

Towards understanding the lithospheric structure of the southern Chilean subduction zone (36°S–42°S) and its role in the gravity field

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SUMMARY

The Southern Andes differ significantly from the Central Andes with respect to topography and crustal structures and are, from a geophysical point of view, less well known. In order to provide insight into the along-strike segmentation of the Andean mountain belt, an integrated 3-D density model was developed for the area between latitudes 36°S and 42°S. The model is based on geophysical and geological data acquired in the region over the past years and was constructed using forward density modelling. In general, the gravity field of the South American margin is characterized by a relatively continuous positive anomaly along the coastline and the forearc region, and by negative anomalies along the trench and the volcanic arc. However, in the forearc region of the central part of the study area, located just to the south of the epicentre of the largest ever recorded earthquake (Valdivia, 1960), the trench-parallel positive anomaly is disrupted. The forearc gravity anomaly differences thus allow the study area to be divided into three segments, the northern Arauco-Lonquimay, the middle Valdivia-Liquiñe, and the southern Bahía-Mansa-Osorno segment, which are also evident in geology. In the proposed model, the observed negative gravity anomaly in the middle segment is reproduced by an approximately 5 km greater depth to the top of the slab beneath the forearc region. The depth to the slab is, however, dependent upon the density of the upper plate structures. Therefore, both the upper and lower plates and their interaction have a significant impact on the subduction-zone gravity field.

Key words: crustal structure, density, forearc, gravity anomalies, subduction zone, southern Andes.

1 INTRODUCTION

The Andes, extending along the western continental margin of South America, are the longest system of high mountain ranges on Earth. They are the result of the convergence of the oceanic Nazca and Antarctic Plates toward the continental South American Plate, with a convergence rate of 62 mm a⁻¹ (Kendrick *et al.* 2003). The Andes are characterized by extreme along-strike variations, creating segments variable in width, altitude, climate and age. This segmentation of the western South American margin is the subject of extensive international research (e.g. Barazangi & Isacks 1976; Jordan *et al.* 1983; Mpodozis & Ramos 1990; Tassara *et al.* 2006). The Southern Andes, being characterized by relatively low topography, a narrow volcanic arc and no plateau, contrast to the Central Andes. In addition, the Nazca plate subducting in the south is younger and seismically less active than in the Central Andes, as is evidenced by the absence of deep-focus events (below 200 km depth) south of 38°S (e.g. Barazangi & Isacks 1976; Herron 1981; Bohm *et al.* 2002).

In 2000 and 2001, two geophysical experiments were performed in the Southern Andes, the aim of which was to collect new offshore

and onshore data in order to image the southern part of the Andean subduction zone. The structures between 36°S and 39°S were imaged along four 2-D east–west wide-angle reflection/refraction seismic profiles (Lüth *et al.* 2003; Krawczyk *et al.* 2006). Additional data obtained from a seismological experiment (Bohm *et al.* 2002) provide constraints on the subducting plate and for a local earthquake tomography model (Bohm 2004). In addition, a receiver-function study along one profile (Yuan *et al.* 2006) was performed. These existing geophysical data were used as constraints for a combined 3-D gravity-seismic model.

A seismic model must be consistent with an observed gravity field (Barton 1986) since seismic velocities and gravity anomalies are both dependent on rock properties, although they are measured separately. Thus, densities employed in the model should satisfy both the gravity field and velocity models. Velocity models are usually tested against gravity anomalies (e.g. Wang *et al.* 2004; Hrubcová *et al.* 2005) and joint inversion of seismic and gravity data is also becoming more common (e.g. Nielsen & Jacobsen 2000). A joint inversion was not performed during this study and only one of the available velocity models has been checked against gravity anomalies (Lüth *et al.* 2003). Thus, one of the aims of the density modelling

presented here is to merge the seismic data, gravity anomalies and all other available geophysical data into a 3-D structure and to test how well these data are correlated.

Although the geophysical data are available between 36°S and 39°S, the density model extends further to the south. It covers the region between 36°S and 42°S and extends from the trench in the Pacific ocean to the Atlantic coast. Hence, the 3-D modelling based on the gravity data should bring more insight into the structures that have not yet been imaged by other geophysical experiments. For this reason, the available gravity database was, prior to the modelling, analysed and reprocessed. A range of models was tested to examine the influence of different modelled structures within the oceanic and continental plates on the gravity field (Section 6 of Tašárová 2004). In addition, several alternative models that reproduce the observed gravity field were derived to identify the possible sources of the gravity anomalies and density inhomogeneities. The primary interest of the modelling and interpretation is the subduction zone between the trench and the main volcanic arc.

2 TECTONIC AND GEOLOGICAL SETTING

The Southern (Neuquén) Andes from 37°S to 39°S form a transitional segment between the Central Andes to the north and the Patagonian Andes to the south and are characterized by significant changes in topography and geology (Melnick *et al.* 2005). From the broad and high (greater than 4000 m) Central Andes, the Principal Cordillera narrows and descends significantly to mean altitudes of around 1000 m towards the south. Also, the Coastal Cordillera,

with a maximum elevation of ~1500 m at latitudes between 36°S and 38°S (in the Nahuelbuta Mountains) decreases to only a few hundred meters south of 39°S (Fig. 1).

The continental plate in the study region may be divided, from west to east, into the forearc, main arc and backarc. The forearc is comprised of two morphotectonic units (Melnick & Echtler 2006), the Coastal Cordillera (or Coastal range) and the Longitudinal Valley, also called the Central Valley or Central Depression (Fig. 2). The main volcanic arc (Principal Cordillera) is a chain of Quaternary centers on top of a Meso-Cenozoic arc and related intra and backarc basins. The Neuquén Basin is a Late Triassic to Early Cretaceous backarc basin (Ramos 1984, 1989) superimposed on a series of Palaeozoic amalgamated blocks: (i) the Río de la Plata craton in the easternmost part of the basin, corresponding to the western margin of Gondwana in Late Proterozoic times; (ii) Cuyania and Chilenia blocks beneath the western part of the basin, two Laurentia-derived continents accreted during the Palaeozoic and (iii) the North Patagonian massif in the southern part, which is potentially connected to the early Permian accretion of Northern Patagonian at the Gondwana margin.

