

Density and Thermal Structure of the Southern Andes and Adjacent Foreland from 32° to 55°S Using Earth Gravity Field Models

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Abstract GOCE satellite data and EGM2008 model are used to calculate the gravity anomaly and the vertical gravity gradient, both corrected by the topographic effect, in order to delineate main tectonic features related to density variations. In particular, using the Bouguer anomaly from GOCE, we calculated the crust–mantle discontinuity obtaining elastic thicknesses in the frame of the isostatic lithospheric flexure model applying the convolution method approach. Results show substantial variations in the density, compositional and thermal structure, and isostatic and flexural behavior of the continental lithosphere along the Southern Andes and adjacent foreland region.

Keywords GOCE · EGM2008 · Isostasy · Density · Andes

1 Introduction

The study of the lithospheric structure related to density variations using gravimetric data has been improved in the past two decades. Since the early Earth gravity field models recovered from satellite orbit perturbation analysis to the last missions, specially designed for gravity information recovery such as GRACE (Gravity Recovery and Climatic Experiment) and GOCE (Gravity Field and Steady State Ocean Circulation Explorer), considerable improvements have been obtained. These satellite missions provide Earth gravity field models with increasing precision, resolution, and homogeneous coverage. The combination of data from

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different satellite missions leads to an excellent performance of both the long and short wavelengths of the gravimetric signal.

In particular, the model *GO_CONS_GCF_2_DIR_R5* (Bruinsma et al. 2013) is a satellite-only model, based on a full combination of GOCE-SGG with GRACE and LAGEOS (LAsER GEODYNAMICS Satellite). This model was obtained by the method of the direct approach, and is one of the maximum degree/order from satellite-only data (processing details are presented in Bruinsma et al. 2010; Pail et al. 2011). This allows obtaining a resolution of approximately $\lambda/2 \approx 2\pi R/N_{\max} \approx 66$ km with R being the mean Earth radius and $N_{\max} = 300$ the maximum degree and order of the harmonic expansion (Li 2001; Hofmann-Wellenhof and Moritz 2006; Barthelmes 2009).

Since the gravity field attenuates at high altitudes of satellite orbits, these satellites provide information only on the long wavelength part of the spectrum (Reguzzoni and Sampietro 2010). Thus, a way to improve the Earth gravity field model resolution is to incorporate terrestrial gravity data, as done in different models such as the EGM2008 (Pavlis et al. 2008, 2012) developed up to $N = 2190$ with $\lambda/2 = 9$ km, and EIGEN models from Förste et al. (2012) as EIGEN-6C4, also developed up to $N = 2190$ but including the last GOCE mission data. Despite its higher spatial resolution and global coverage, the disadvantage of these models is their inhomogeneous precision. For well-surveyed regions, where high quality terrestrial mean gravity anomalies are available, such as in North America, Europe, and Australia these models present geoid RMS-differences in the order of 4–6 cm (Yi and Rummel 2014). However, where surface gravity data are poor such as in South America, Africa, South-East Asia, and China regions, the RMS differences span between 20 and 38 cm (Yi and Rummel 2014).

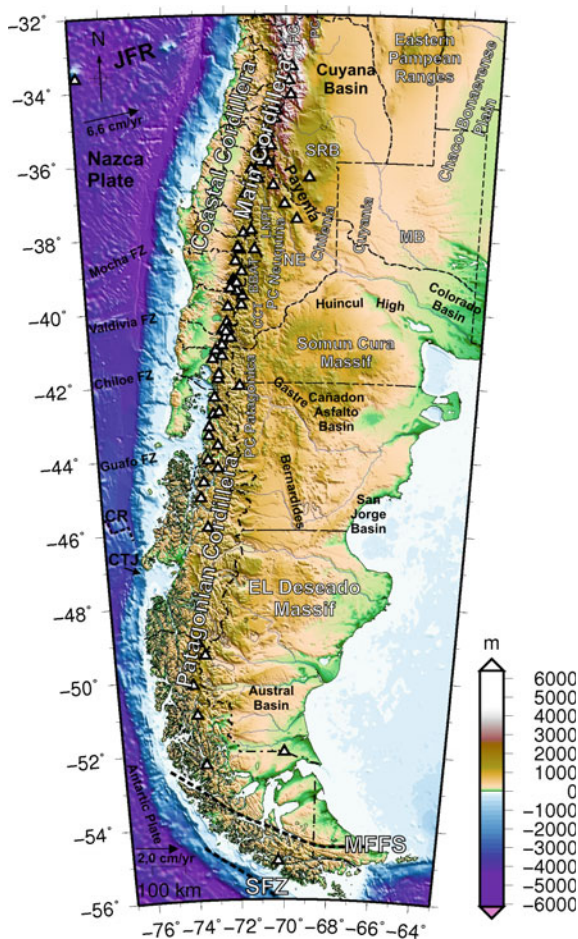
The combined use of both kinds of models allow interpretation of medium-to-high (satellite-only models) and low (integrated models as EGM2008, EIGEN) wavelength anomalies related to crustal inhomogeneities. Nevertheless, a statistical comparison is needed between satellite-only and integrated models in order to estimate the error assessment prior to interpretation and analysis. Satellite only models ensure data quality, since these present homogeneous precision and no errors or sampling biases are present. Köther et al. (2012) explained that combined gravity models (such as EGM2008) can be used for density modeling of relatively smaller features such as shallower crustal structures, while satellite-only models are not appropriate for this purpose due to a lower spatial resolution.

Despite this, the homogeneous precision and regional coverage make them useful for regional studies, allowing interpretation of the gravity potential-derived signals at medium-to-high wavelengths (Braitenberg et al. 2011; Álvarez et al. 2012, 2014, 2015; Mariani et al. 2013; Li et al. 2013; Braitenberg 2014). In the present work, we use the satellite-only model from *GO_CONS_GCF_2_DIR_R5* (Bruinsma et al. 2013) to delineate the density and thermal structure of the Andes and adjacent foreland region from 32° to 55°S. In those regions, where the EGM2008 model (Pavlis et al. 2008, 2012) has an acceptable performance, density anomalies are interpreted as related to medium-to-low wavelengths.

