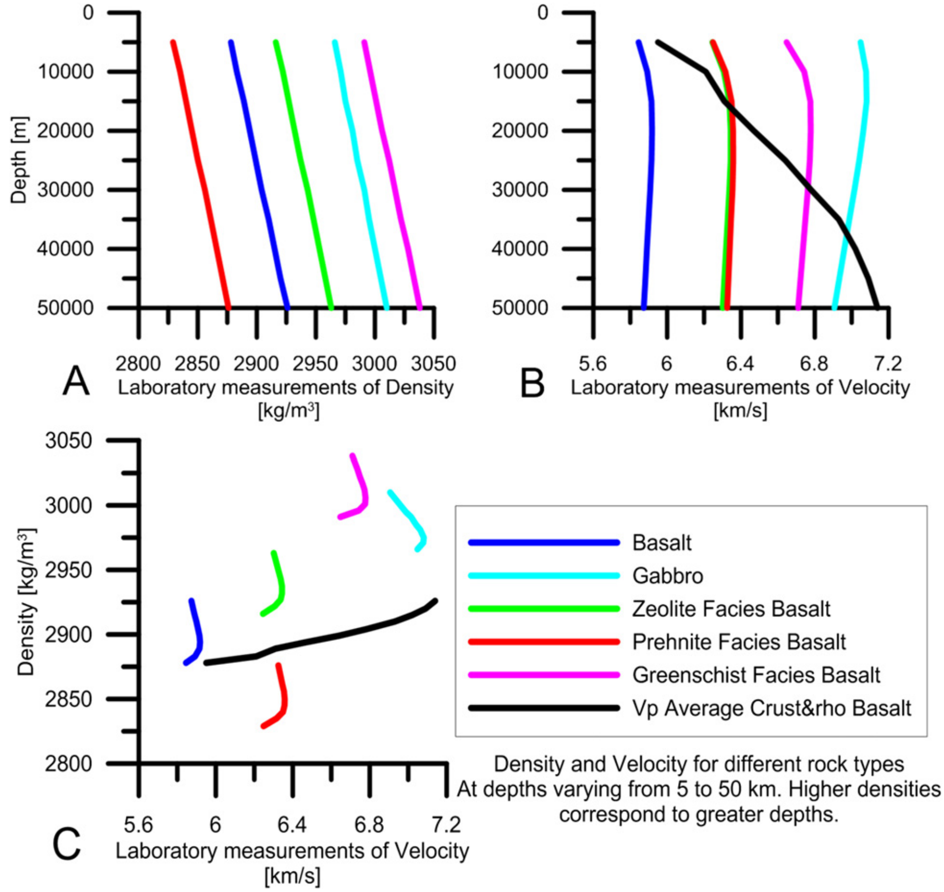


Fig. 11. Depth dependence of density and seismic velocity vp from laboratory measurements for the crust for different basalt types and gabbro (Christensen and Mooney, 1995). (A) Density in function of depth. (B) Compressional velocity in function of depth. Also shown is the velocity variation for the average crust. It shows that underplated gabbro is transparent to seismology as it has similar velocities to the average crust. (C) Density–velocity relation at varying depths. The continuous curve shows as reference the curve for basalt. Greenschist–facies–basalt and gabbro have considerably higher density than basalt.

50 km depth. It is found that the São Francisco has normal velocity, the Amazon craton has increased velocity, and the Paraná basin has reduced velocity. To the west and east of the Paraná velocity low, the velocity increases in the Chaco basin, and towards the coast, respectively.

When looking for a possible rock-type producing high density and low velocity, then the basalt would fit these requirements (Christensen and Mooney, 1995), as it is seen in Fig. 11B, where the Vp-velocity to density relation is given for different rock types. Basalt and gabbro are seen to have low velocity in the lower crust, in comparison with the average crustal velocity variation. Considering the standard density reference column, gabbro and basalt contribute a positive density contrast. This would imply that the underplated material would be seen in the gravity signal, as a positive anomaly, and in the velocity as a negative anomaly, which is what is found in the Paraná basin. In Fig. 11C the density–velocity relation for a range of depths from 5 to 50 km is shown. The values have been obtained from laboratory measurements up to 1 GPa, corresponding to about 35 km depth, and extrapolated to the depth of 50 km and refer to the table 4 of Christensen and Mooney (1995). It seems that, in general terms, both basalt and gabbro contribute to a higher density than the average crust for a given velocity value.