The offshore forearc is dominated by offshore basins (Itata, Arauco and Valdivia) (Fig. 2) filled with Triassic continental and Late Cretaceous to Quaternary marine and continental sedimentary sequences (e.g. Mordojovich 1981; Pineda 1986) with a thickness of up to 2.5 km (Reichert & Schreckenberger 2002). The Coastal Cordillera in the onshore forearc consists mainly of a low-grade late Palaeozoic accretionary complex (also called accretionary prism), represented by the metamorphic rocks of two parallel belts. These belts are called the Western Series and the Eastern Series, and in the

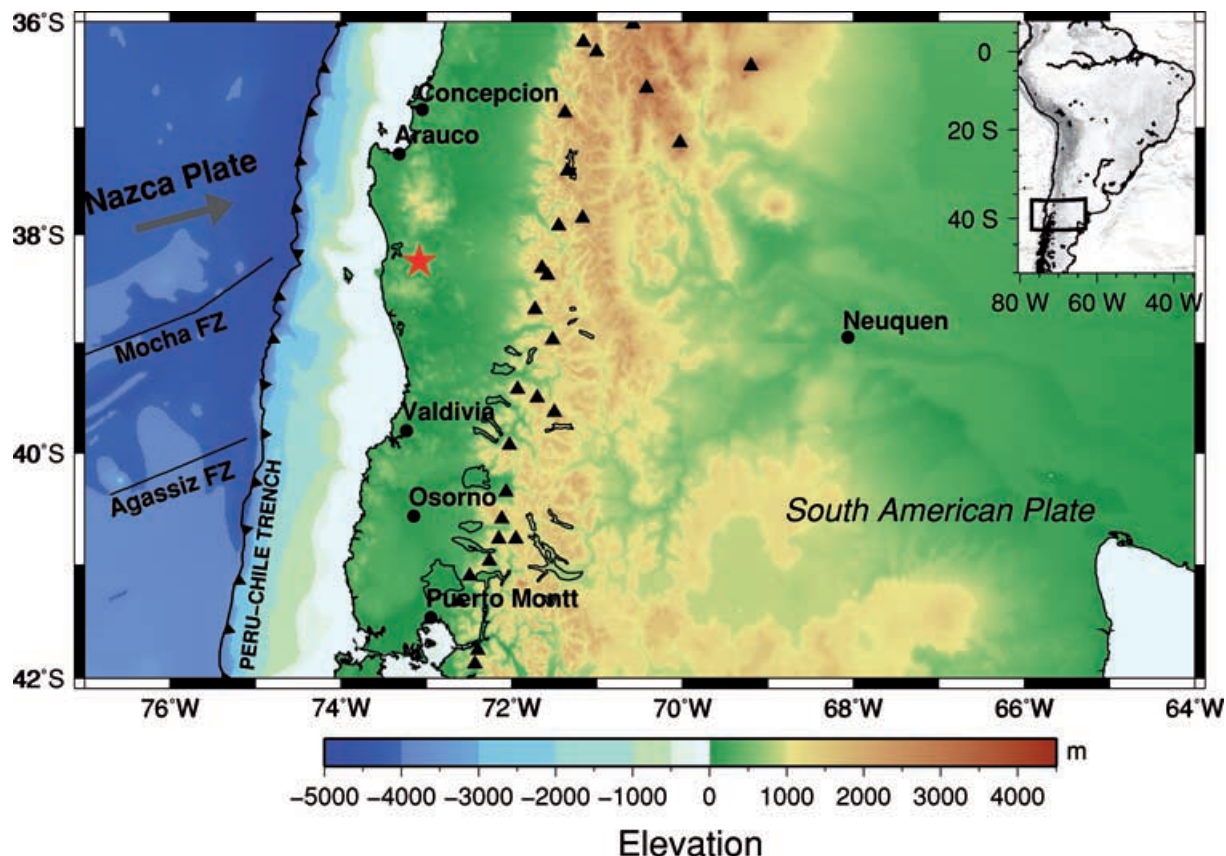


Figure 1. Bathymetry and topography of the study area (GLOBE DEM 1999). The red star indicates the epicentre of the Valdivia earthquake (Krawczyk *et al.* 2003) and black triangles show active volcanoes.