2 Main Morphostructural Units of the Southern Andes

The northern analyzed region is located in the southern part of the Pampean flat slab zone. At these latitudes (32°S – 33°S), the Andean structure is composed of a series of mountain ranges that comprises the Main Cordillera, the Frontal Cordillera, the southernmost region of the Precordillera, and the Sierras Pampeanas (Fig. 1; and see Chap. 1). The Frontal Cordillera is formed by a series of basement blocks that exposes Neo-Proterozoic to Paleozoic rocks, constituting the highest part of the fold and thrust belt at these latitudes (Ramos 1999). The Precordillera to the east is formed by thin- and thick-skinned imbricate fans, forming east-verging systems that expose Paleozoic to Triassic sequences, whose basal Laurentia-derived terms (Early Cambrian to Early Ordovician; Cuyania terrane; Fig. 1) were accreted to the Gondwana margin in Late Ordovician times (see Ramos 2004a, for a synthesis).

Fig. 1 Main morphostructural units mentioned in this chapter (*FC* Frontal Cordillera, *PC* Precordillera, *SRB* San Rafael Block, *NE* Neuquén Embayment, *MB* Maudas Block, *LNPT* Loncopué Through, *BBAT* BioBio Aluminé Through, *CCT* Collón Cura Through, *CR* Chile Rise, *CTJ* Chile Triple Junction), *MFFS* Magallanes Fagnano fault zone, *SFZ* Scotia fault zone. The Chilean–Argentinian border is indicated by a dot and dashed line and the coastal zone by a solid black line



Deformation in the Precordillera responsible for its present morphology occurred in the last 10 Ma, partly synchronous with the rise of the Pampean Ranges to the east in the foreland area (Jordan and Allmendinger 1986; Ramos et al. 2002). The thick-skinned Sierras Pampeanas constitute a set of Laramide-style basement blocks uplifted during the shallowing of the subducted Nazca Plate since late Miocene times (Ramos et al. 2002; Kay and Coira 2009). The NE region of the area under study comprises the Eastern Sierras Pampeanas that expose latest Proterozoic to Cambrian metamorphic rocks intruded by subduction-related to anorogenic magmatic rocks (Ramos 1988; Rapela et al. 1998; Lira et al. 1997). Late Miocene to Quaternary volcanic rocks unconformably overlying the Proterozoic to Paleozoic basement at the foreland area indicate the eastern migration of the volcanic arc during the flattening of the Nazca Plate in the last 17 Ma (Brogiotti 1990; Kay and Gordillo 1994).

South of this region (-35°S), the Payenia volcanic zone is characterized by mainly Quaternary extensive volcanic fields, located in the back-arc zone. These young volcanic rocks overly Paleozoic to Cenozoic rocks and more locally, Mesozoic sedimentary sections of the Neuquén embayment. The Neuquén embayment is segmented by basement structures with N-to-NW structures and more locally by the Huincul Ridge, a transverse-to-the Andes regional lineament (see Ramos 1999 and references therein), corresponding to a first order structural feature (Turner and Baldi 1978), interpreted in some works as the extensional reactivation of the northern limit of the Patagonia Paleozoic terrane (Ramos et al. 2004b). North of this, the Mahuidas basement block is constituted by Precambrian metamorphic rocks intruded by Paleozoic plutonic rocks (Linares et al. 1980).

The Patagonian Cordillera to the south, a relatively narrow mountain system compared with the Central Andes to the north, is characterized by a continuous Mesozoic to Cenozoic batholith emplaced along the western slope and axial zone of the Andes, from nearly 39°S to about 52°S (Hervé et al. 2004). A strike-slip crustal structure, the Liquiñe–Ofqui fault zone (LOFZ) (Fig. 1) (Hervé et al. 1996; Cembrano et al. 1996, among others) crosses the axial part of the Northern Patagonian Andes, controlling the emplacement of the volcanic arc corresponding to the Southern Volcanic Zone (see Ramos 2009 and references therein).

Two basement-foreland cores characterize the extra-Andean region, the Somuncura Massif (Windhausen 1931) and the Deseado Massif. The central zone of these foreland Paleozoic cores is partly covered by siliceous Jurassic rocks associated with widespread synextensional volcanic activity during Pangea breakup (Uliana et al. 1985; Alric et al. 1996; Feraud et al. 1999; Ghidella et al. 2002; Ramos 2004b). These foreland cores are flanked by Mesozoic extensional basins such as Cañadón Asfalto, Río Mayo, Austral, and San Jorge Gulf basins (Fig. 1). The Precordillera Patagonica across the Patagonian foreland zone constitutes a feature separated from the Andes to the west that results from the partial inversion of these extensional basins (Ramos 1999).

3 The Topography-Reduced Gravity Anomaly (Ga)

The observed potential is obtained from the Earth gravity field model. The disturbing potential (T) is calculated subtracting the potential field from the reference ellipsoid (WGS84) (Janak and Sprlak 2006). This disturbing potential allows calculation of different derived quantities, related to crustal density heterogeneities. The topography reduced gravity anomaly ($\Delta_{g_{rr}}$) is the difference between the real gravity (g) at a given point (h, λ, ϕ), and the gravity of the reference ellipsoid (γ) at the same coordinates, but at the ellipsoidal height $h - \zeta_g$ (where ζ_g is the generalized height anomaly) and without considering the effect of the topographic masses above the geoid (g_r) (Molodensky et al. 1962; Hofmann-Wellenhof and Moritz 2006; Barthelmes 2009). The height h is assumed on or above the Earth surface ($h \geq h_r$). Thus, the measured gravity at the Earth surface can be used for calculation of the gravity anomaly, in the modern definition, without downward continuation or any reduction as it is a function in the space outside the masses (Barthelmes 2009).