The considerations on the velocity anomalies would suggest the hidden mass to be located in the mid-lower crust, rather than in the upper crust. Assuming a density contrast of 200 kg/m^3, we estimate the thickness of the underplated body by inverting the gravity residual. In Fig. 12A the location of the 2D profiles analyzed in Fig. 12B and C is shown. The model assumes the reference depth of the body to be located at 20, 30 or 40 km, and the inversion determines the geometry of the body, given the density contrast. The reference depth defines the top of the body. We find the total thickness of the body to be over 10 km (Fig. 12); the deeper the body is assumed to be, the bigger its mass must be to explain the gravity residual. If the density of the underplated material is greater, then the thickness of the body is proportionally smaller, as can be seen in Fig. 12B, where a density contrast of 300 kg/m^3 was used to illustrate the effect of a varying density. A density contrast between 100 kg/m^3 and 200 kg/m^3 is to be expected when considering the density of gabbro (see Fig. 11) and the density of the normal lower crust lower (2900 kg/m^3).

7. Discussion

Our work aims at the full use of the new gravity observations derived from the satellite GOCE in obtaining a better understanding of the Paraná basin and its surroundings. The Paraná basin is known in its superficial parts since the 1960–70s from exploration drills and seismics, that defined the maximum depth of the basin at 5600 m (Zalán et al., 1987). The knowledge of the deeper crustal structure, as the crustal thickness, has been known since recent years from seismologic investigations (Feng et al., 2007; Lloyd et al., 2010; Assumpção et al., 2013b). Already earlier works concerned with the gravity field and the isostatic equilibrium had encountered

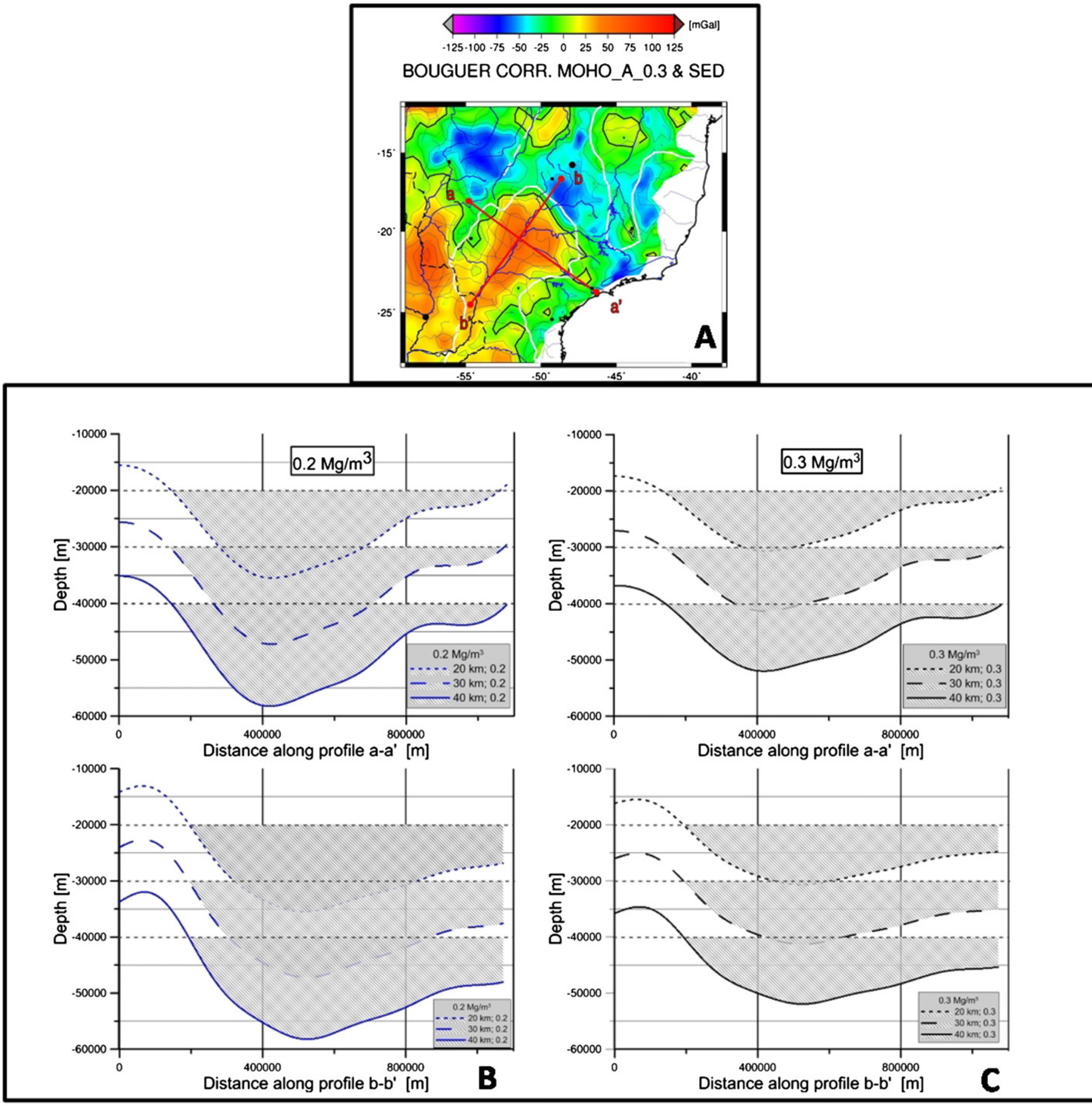


Fig. 12. Geometry of the high density body in lower crust along the two profiles a–a′ and b–b′ from gravity inversion at different reference depths (20, 30, 40 km). (A) Location of profiles; (B) Inversion with density contrast 200 kg/m^3 along profiles a–a′ and b–b′; (C) Inversion with density contrast 300 kg/m^3 along profiles a–a′ and b–b′.