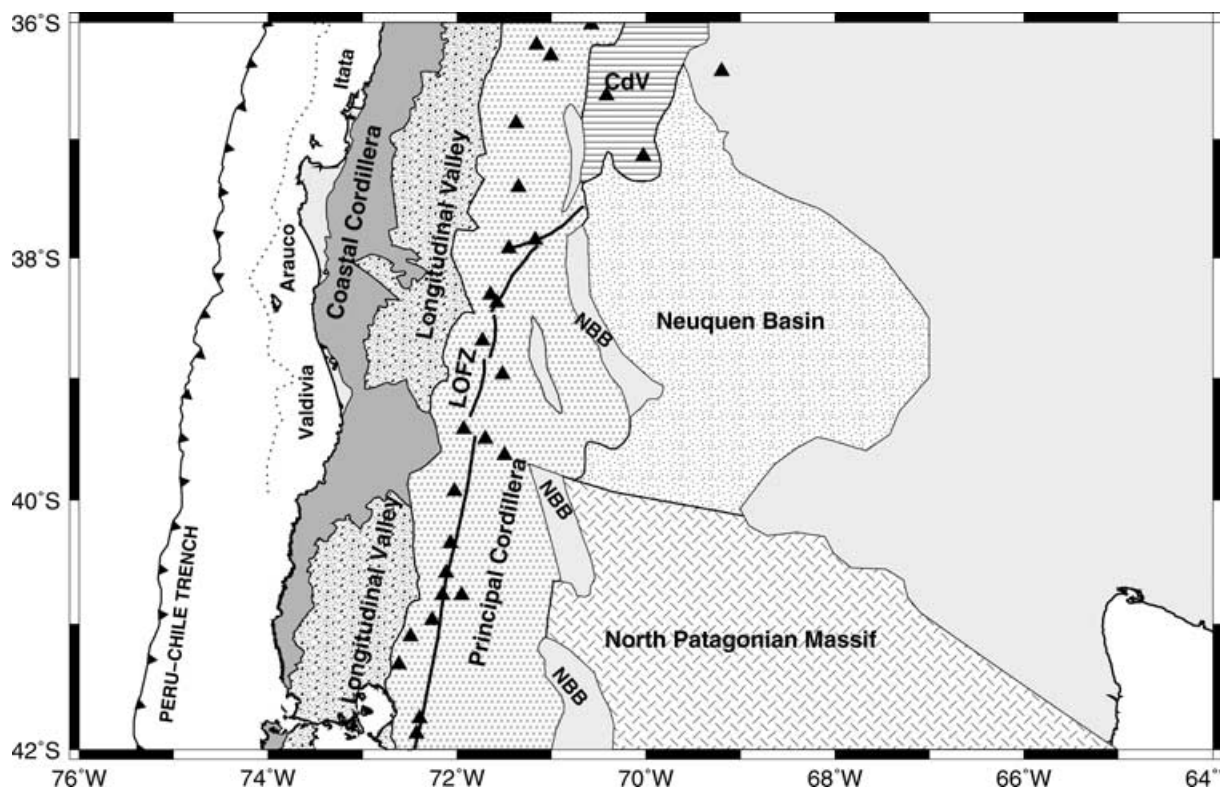


Figure 2. Morphotectonic units of the study area based on Melnick & Echtler (2006). CdV stands for Cordillera del Viento massif, LOFZ for Liquiñe-Ofqui Fault Zone and NBB for Neogene backarc basins. The dotted line marks the western extent of three offshore forearc basins according to Reichert & Schreckenberger (2002).

northern part of the region they are intruded by the synmetamorphic granitoids of the Nahuelbuta Mountains (e.g. Cingolani *et al.* 1991; Glodny *et al.* 2005). The Longitudinal Valley, an active extensional depocentre since the Late Oligocene, is filled with sediments up to several kilometers thick (Jordan *et al.* 2001). It runs through the study area except for an interruption between 39°S and 40°S (Fig. 2).

The volcanoes of the study area belong to the Southern Volcanic Zone (SVZ), extending from 33°S to 46°S. The arc is formed by numerous andesitic to basaltic stratovolcanoes and hundreds of minor eruptive centres (López-Escobar *et al.* 1995). North of ~39°S, the Jurassic to Miocene deposits are locally preserved and, together with the volcanic rocks, form intra-arc volcano-sedimentary basins (Mpodozis & Ramos 1990; Jordan *et al.* 2001). The volcanic arc south of 38°S is superimposed on the igneous rocks of the North Patagonian Batholith that are Cretaceous-Late Tertiary in age (e.g. Cingolani *et al.* 1991). The arc is aligned along the Liquiñe-Ofqui Fault Zone (LOFZ) (Fig. 2), which is a dextral strike-slip fault system cutting the upper plate for more than 1000 km between 38°S and 46.5°S (Hervé 1994; Cembrano *et al.* 2000).

The oceanic plate subducting under the continental plate within the study area originates at two different spreading centres. North of 38°S, crust formed at the East-Pacific Rise at least 35 Myr ago is subducting, while the oceanic crust between 40°S and the Chile triple junction at ~46°S was generated within the past 20 Myr at the Chile Ridge (Herron 1981). This results in a southward decrease in the age of the oceanic lithosphere being subducted at present. The northern limit of the crust generated at the Chile Ridge is defined by the Agassiz Fracture Zone (FZ) at ~40°S, an extension of the Valdivia FZ (Herron 1981; Tebbens *et al.* 1997). However, the age

offset is already pronounced across the Mocha FZ located north of the Agassiz FZ at ~39°S (Fig. 1).

3 GRAVITY DATA AND GRAVITY FIELD

The offshore free-air anomaly and onshore Bouguer anomaly were used in the forward modelling. The offshore gravity data were collected in the framework of the experiment 'Subduction Processes Off Chile' (SPOC). The ship-borne gravity data were measured along profiles (grey lines in Fig. 3) and merged with the KMS global satellite altimetry-derived gravity data set (KMS2001: Andersen & Knudsen 2001) where ship-borne data were not measured (Götze *et al.* 2002; Schreckenberger *et al.* 2002).

The onshore gravity database includes more than 54 000 data, all usually measured along main roads with an average spacing of 3–5 km (Fig. 3). The database is a compilation of data from several different sources measured over the past 30 years (Table 1), thus creating a rather irregularly distributed data set. Around 2000 stations were occupied along the South American Lithosphere Transect between 36° and 42°S from the Pacific to the Atlantic coastlines within the framework of the collaborative research program SFB267 'Deformation Processes in the Andes' (MIGRA 2000 field campaign; Schmidt & Götze 2006). Some 40 000 gravity stations in Argentina were released by Repsol-YPF (Yacimientos Petrolíferos Fiscales, Argentinean National Oil Company). These data were reprocessed, homogenized and merged with the above-mentioned data set (Wienecke 2002). The 12 300 remaining and, for the modelling, most important stations cover mainly the forearc/arc region (dashed box, Fig. 3). These data were compiled as part of a Volkswagen