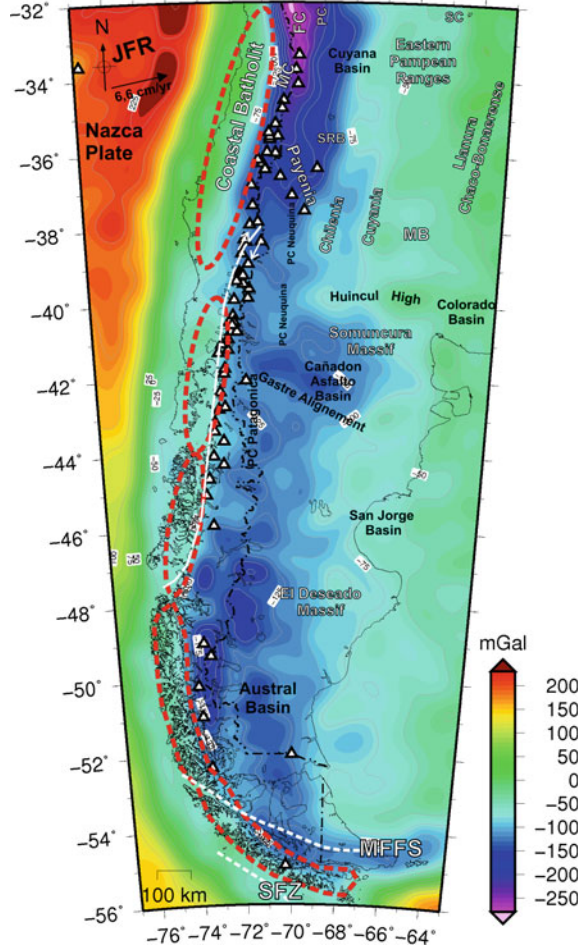
$$\Delta_{g_{rr}}(h, \lambda, \phi) = g(h, \lambda, \phi) - g_r(h, \lambda, \phi) - \gamma(h - \zeta_g, \lambda, \phi) \quad (1)$$

For calculation of the gravity anomalies (Janak and Sprlak 2006), we used the model of GOCE *GO_CONS_GCF_2_DIR_R5* (Bruinsma et al. 2013) up to its maximum degree/order = 300 (Fig. 2), and the EGM2008 (Pavlis et al. 2008, 2012) up to degree/order = 2159 (Fig. 3). For an approximate calculation of the effect of the topographic masses, we used a global relief model which includes ocean bathymetry (ETOPO1, Amante and Eakins 2009) and a mass density distribution hypothesis. We utilized standard densities of 2670 kg/cm³ for continental crust and a density of 1030 kg/cm³ for the sea. The topographic effect was calculated at a height of 7000 m to ensure that all values are above the topography and in a spherical coordinate system. The topographic elements were approximated using spherical prisms or Tesseroids (Uieda et al. 2010; Alvarez et al. 2013) taking into account the Earth curvature in order to avoid the error introduced when using a planar approximation (Grombein et al. 2013; Bouman et al. 2013). All calculations were carried out with respect to the system WGS84 on a regular grid of 0.05° grid cell size. These topography corrected gravity anomalies are useful for highlighting the effects of different rock densities into the crust, geological structures and basins.

4 The Topography Corrected Vertical Gravity Gradient (Tzz)

The gravity gradient tensor (Marussi tensor) is obtained as the second derivative of the disturbing potential T (e.g., Hoffman-Wellenhof and Moritz 2006) and is composed of five independent elements. The Marussi tensor components $M = T_{ij}$ can be expressed and solved numerically in a spherical coordinate system

Fig. 2 Gravity anomaly from GOCE model GO_CONS_GCF_2_DIR_R5 (Bruinsma et al. 2013) corrected by the topographic effect (FC Frontal Cordillera, PC Precordillera, SC Sierras de Cordoba, MC Main Cordillera, SRB San Rafael Block, MB Mahuidas Block, and JFR Juan Fernandez Ridge). Thin white dashed line represents the Liquiñe Ofqui fault zone, MFFS Magallanes Fagnano fault zone, SFZ Scotia fault zone. The Chilean–Argentinian border is indicated by a dot and dashed line and the coastal zone by a solid black line

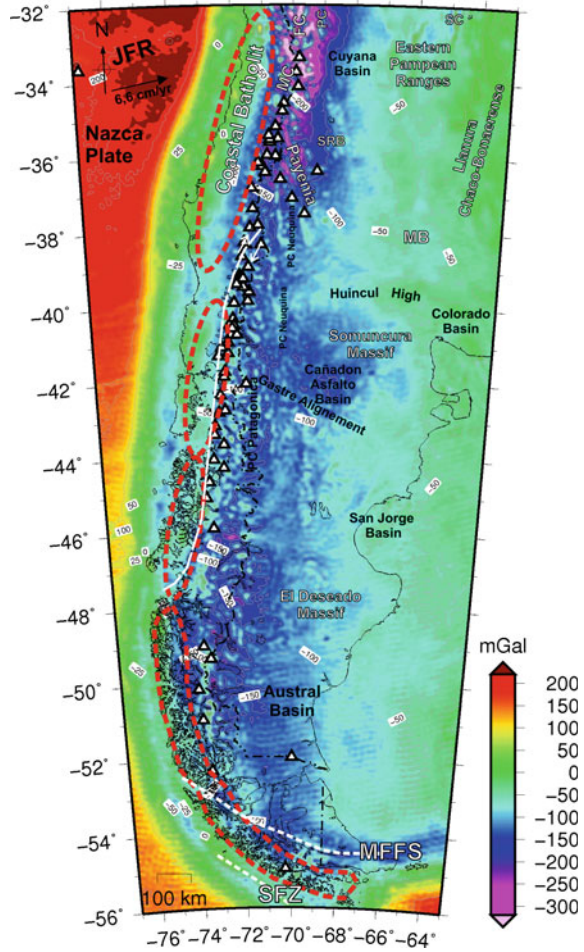


(Tscherning 1976; Rummel et al. 2011). The vertical gravity (T_{zz}) gradient is the second radial derivative of the disturbance potential T .

$$T_{zz} = \frac{\partial^2 T}{\partial r^2} \left[1\text{Eötvös} = 10^{-4} \frac{\text{mGal}}{\text{m}} \right] \quad (2)$$

The vertical gravity gradient highlights superficial density anomalies and allows delineating the location of an anomalous mass with better detail and accuracy than the gravity anomaly itself (Braitenberg et al. 2011). Since the T_{zz} is a derivative of gravity, the spectral power of gravity gradient signals is pushed to higher frequencies, resulting in a signal more focalized to the source than the gravity anomaly (Li 2001). Therefore T_{zz} is better for detection of the edge of geological structures and to distinguish the signal due to a smaller superficial density variation from an