some evidence of extra-mass to explain the gravity field in the Paraná basin (Molina et al., 1987; Vidotti et al., 1998). The new gravity data from satellite GOCE and the new seismologic investigations permitted us to analyze the crustal structure in more detail and to address the question of the isostatic equilibrium. This study provides a step forward in defining the properties of the crust below this Large Igneous Province (LIP).

We have undertaken two lines of investigations: the first considers the gravity field, the second the isostatic state, and both point to the fact that there must be surplus mass at crustal level as already proposed by Molina et al. (1987) and Piccirillo et al. (1987), however at the time, no Moho depth estimates were available. The correction of the gravity effect of the two known sedimentary sequences and the basalts, has shown that the lighter sedimentary rocks have a greater contribution to gravity with respect to the basalt layer, and therefore the net effect is negative. This means that notwithstanding the basalts of the Paraná belong to a LIP, the mass of the pre-volcanic and post-volcanic rocks add up to a greater value than the basalt layer. In terms of the second line of investigation, the isostatic response of the crust, the sedimentary rocks therefore contribute to a load deficit, and we would expect the crustal thickness to be reduced with respect to the reference crust, which is 35 km thick. Tassara et al. (2007) found an elastic thickness value near to 30 km for the Paraná, which shows that the crust is expected to respond in thickness to the load. Only for a very high elastic thickness like 100 km, does the crust not respond to loads and has a flat Moho. But instead of a reduced thickness, the seismological results discussed above agree in a thickened crust of up to 46 km, way over the value of a standard crustal thickness. Considering our two lines of investigations, we have estimated the gravity effect of the thickened crust, which is a negative gravity contribution, which adds up to the negative contribution of the sedimentary sequence. Thus the summed effect of the crustal thickness and the sedimentary rocks is a negative gravity effect, whereas the observed Bouguer gravity is positive. This apparent discrepancy means that there is a surplus mass at crustal level that compensates for the negative effect of sedimentary layers and thickened crust. Our second line of investigation, the isostasy, also shows a discrepancy between the expected uplifted Moho and the observed deep Moho; only a surplus mass at crustal level can give the load that is needed to counteract the load deficit created by the sedimentary rocks.

Our observations are based on the integration of two pieces of information: the sediment layer thickness including the basalt layer, and the seismologic results on crustal thickness.

The sedimentary sequences have been controlled by exploration drilling, so there is good control on their minimum depth. Part of the basalt layer has been eroded (Bryan and Ernst, 2008; Gomes et al., 2011) due to uplift of the area, but nonetheless we are considering the present equilibrium, so the mass we need to explain the observations must be there today.

The seismologic investigations have several problems in the eastern South American continent due to the unsymmetric distribution of earthquakes, which are mostly from the Pacific side of the continent, the Atlantic side being near to aseismic. The different methodologies used in the studies result in differences in the crustal thicknesses, but all publications agree in a thick crust of 40–46 km in the Paraná area (Assumpção et al., 2002, 2004a, 2006; An and Assumpção, 2006; Feng et al., 2004, 2007; Julià et al., 2008; Lloyd et al., 2010).

In northern Italy, in the Ivrea-Verbano area, a crustal section which was below a large volcanic source is exposed, revealing a large magmatic complex extending in the lower crust with amphibole gabbro and gabbro that form an underplated body that was incorporated into the metamorphic crust (Quick et al., 2009). Laboratory measurements of seismic velocities of rock samples were fed into a simulation that produced a synthetic seismic section (Rutter et al., 1999) which showed that the underplated gabbro would be transparent to the seismologic investigations, and the Moho would be retrieved correctly below the underplated material. Translated to the Paraná basin, it follows that the lowest velocity reflector picked by the seismologic observations is the Moho, and that the hidden mass we need to explain the gravity observations is located above the reflector, and not below it. The underplating has been observed in some seismologic stations showing increased vp/vs ratio greater than 1.83 (Assumpção et al., 2004b; Julià et al., 2008).