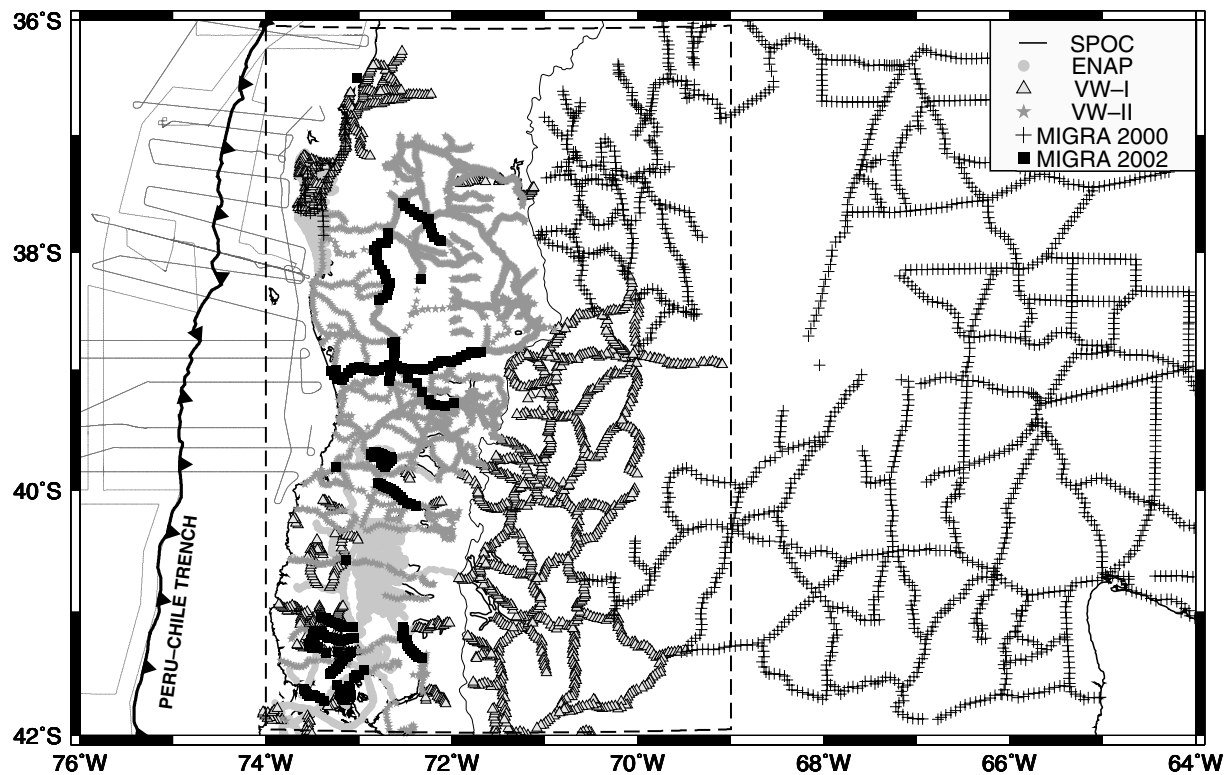


Figure 3. Gravity data used for the modelling. The offshore lines represent the ship-tracks along which the ship-borne data were measured. The light grey points represent the ENAP data, the darker grey stars and triangles are stations from the VW data set, the crosses are from the MIGRA 2000 campaign, and the control measurements (MIGRA 2002 campaign) are shown by black squares. The data located within the dashed-frame were the subject of the detailed analysis described in this paper.

(VW) Stiftung Project (Araneda *et al.* 1999a,b) and were provided under an agreement. The VW data set is comprised of several data files, including data from Empresa Nacional del Petroleo (ENAP).

3.1 Analysis of the gravity data

The different techniques, instrumentation and data processing methods used to acquire each data set resulted in some severe problems within and inconsistencies between the different data files. In an attempt to identify the cause of the problems within the database, a detailed analysis of the data was performed. For the purposes of analysis, the VW data set was divided into the three main parts (ENAP, VW-I and VW-II). The absolute gravity values, heights and Bouguer anomaly values were compared along identical profiles common to the different data files. Discrepancies were found in all three tested quantities (absolute gravity values, station elevations and the Bouguer anomaly), meaning that all data sets had random problems in all three values. Thus, the cause of the differences was not simply a shift resulting from the use of different gravity datums.

The largest misfits in the Bouguer anomaly data of ~ 20 mGal, reaching at times 30 mGal, occur in the southern part of the forearc region (south of 40°S) where the Bouguer anomaly values reach ~ 60 mGal. The area of the gravity maximum is located in the forearc where it is influenced by the subducting plate the most and is thus a very crucial region for the modelling and interpretation. Also, all additional geophysical data are concentrated in the forearc region between 36°S and 39°S . Since the density model extends to 42°S , the large data discrepancies found in the southern forearc region had to be fixed before modelling began. However, without supplementary

information (measurement records etc.), which unfortunately is not available for a large part of the database, this task was impossible. Hence, some of the critical profiles had to be remeasured. Control measurements were conducted in Chile during the MIGRA 2002 field campaign, during which 220 stations on 15 selected profiles were remeasured (black squares, Fig. 3) (Tašárová 2004). After the remeasurements and additional analysis of the database, ~ 500 of the 12 300 stations with obvious errors were removed. The updated VW data set was then merged with the MIGRA 2000 data set. All onshore data were reprocessed and the complete Bouguer anomalies (i.e. including topographic correction) were computed. For the calculation of the Bouguer anomalies, a spherical Bouguer cap with a radius of 166.7 km and correction densities of 2.67 and 1.64 Mg m^{-3} for onshore and offshore areas, respectively, were used (Schmidt & Götze 2006). The correction density of 2.67 Mg m^{-3} roughly corresponds to an average density of the surface granitic rocks and it is a standard density value that enables various Bouguer anomaly data sets to be more easily compared and merged (Hinze 2003 and references therein). The normal gravity formula of 1967 (e.g. Torge 1989) was applied and all the stations were tied to the IGSN71 gravity datum (Morelli *et al.* 1974). For onshore topographic corrections, the GLOBE Digital Elevation Model (DEM) (GLOBE Task Team and others 1999) with a spatial resolution of $1 \text{ km} \times 1 \text{ km}$ (30 arc-sec) was used, while offshore corrections used the ETOPO 5 DEM (NOAA 1988) combined with data from Zapata (2001), also at a resolution $1 \text{ km} \times 1 \text{ km}$. Due to uncertainties in the measurements of station elevation in the original data, the onshore Bouguer anomaly was calculated with an estimated accuracy of 5–10 mGal (Tašárová 2004).

Table 1. The number of gravity stations and the bodies responsible for their collection

Source	Location	Number
MIGRA 2000	Argentina	2000
MIGRA 2002	Chile	220
YPF	Argentina	~40,000
ENAP	Chile	5900
VW Project	Chile/Argentina	6400
Total		~54,520

3.2 Gravity field

The free-air anomaly offshore and Bouguer anomaly onshore, calculated based on the improved database, show notable variations in the trench-parallel anomalies in the forearc region. The variations are particularly obvious along the coastline and thus, three forearc segments have been identified (Fig. 4). These segments are also evident in the local geoid and in the geology (Hackney *et al.* 2006).