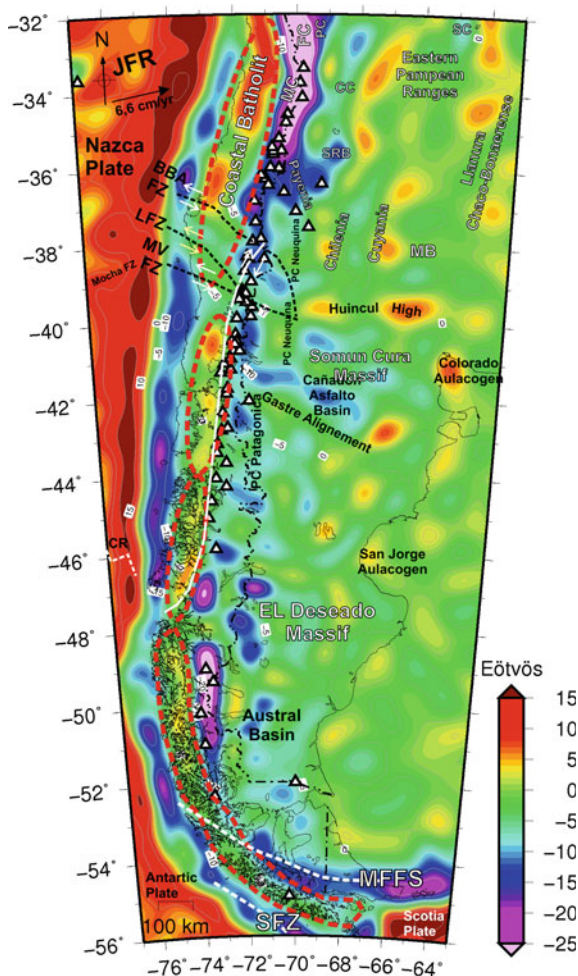
Fig. 3 Gravity anomaly from EGM2008 model (Pavlis et al. 2008, 2012) corrected by the topographic effect



extensive deeper mass, being the *Ga* more sensitive to regional signals and deeper sources (Alvarez et al. 2012). Relative denser bodies are related to a positive gradient value while a negative gradient is related to less dense bodies. Abrupt changes may indicate a high-density contrast between two different lithologies.

We calculated the vertical gravity gradient from GOCE *GO_CONS_GCF_2_DIR_R5* (Bruinsma et al. 2013) in a geocentric spherical coordinate system at the calculation height of 7000 m to ensure that all values were above the topography (Fig. 4) and also for the EGM2008 model (Fig. 5). The values were calculated on a regular grid with a cell size of 0.05°, with the maximum degree and order of the harmonic expansion ($N = 300$, $N = 2159$) for each model. The topographic effect was removed from the fields in order to eliminate the correlation with the topography (Alvarez et al. 2013) as made for the gravity anomaly. The topographic correction amounts up to tens of Eötvös for the T_{zz} , becoming higher over the

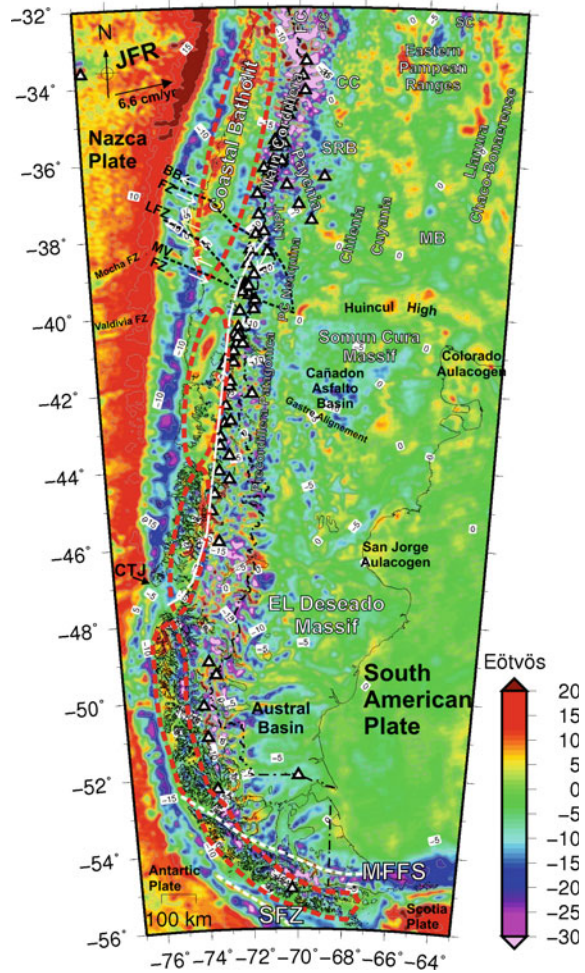
Fig. 4 Vertical gravity gradient from GOCE model GO_CONS_GCF_2_DIR_R5 (Bruinsma et al. 2013) corrected by the topographic effect (BBAFZ BioBio Alumine fault zone, LFZ Llanahue fault zone, MVFZ Mocha Villarica fault zone, MFFS Magallanes Fagnano fault zone, SFZ Scotia fault zone)



maximum topographic elevations (e.g., the Main Andes) and lower over the topographic depressions such as the Chilean Trench.

After a statistical analysis and considering the sparseness of terrestrial data in the region under study (Pavlis et al. 2008, 2012), the most reliable areas to apply the EGM2008 model are mainly offshore and forearc regions comprised between 36° and 48°S. Thus, the GOCE model (TIM_R5) becomes more appropriate than the EGM2008 model in the region under study despite its lower resolution (but homogeneous precision). Results obtained by means of the EGM2008 model will only be analyzed and compared to GOCE in the regions of higher performance, in order to solve the different anomalies in greater detail. The combined use of both models, considering their best individual qualities, has been already tested in different studies (Braitenberg et al. 2011; Alvarez et al. 2012, 2014, 2015).