The presence of a dense mid to lower crust below the Paraná basalts is conform to the petrologic models that postulate the presence of dykes and sills in the northern part of the basin. Superficial dykes are observed in several areas of the basin like in Ponta Grossa (Ernesto et al., 2002). The regional distribution of the sills has been postulated to be more frequent in the central and northern part of the basin (Melfi et al., 1987). Moreover the petrologic models predict the presence of dense magmatic products in the lower crust that were left behind by the differentiation process that has produced the basalts (Piccirillo et al., 1987). Generally, petrologic models predict that the basalts of a LIP found at the surface are only a small part of the melting process, estimated in a fraction of one-tenth of the total volume magmatic crustal material (Bryan and Ernst, 2008). The missing mass we must place in the crust would be this melted crustal material that has a greater density than the reference crust.

Analogies can be found for the rhyolites of Karoo, that have been associated to underplating and the observations in Valsesia (Italy), where lower crust of a large volcanic system, the so-called supervolcano is exposed and shows evident gabbroic underplating (Voshage et al., 1990; Quick et al., 2009). Another LIP associated to one of the largest sedimentary basins are the Siberian traps, that continue as a basalt layer below the West Siberian basin. Analogous calculations as the one done in the Paraná, had shown that the lower crust must be dense, in order to explain the gravity values, the deep Moho and the great depth of the sedimentary basin (Braitenberg et al., 2004).

8. Conclusions

We use the new gravity observations of the GOCE satellite, that guarantee homogeneity of the precision, due to the fact that they are independent from the availability of terrestrial data. The spatial resolution of the GOCE observations (80 km) is sufficient for the scale of our problem, that covers the entire Paraná basin. Although different models exist, for the different publications, they all agree on the thickened crust below the northern Paraná basin. We build a crustal density model based on geophysical and geological constraints on geometry and density.

The recent seismological studies have shown that the crust–mantle interface below the Paraná basin is deep, and the deepest values are found in the northern part of the basin, depending on the specific model between 40 and 46 km. The proposed crustal model in the present paper is based on gravity inversion and on the isostatic flexure model. The gravity model is based on the inversion of the Bouguer values, assuming a constant density contrast across the discontinuity. The isostatic model uses parameters on elastic thickness and the intracrustal and topographic loads. The gravity model and the isostatic model predict a relatively shallow Moho, with depths near to 35 km below northern Paraná basin. The isostatic and gravity Moho are therefore shallower than the seismologic Moho. A thick crust, as the one predicted by seismology, generally generates a strongly negative Bouguer anomaly, which we do not find for the Paraná basin. There is a relative Bouguer anomaly high along the maximum sediment accumulation which follows the NE–SW trend of the Paraná river drainage basin Paraná, instead of the Bouguer minimum expected for such a deep Moho.

The volcanic deposits of the basin are too thin to explain the relative gravity high, the source of which has to be seeked at lower crustal levels, which adds to the mass needed to explain the lack of negative Bouguer anomaly expected for such a thick crust. We propose that the anomalously thick crust is related to the basalt volcanism, and is presumably a general phenomenon, not limited to the Paraná basin. The densified lower crust would be interpreted as the magmatic products that were left behind by the ascending basalt that produced the flood basalt. The magmatic material can be interpreted as magmatic underplating at the lower crustal boundary, or also as sills which intruded the crust. Both phenomena contributed to increase the thickness of the crust with rocks of higher density than normal average crust.

The underplated or intruded material could have been the mechanism driving the subsidence of the Bauru Group (Late Cretaceous) due to a crustal overloading and cooling. The Bauru deposition though did not happen immediately, as there is a time interval between magma emplacement and Bauru deposition

(∼40 Ma). If the intrusion (at 133 Ma) controlled the Bauru deposits, then, the age of this group should be immediately after. Nonetheless, the underplating could have generated the space and topographic depression, then, later, another erosional process of the surrounding uplifted areas carried the Bauru sedimentary layer into the basin.

Petrographic evidences in favour of our hypothesis are the lack of primitive magma on the surface, and the chemistry of the effusive rocks of the basin, that are characterized by a partially evolved magmatism. This implies that the magmatic products must have followed a fractionation, with the denser products left behind, the lighter products having reached the surface. The hypothesis of underplating is related to intrusions of mantle origin that have density intermediate to the mid crust and upper mantle, and therefore get trapped in the lower crust.

Other places where underplating occurs are in the LIP of West Siberia Basin, Karoo and also at the Valsesia super-volcano (Western Alps, Italy). We therefore are inclined to think that the underplating or the general densification of the crust below a LIP is a common feature, and related to the fact that only a small portion of the melted material reaches the surface, the remainder greater part contributing to both increase the crustal thickness and to increase the average crustal density. This result is important for the maturation history of the hydrocarbons in the basin, as the underplating rises the thermal isotherms with respect to a standard crust, and thus results in increased temperature in the sediments.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.jsames.2013.03.008.

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