The northern Arauco-Lonquimay segment (36°S–39°S) is characterized by a pronounced gravity high of 60–90 mGal along the coast, particularly over the Arauco peninsula, whereas the volcanic arc and the backarc are associated with gravity lows of down to –180 mGal. The offshore minimum of some –120 mGal in the free-

air anomaly is related to the trench. The average free-air anomaly offshore reaches ~35 mGal near the coastline, but there are two areas characterized by higher values. One is around the Arauco peninsula and north of it (Santa María Island, ~37°S), where the free-air anomaly reaches 50–70 mGal. Similarly, the second region, located around Mocha Island (~38.3°S), is characterized by an average of ~60 mGal (Figs 1 and 4).

The middle Valdivia-Liquiñe segment (39°S–40°S) is characterized by a low gravity of –30 to 0 mGal in the onshore forearc and an increase in the minimum gravity values associated with the main arc and the backarc to some –80 mGal. Also, the minimum in the offshore free-air anomaly characterizing the trench has, compared to the northern segment, more positive values of –60 mGal, reflecting the shallower bathymetry of the trench south of 39°S (Reichert & Schreckenberger 2002). The free-air anomaly in the offshore forearc reaches –20 mGal on average, with a local minimum of –45 mGal close to the coastline (Valdivia sedimentary basin).

The southern Bahía-Mansa-Osorno segment (40°S–42°S) is characterized by a forearc gravity high of 50–80 mGal with a longitudinal offset compared to the northern segment. Thus, the gravity high is centred on the western part of the Longitudinal Valley rather than the coastline. The minimum Bouguer gravity anomaly related to the main arc and the backarc regions reaches values of –100 mGal. The offshore free-air anomaly is characterized by values near 0 mGal on

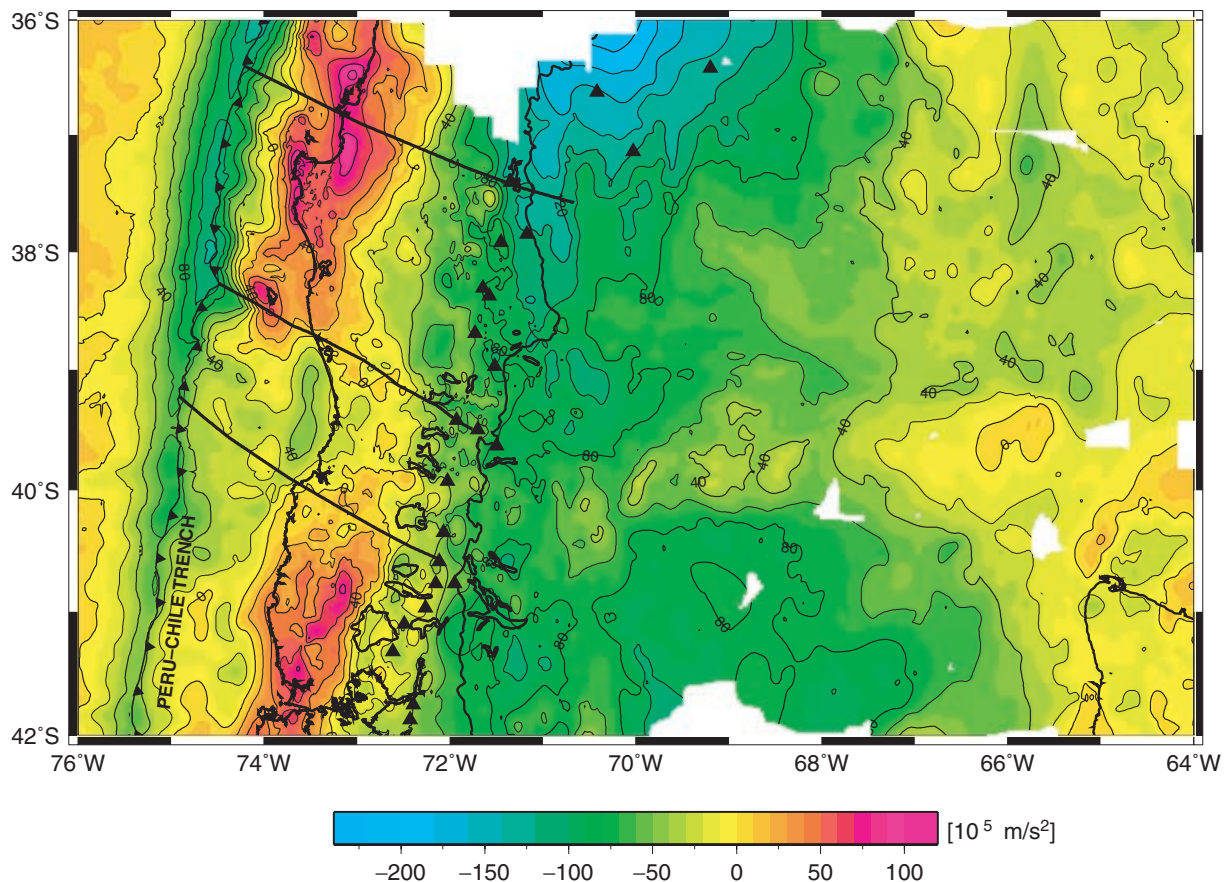


Figure 4. The Bouguer anomaly onshore and the free-air anomaly offshore. The two pronounced maxima reach values of ~80 and ~60 mGal in the northern and southern parts, respectively, of the forearc region. The forearc of the middle segment is dominated by negative values both offshore and onshore. The negative Bouguer anomaly characterizes the entire arc and backarc regions, with a minimum of about –180 mGal in the northern part of the arc. A negative free-air anomaly of –100 mGal is observed at the trench in the northern part of the study area, which increases southwards to about –50 mGal. Offshore areas not covered by the SPOC experiment are filled with KMS global free-air anomaly data (Andersen & Knudsen, 2001). The segments identified in the study area are indicated by oblique black lines and the black triangles mark the active volcanic arc.

average, with only the area above the trench having lower values of some -70 mGal (Fig. 4).