Fig. 5 Vertical gravity gradient from EGM2008 model (Pavlis et al. 2008, 2012) corrected by topographic effect. *CTJ* Chile triple junction



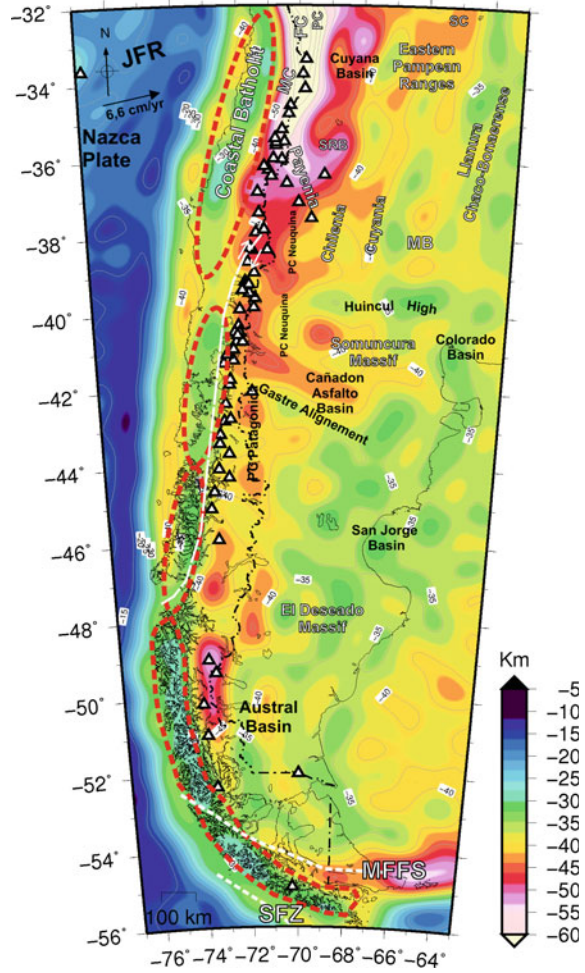
5 Isostatic Modeling

Isostatic theory explains the elastic behavior of the lithosphere, describing the effect of the inhomogeneous distribution of the crustal loads. The Airy model considers that different cortical sections behave in hydrostatic equilibrium, constituting a “topographic load” compensated by a “crustal root”

$$\Delta R = \frac{\sigma}{\sigma_1 - \sigma} h \quad (3)$$

where σ is the crustal density, σ_1 is the mantle density, ΔR : the crustal root or compensation and h is the topography. As this expression indicates, crustal and

Fig. 6 Moho from
GOCE model
GO_CONS_GCF_2_DIR_R5
(Bruinsma et al. 2013)
obtained from inversion of the
Bouguer anomaly

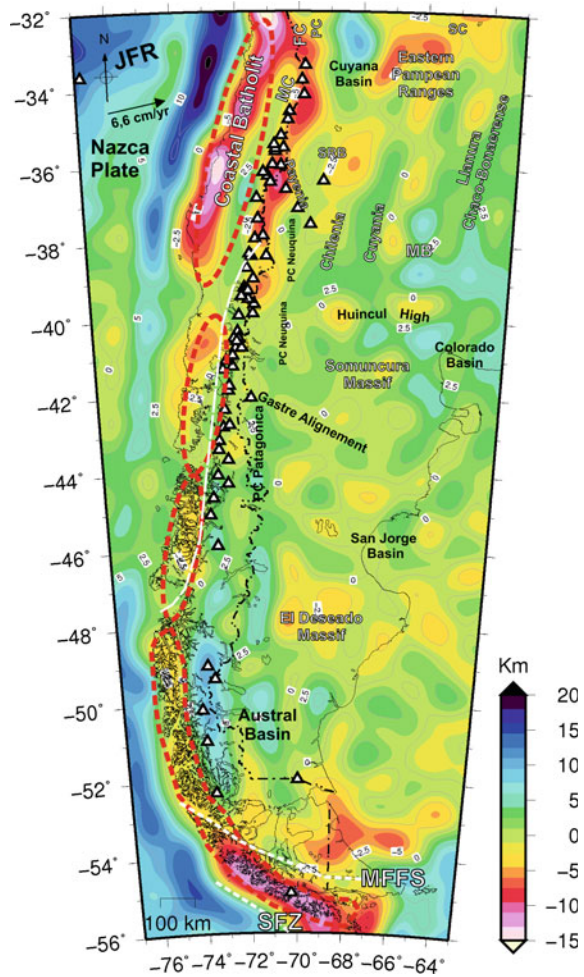


mantle densities remain constant and consequently the depth of compensation varies in function of the topography. For calculation of the Airy response we considered a crust–mantle density contrast of 400 kg/cm^3 and the relief model ETOPO1.

Then, we obtained the depth to the crust–mantle interface (Hydrostatic Moho) by using Eq. (3), taking into account as normal crustal thickness $T_n = 35 \text{ km}$. This Moho was then compared (by subtracting) to the crust–mantle interface (Fig. 6) obtained by inversion of the Bouguer anomaly (Fig. 2) calculated from the GOCE model *GO_CONS_GCF_2_DIR_R5* (Bruinsma et al. 2013). Near to zero values of this “residual” Moho (Fig. 7) indicates different regions that are near the isostatic equilibrium or that behave as compensated by a crustal root, while negative values indicate an overcompensation (the crustal root obtained by means of the Bouguer

Fig. 7 Residual between the isostatic Moho (Airy) and Moho obtained by inversion of the Bouguer anomaly from GOCE model

GO_CONS_GCF_2_DIR_R5 (Bruinsma et al. 2013). Near zero values indicates a hydrostatic balance, while negative values indicate an overcompensation of the crustal root and positive ones the expression of a deficiency in isostatic compensation



anomaly is deeper than the one obtained following the Airy hypothesis), and positive values indicate an undercompensation, i.e., the crustal root is not deep enough to compensate the topographic loads.

6 Effective Elastic Thickness

The compensation of the topographic masses is approximated through the Airy theoretical model that considers a local compensation and the Vening Meinesz model that indicates that the compensation is performed by a regional bending of

the plate, which best describes the mechanical behavior and stress state of a loaded lithosphere (Introcaso 2006).

The thickness and viscosity of the lithosphere, or lithospheric strength, is related to the thermal state, rheological and compositional properties of the crust and is well characterized by the flexural rigidity (Lowry et al. 2000). The last can be interpreted in terms of the elastic thickness (T_e) by making some assumptions regarding the Young modulus and the Poisson ratio. T_e defines the maximum size and wavelength of the surface loading that can be supported without an elastic break of the lithosphere. The flexural rigidity D can be expressed by:

$$D = T_e^3 \cdot \frac{E}{12(1 - \nu^2)} \quad (4)$$

where $E = 100 \text{ GPA} = 10^{11} \text{ N/m}^2$ is the Young's modulus, and $\nu = 0.25$ is the Poisson's ratio. The isostatic state and deformation of the upper crust are reflected in the spatial distribution of the T_e , whose variation can be explained by heat flow distribution and a change in the Young modulus. A relation between T_e and the geometry and composition of the flexured plate, external forces and the thermal structure has been reported by different authors (e.g., Göetze and Evans 1979; Lyon-Caen and Molnar 1983; Burov and Diamant 1995; Hackney et al. 2006).

Different methods have been developed for estimation of the elastic thickness, as the flexural coherence analysis, the spectral methods, and the convolution approach (Braitenberg et al. 2002; Wienecke 2006). The last has the advantage of being a method of double entry that calculates the flexure parameters by the best fit of the observed crust–mantle interface obtained by gravity inversion and a crust–mantle interface computed due to a flexure model see Fig. 8 (www.lithoflex.org) (Wienecke 2006; Wienecke et al. 2007; Braitenberg et al. 2007). Isostatic modeling adopts the isostatic thin plate flexure model (e.g., Watts 2001). To estimate the elastic properties of the plate for a known load are needed a crustal load (combination of the overlying topography plus a density model) and the crust–mantle interface to be used as a reference surface (Wienecke 2006).

The topographic load was calculated using topographic/bathymetric data from ETOPO1 (Amante and Eakins 2009). The densities used for calculation were 1030 kg/m^3 for water and 2800 kg/m^3 for the crust. The undulating boundary corresponding to the Moho was calculated by gravity inversion of the Bouguer anomaly field (from GOCE *GO_CONS_GCF_2_DIR_R5*, <http://icgem.gfz-potsdam.de/ICGEM/>, Bruinsma et al. 2013). This method requires two input parameters, the density contrast (400 kg/m^3) and reference depth or normal crustal thickness ($T_n = 35 \text{ km}$) (values between 30 and 40 km have been adopted in numerous works as normal crustal thickness).

The flexural rigidity was inverted in order to match the known loads with the known crustal thickness model (i.e., to model the gravity Moho in terms of an isostatic model). The elastic thickness was allowed varying in the range of $1 < T_e <$

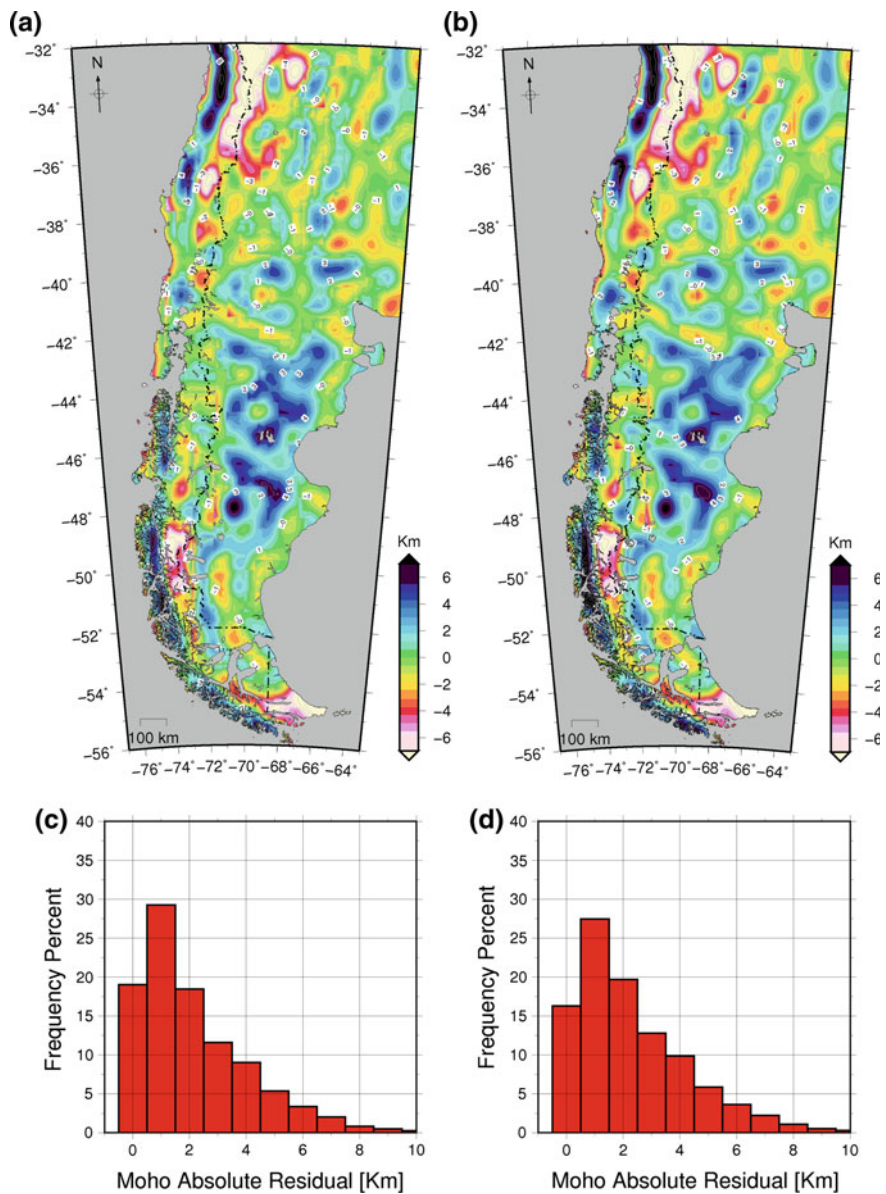


Fig. 8 Up Residual obtained by subtracting the crust-mantle interface obtained by gravity inversion of the Bouguer anomaly minus the crust-mantle interface obtained from the flexural model. **a** Window size 80×80 km. **b** Window size 120×120 km. Down Histograms of the residual Moho between the crust-mantle interface obtained by gravity inversion and the crust-mantle interface from the flexural model. More than 80 % of the error is less than 4 km. **c** Window size 80×80 km. **d** Window size 120×120 km. The residual obtained for the window size of 80×80 km shows a slight better fitting