4 MODEL CONSTRUCTION

The density model was constructed using the Interactive Gravity and Magnetic Application System (IGMAS). It is a tool developed for the interpretation of measured gravity and magnetic fields using forward modelling. It works by means of a numerical simulation of underground structures that are described as closed polyhedrons of constant density/susceptibility (e.g. Götze 1976; Götze 1984; Götze & Lahmeyer 1988). 3-D structure is achieved in IGMAS by defining several vertical 2-D planes on which geological bodies are geometrically defined in the form of polygons (Schmidt 1996). The 2-D vertical planes are connected via triangulation, thereby forming 3-D structure. The geometry of the geological bodies is defined along the 2-D cross-sections, whereas the geometry in the area between them is interpolated. Hence, in order to perform more detailed modelling, a greater number of 2-D cross-sections must be included. The 3-D density model to be presented was, therefore, developed using 28 parallel east–west cross-sections. The distance between these cross-sections is 20 km in the northern part of the working area where other geophysical measurements are concentrated, and 30 km south of 40°S where no other geophysical constraints are available (Fig. 5). The density model consists of about 40 bodies that represent the following geological units: ocean, the oceanic crust and lithospheric mantle of the subducting slab, continental crust (three layers), continental lithospheric mantle and asthenospheric mantle (Fig. 6). As input, IGMAS uses gravity data either in a station file with the locations of the measurement points, or a gridded data file.

Since the data coverage in the region studied is good, although the stations distribution is uneven (Fig. 3), in order to retain the measured information, a station file was used in this case. The gravity effect of the modelled structures is calculated at the measurement points, while the gravity anomalies in the areas with no measured data are linearly interpolated.

4.1 The reference model

The measured gravity is influenced by the mass of the entire Earth, whereas in density models, only a part of the Earth is considered (typically the upper tens to hundreds of kilometres.) Hence, the modelled space does not include all the masses below and around the measurement points. For this reason, the levels of the observed and modelled values are not identical. Additionally, the use of absolute density values, which are directly comparable with seismic velocities, increases the offset of the observed and modelled values even more (Kirchner 1997). This offset can be computed and removed automatically by the IGMAS program. In this case, the modelled density values are always relative to a chosen reference density (e.g. 2.67 Mg m^{-3} , which is usually used for calculating Bouguer anomalies). Another possibility to remove the offset is to introduce a reference model in the background of the density model (Kirchner 1997; Ebbing *et al.* 2001). The reference model has only a few layers representing the structure of the ‘normal Earth’, but extends to the same depth as the 3-D model and its densities are assigned negative values. The modelled densities are then relative to this reference model.

The selected reference model (Table 2) is constrained by the studies of Christensen & Mooney (1995) and Rudnick & Fountain

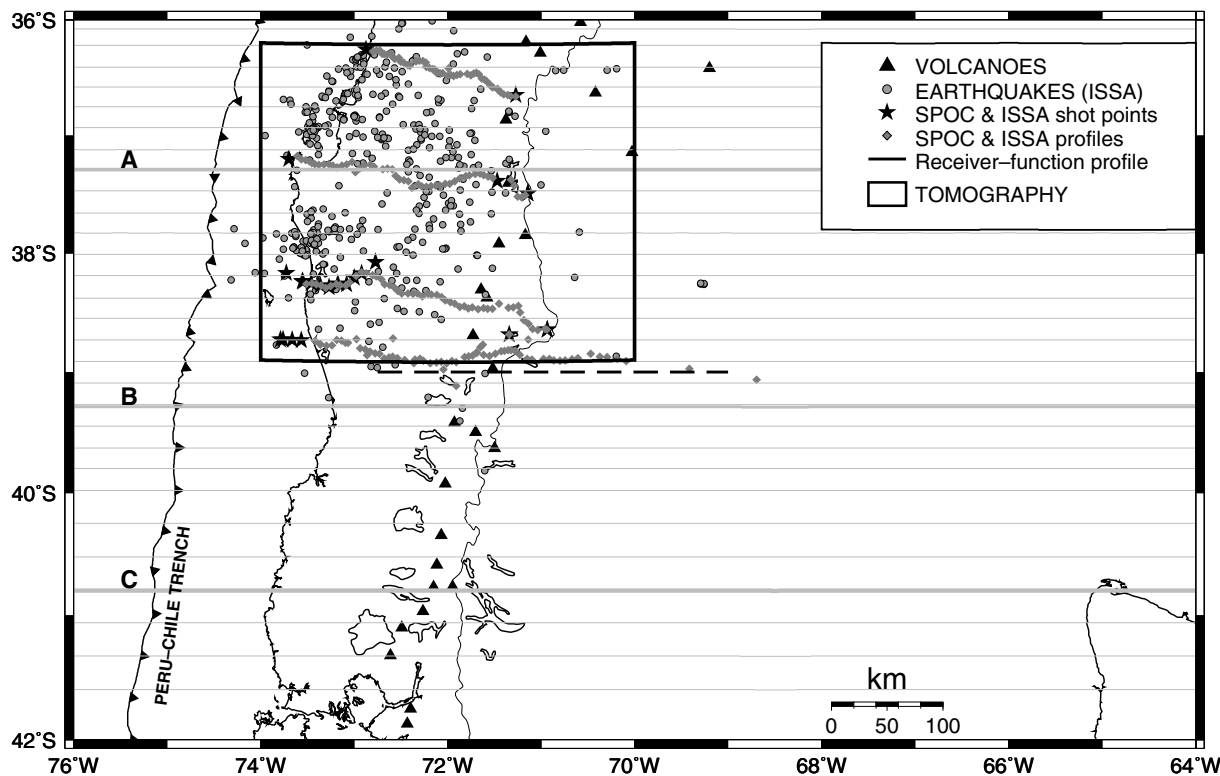
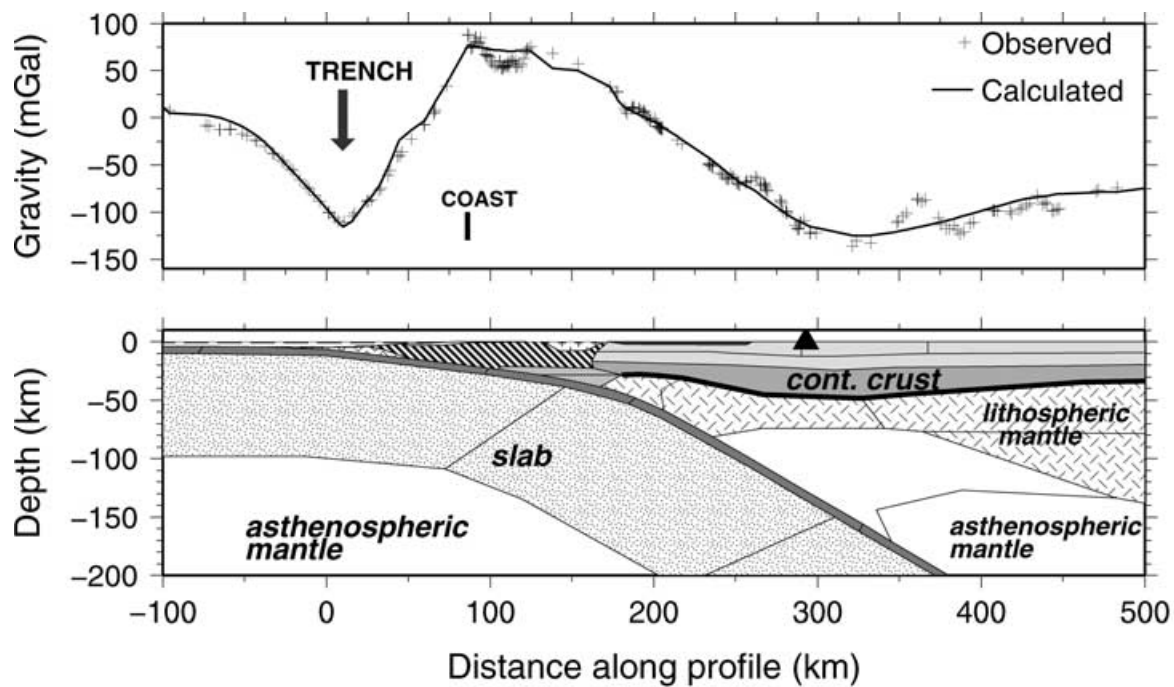


Figure 5. Constraining data used for the gravity modelling. The black frame marks the area of the local earthquake tomography model and the dashed line marks the location of the receiver-function profile (see legend, top right corner). The cross-sections of the 3-D model are also marked by the thin grey east–west lines. The profiles A, B and C (thick lines) are shown in Figs 6, 7, 10 and 11.



Densities [Mg/m^3]

 (1.03)	oceanic water	 (3.15–3.2)	unidentified body
 (2.95–3.45)	oceanic crust	 (2.65–2.88)	cont. upper & middle crust
 (2.4)	forearc basins	 (3.1)	continental lower crust
 (2.6)	active wedge	 (3.23–3.32)	continental lithosphere
 (2.85)	accretionary complex	 (3.24–3.41)	oceanic lithosphere
 (2.6)	sediments (Long Valley)	 (3.34)	oceanic and continental asthenosphere
 (2.72)	granites		

Figure 6. A representative cross-section from the 3-D density model. This cross-section at $\sim 37.3^\circ\text{S}$ (profile A, Fig. 5) crosses the Arauco Peninsula, which is characterized by the pronounced positive gravity anomaly. The upper part of the image shows the observed gravity values (crosses) and the modelled gravity curve (black line). The lower part shows the structures within the density model. The position of the volcanic arc is indicated by the black triangle. The observed gravity data from the original station file are projected onto a profile from a 4 km wide stripe (2 km north and 2 km south of the profile). In areas lacking data, the calculated gravity anomaly is interpolated by the IGMAS program. The calculated anomaly is shown here as a line, but in the modelling software it is calculated at the station locations. The same holds true in Figs 7, 10 and 11.

Table 2. Depth and density parameters of the reference model used for gravity modelling

Layer	Depth (km)	Density (Mg m^{-3})
Upper crust	0–20	–2.68
Lower crust	20–40	–3.0
Upper mantle	40–300	–3.35

(1995). The continental crust is divided into two layers with densities of -2.68 and -3.00 Mg m^{-3} for the upper and lower layer, respectively. The thickness of each crustal layer is 20 km, thus producing 40 km thick crust with an average density of -2.84 Mg m^{-3} , which is consistent with the global average (Christensen & Mooney

1995) and with the crustal structure of the ‘Arcs’ and ‘Shields and Platforms’ analysed by Rudnick & Fountain (1995). The mantle density of -3.35 Mg m^{-3} , which extends to a depth of 300 km, corresponds to the average upper-mantle value (Poudjom Djomani *et al.* 2001).

4.2 Constraining data

The density model is constrained in the northern segment by geophysical data from two geophysical experiments—‘Integrated Seismological experiment in the Southern Andes’ (ISSA 2000) and ‘Subduction Processes Off Chile’ (SPOC) (Fig. 5). Data from the ISSA 2000 experiment provide the first accurate image of the

Benioff zone in the Southern Andes (Bohm *et al.* 2002). Other results from this experiment include a local earthquake tomography model (Bohm 2004), a 2-D refraction seismic profile at 38.8°S (ISSA profile) (Lüth *et al.* 2003) and a receiver-function study at ~39°S (Yuan *et al.* 2006). The SPOC project was a combined offshore–onshore, active-passive survey focused on the forearc and the seismogenic-coupling zone (Krawczyk *et al.* 2003, 2006). The onshore experiment comprised a 54 km near-vertical incidence reflection profile and three wide-angle reflection/refraction seismic profiles that provide *P*-wave velocity images along east–west transects at ~36.2°, 37.2° and 38.2°S (SPOC north, middle and south profiles) (Lüth *et al.* 2004; Krawczyk *et al.* 2006). The near-surface structure of the density model is based on a geological model for this region (Echtler & Melnick, personal communication, 2002).

5 INDIRECT DENSITY DETERMINATION

The density and geometry of modelled structures are the primary input parameters for gravity modelling, which means that all available information which constrain these parameters must be used. Seismic velocities are one of the most important geophysical measurements for providing fundamental information about rock densities and deeper crustal/lithospheric structures. Based on studies of the physical and chemical properties of various rocks, including laboratory measurements of densities and velocities, relationships between both quantities have been derived (e.g. Birch 1961; Ludwig *et al.* 1970; Gardner *et al.* 1974; Christensen and Mooney 1995). These empirical velocity–density curves, or linear relationships, provide a mean density of a particular rock unit, which internally may be quite variable. Nevertheless, a density model always approximates the real structures in a simplified way, with larger units extending sometimes over tens of kilometres (in distance and depth) that are characterized by a single (constant) density value. Therefore, although a significant scatter around the mean value is associated with empirical velocity–density relationships (Ludwig *et al.* 1970;

Barton 1986), they may be applied to larger-scale lithospheric models. In models with a characteristic scale of several tens or hundreds of kilometres, a stronger correlation between velocities and densities is expected, particularly in the middle and lower crust. This is due to the decreased scattering related to the increased wavelength of the model (Kozlovskaya *et al.* 2004a, b and references therein).