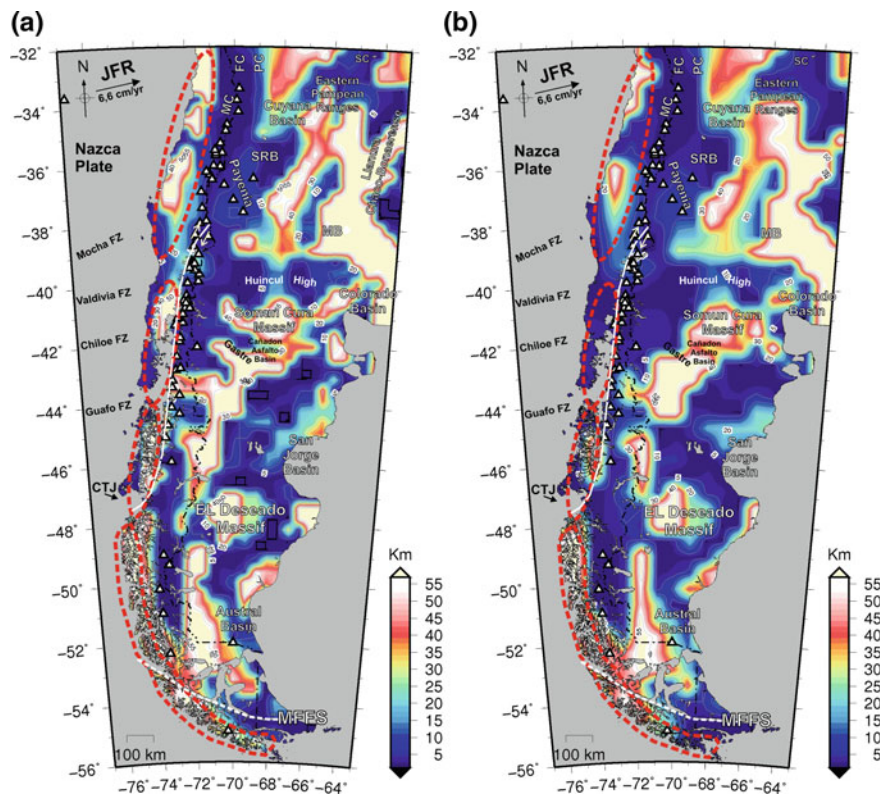


Fig. 9 Effective elastic thickness for the Southern Andes and adjacent foreland region from GOCE model. **a** Window size of 80 × 80 km. **b** Window size of 120 × 120 km

Table 1 Parameters used in the flexural modeling

Masses above sea level	ρ_s	2670 kg/m ³
Upper crustal density	ρ_{uc}	2700 kg/m ³
Lower crustal density	ρ_{lc}	2900 kg/m ³
Mean crustal density	ρ_{mc}	2800 kg/m ³
Upper mantle density	ρ_m	3300 kg/m ³
Young modulus	E	10 ¹¹ N/m ²
Poisson ratio	Σ	0.25

60 km and was iteratively estimated over moving windows of 40 × 40 km size (Fig. 9a) and with 60 × 60 km window size (Fig. 9b). Model parameters (Table 1) and adopted densities are standard values already used by Introcaso et al. (2000); Gimenez et al. (2000, 2009); Alvarez et al. (2015) in the region under study.

7 Results

Low gravity values (less than -250 mGal) north of 38°S along the Andes detected in the Ga from GOCE (Fig. 2) and from EGM2008 (Fig. 3) are related to the negative gravimetric response of the Andean roots. Computed Moho depths (Fig. 6) are consistent in the southern section of the Pampean flat subduction zone ($32\text{--}34^{\circ}\text{S}$) with those obtained by Gans et al. (2011), based on receiver functions. Low-gravity anomalies (Figs. 2, 3), a deep Moho (Fig. 6) and low Te values (Fig. 9) are a consequence of a flexural model in which the plate has little to no strength. The isostatic residual Moho (Fig. 7) indicates that this region of the Andes is overcompensated along the active arc front, while previous results had indicated that this region was near isostatic balance (Introcaso et al. 1992; Stewart and Watts 1997; Tassara and Yañez 2003; Tassara et al. 2007; Pérez-Gussinyé et al. 2008; Sacek and Ussami 2009).

Seaward of the trench at the latitudes of the Pampean flat subduction zone, the flexural outer rise of the Nazca Plate, is marked by a positive gravity anomaly higher than $+200$ mGal (Figs. 2, 3) and more than $+20$ Eötvös (Figs. 4, 5). Particularly, the Juan Fernández Ridge over the Nazca Plate is delineated by a well-defined signal in both T_{zz} and gravity anomaly differentiated from the surrounding sea-floor (Alvarez et al. 2015). The Moho is deeper along the Juan Fernández Ridge indicating an over-thickened oceanic crust beneath the ridge as described by von Huene et al. (1997) based on wide-angle seismic data and by Sandwell and Smith (1997) who related negative satellite-derived gravity anomalies to a crustal root indicative of crustal flexure from loading.

Relative low Ga values to the south of 38°S (Fig. 2) reflect the reduction of the Andean roots (Fig. 6) in coincidence with a segment of lower Andean elevations. In particular, between 39° and 48°S the residual Moho indicates a state near isostatic balance (Fig. 7). Contrastingly, south of the inception zone of the Chilean Ridge into the trench along the Southern Patagonian Cordillera, the residual Moho indicates an undercompensation of the Andean root.

Paleozoic sedimentary rocks exposed in the Precordillera region above the Pampean flat subduction zone are identified by a more positive gravimetric signal than the general negative gravimetric response of the Main Andes (Fig. 2). Other works (Introcaso et al. 1992; Gimenez et al. 2000, among others) had also reported, based on terrestrial gravimetric data, a high gravimetric signal for the Precordillera area (see Alvarez et al. 2012, 2014 for a detailed analysis). The Late Triassic Cuyo sedimentary basin located SE of the Precordillera is detected by low gradient values of less than -20 Eötvös and less than -150 mGal. The crust–mantle interface is shallower through the Precordillera region (Fig. 6) and behaves near the isostatic balance (Fig. 7). Higher values of Te are detected in this area in relation to the Main Cordillera. The T_{zz} shows a high gradient signal for the Precordillera in both GOCE (Fig. 4) and EGM2008 (Fig. 5). The flat slab region (north of 33°S) presents high-rigidity values most likely delineating the Cuyania Terrane boundaries as described by Alvarez et al. (2015). High Te values over flat-slab regions have been

related to low thermal regimes (Stewart and Watts 1997; Tassara 2005; Alvarez et al. 2015). In this sense, Pérez-Gussinyé et al. (2008) considered that flat slabs are characterized by several parameters among which high T_e values, high shear wave velocities, thick thermal lithospheres, and low heat flows are characteristic, indicating that the continental lithosphere above these settings become cooler and consequently thicker.

East of the Precordilleran region, a broad broken foreland zone is developed associated with the Pampean flat slab, where the Eastern Sierras Pampeanas are uplifted in the eastern border of the Rio de la Plata Craton. These uplifts are represented by Neoproterozoic basement blocks which present low gradient gravity values characteristic of their composition. Gravity data indicate that these mountain ranges are developed over intermediate Moho depths (approx. 40 km), consistent with previous works (Gans et al. 2011). Even shallower Moho depths are calculated beneath the Rio de la Plata Craton to the east. We found intermediate to low T_e values in the Eastern Sierras Pampeanas with increasing values eastwards to the foreland region; this is consistent with the residual Moho that indicates a regional overcompensation under this zone. Different authors have found that the Eastern Sierras Pampeanas are undercompensated (Introcaso et al. 1992, 2000) for wavelengths in the order of range topography (less than 50 km), based on terrestrial gravity data. This is consistent with the tectonic evolution of these ranges that were uplifted by listric faults as is the case of the Western Sierras Pampeanas (Weidmann et al. 2013).

The Payenia region and the San Rafael block south of the Pampean flat subduction zone present low T_{zz} values, a relatively deep Moho and low rigidity values. In particular, in the Payenia region the isostatic residual Moho is near-to-zero and the T_e is near 10 km (Fig. 9), contrastingly to the San Rafael Block that is overcompensated, with a lower T_e (less than 5 km), similarly to the Sierras Pampeanas to the north (at higher wavelength), showing the continuity of the broken foreland zone south of the Pampean flat subduction zone.

South of the Juan Fernández Ridge up to the latitudes of the triple point determined by the Chile Ridge with the trench, the region comprehended between this and the coastline is defined by a negative T_{zz} anomaly, less than -10 Eötvös. This indicates the existence of low-density materials filling the trench and a thinner crust beneath the forearc (Alvarez et al. 2014). Sediment filled slope basins, resting upon thinned continental crust that extends to about 50 km off the trench, had been recognized in free-air gravity maps and in seismic profiles of the southern Chile margin (ANCORP Working Group 2003; Flueh et al. 1998; von Huene et al. 1997). Wells et al. (2003) had found prominent free-air gravity lows following the physiographic slope and basins in areas of high sedimentation rates as in the Southern Chile continental margin. This low-gravity signal is segmented by transform zones over the Nazca plate that segment the Chile Ridge that enter obliquely to the trench. These transform faults are described as highly serpentinized crustal settings by Contreras-Reyes (2008), where oceanic lithosphere is segmented by its different ages and consequently by variable lithospheric thicknesses (Tebbens and Cande 1997).

West of the Main Andes, positive gravimetric values and higher T_{zz} parallel to the coastline are the expression of the Coastal Batholith exhumed along the Coastal Cordillera (Hackney et al. 2006). The residual Moho over this region indicates an overcompensation of the Coastal Batholith adjacent to an undercompensated region along the Central Valley of Chile west of the Main Andes.

At 39°S, a displacement of 100 km between the axis of the Coastal Cordillera Batholith and the Paleozoic Patagonian Batholith is observed. An interruption of the gradient signal in coincidence with this jump coincides with the Bio-Bio Aluminé lineament along the arc and retroarc zones (García Morabito et al. 2003; Rosenau et al. 2006). In this region, the upper plate is segmented in the forearc zone by major fault zones with general NW-trends (Sick et al. 2006; Hackney et al. 2006; Krawczyk et al. 2006; Ramos et al. 2011; among others).

Between 39° and 52°S, high T_{zz} values detected along the back-arc zone are related to plutonic rocks of the Patagonian Batholith. Along this Northern region of the Patagonian retroarc, Moho depths are rather constant (Fig. 6) indicating in the residual Moho a certain degree of overcompensation (Fig. 7).

Near the Taitao peninsula, where the Chile Ridge is presently colliding, low Ga values (<0 mGal) and less than -10 Eötvös interrupt the high-gravity values characteristic of the Coastal Batholith. South of this low, the alternation of high and lows in the T_{zz} values probably indicates a truncated subduction channel most likely as a consequence of the collision of Chile Ridge against the continental margin in the last 3–6 My (Ranero et al. 2006; Bourgois et al. 1996; Cande and Leslie 1986).

Between 48° and 55°S, the Patagonian Batholith is expressed by high T_{zz} values that are interrupted at 53°S by the Magallanes–Fagnano fault system (MFFS) which marks the limit between the South American and Scotia plates (Dalziel et al. 1974; Klepeis 1994; Olivero 1998; Lodolo et al. 2003; Ghiglione and Ramos 2005; Ramos 2009). Computed Moho depths (Fig. 6) are shallower in the southernmost region of the Patagonian Andes (-25 km) next to this plate limit than to the north (-35 km). The isostatic Residual Moho (Fig. 7) shows a certain degree of overcompensation which increases south of the Magallanes–Fagnano fault system.

In the north central Patagonia, the North Patagonian Massif is characterized by high T_{zz} values, with low Ga values indicating the existence of a crustal root. Low T_{zz} values in the south central region of the massif are the expression of the Cañadon Asfalto basin, while north of it, the Huincul High (Bettini 1984; Ramos et al. 2004a; Mosquera and Ramos 2006) is expressed by an arched and elongated anomaly of high T_{zz} values, also detected in the Ga maps. Positive magnetic anomalies along this feature were related to mafic and ultramafic rocks potentially emplaced in a suture between Northern Patagonia and the southern Gondwana margin (Chernicoff et al. 2008). Moho depths are shallower along the Huincul High than to the south over the Somuncura Massif. The central region of the Somuncura and Deseado Massifs to the south presents high rigidity values (Fig. 9).

The Deseado Massif is expressed by high T_{zz} values and shows to be slightly overcompensated isostatically. The Cretaceous sedimentary deposits filling the Austral basin, and further north in the Río Mayo embayment (Ramos and Aguirre-Urreta 1994) are denoted by low T_{zz} values at the western and southern sides of the Deseado Massif.

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