In this study, the linear relationships of Sobolev & Babeyko (1994) and Christensen & Mooney (1995) were used to determine the density values from the *P*-wave velocities of SPOC and ISSA models (Table 3). The results obtained were also compared to the densities calculated based on the Nafe–Drake curve (Ludwig *et al.* 1970). In addition, the relationships of Gardner *et al.* (1974) and Wang (2000) were used to convert seismic velocities to densities for the sedimentary rocks (Section 6.4). The approach of Sobolev & Babeyko (1994) takes into account *in situ* temperature and pressure conditions and, in three steps, converts measured velocities into *in situ* densities (e.g. Kirchner 1997; Ebbing 2002). The pressure and temperature are input parameters required for the density calculation, hence a user is able to calculate the densities for regions characterized by different thermal states. If no thermal model exists for a study region, temperatures can be calculated based on the surface heat flow density distribution or radiogenic heat production using the heat conduction equations (e.g. Turcotte & Schubert 2002). The *in situ* temperatures can also be determined using Ranalli's (1997) graphical relationship between surface heat flow, temperature and depth. Similarly, pressure can be estimated as a function of depth, standard density and overpressure factor (Turcotte & Schubert 2002). Although heat flow measurements from the area studied are sparse and limited to the volcanic arc, the forearc and volcanic arc are, similarly to the Central Andes, assumed to be characterized by different temperature conditions (Hamza & Muñoz 1996). Thus, the approach of Sobolev & Babeyko (1994) was found to be more appropriate than other empirical relationships in which the temperature cannot be changed. However, this method is valid for velocities greater than 6.05 km s⁻¹ and should only be used for crustal rocks.

Table 3. Densities for particular depths derived from *vp* velocities using the approach of Sobolev & Babeyko (1994) (S&B), Christensen & Mooney (1995) (C&M) and Ludwig *et al.* (1970).

Depth (km)	<i>V_p</i> (km s ⁻¹)	rho (S&B) colder (Mg m ⁻³)	rho (S&B) warmer (Mg m ⁻³)	rho (C&M) (A) (Mg m ⁻³)	rho (C&M) (B) (Mg m ⁻³)	rho (C&M) (C) (Mg m ⁻³)	rho (Ludwig) (Mg m ⁻³)
10	6.05	2.64	2.65	2.73	2.74	2.72	2.73
10	6.1	2.66	2.67	2.75	2.75	2.74	2.74
10	6.2	2.70	2.71	2.79	2.78	2.77	2.76
20	6.3	2.76	2.78	2.82	2.82	2.81	2.78
20	6.4	2.81	2.82	2.85	2.85	2.85	2.81
20	6.5	2.85	2.87	2.89	2.88	2.88	2.83
30	6.6	2.91	2.93	2.94	2.93	2.94	2.86
30	6.7	2.95	2.97	2.97	2.96	2.98	2.89
30	6.8	2.99	3.02	3.00	2.99	3.02	2.91
30	6.9	3.04	3.07	3.03	3.02	3.06	2.94
30	7.0	3.09	3.12	3.06	3.05	3.10	2.97
40	7.1	3.15	3.18	3.12	3.10	3.17	3.00
40	7.2	3.20	3.23	3.15	3.13	3.21	3.03
40	7.3	3.24	3.28	3.18	3.16	3.25	3.06
40	7.4	3.29	3.33	3.20	3.19	3.29	3.09

Notes: The temperatures for the approach of Sobolev & Babeyko (1994) were calculated assuming surface heat flow of 45 mW m⁻² for a colder and 70 mW m⁻² for a warmer region. There are three relationships of Christensen & Mooney (1995), one non-linear and two linear. The non-linear relationship (A) was derived for all rock types (including the upper mantle rocks) and is recommended for crust–mantle sections. One of the linear relationships considers all rock types (B), the other is valid for all rock types excluding the volcanic and mono-mineralic rocks (C).

Table 4. Velocities, densities and for lower-crustal and upper-mantle rocks additionally also V_p/V_s and Poisson's ratios calculated based on published mineralogical composition of the selected rocks using the method of Hacker & Abers (2004).

Depth (km)	Rocks	V_p (km s ⁻¹)	rho (Mg m ⁻³)	V_p/V_s ratio	Poisson's ratio
10	Granit	6.20	2.67	—	—
10	Granit 2	6.34	2.67	—	—
20	Granodiorite	6.35	2.78	—	—
20	Tonalite	6.5	2.82	—	—
40	Gabbro	6.67–6.85	3.05–3.06	1.8–1.82	0.28–0.29
40	Olivin-granulite	7.22	3.05	1.84	0.29
40	Garnet-granulite facies	7.4	3.31	1.79	0.27
40	20per cent serpentinized periodite	7.54	3.13	1.81	0.28
40	30per cent serpentinized periodite	7.35	3.09	1.84	0.29
40	Lherzolite (29per cent serpentine)	6.99	2.92	1.86	0.30
40	95per cent serpentinized periodite	5.88	2.72	2.19	0.37
40	Lherzolite D	7.8	3.3	1.76	0.26
40	Lherzolite M	8.01	3.35	1.75	0.26
50	Dunite	8.23	3.32	1.75	0.26

Notes: For both velocities and densities, the pressure and temperature values change at different depths. The temperature is calculated according to the heat conduction equation (e.g. Turcotte & Schubert 2002) with an average surface heat flow value of 45 mW m⁻², thermal conductivity of 3 Wm⁻¹ K⁻¹, surface temperature of 25 °C, heat production of 1.5 µWm⁻³ and thickness of the radiogenic layer of 10 km. The pressure was calculated for each depth after Turcotte and Schubert (2002) using a standard density of 2.8 Mg m⁻³ and overpressure factor of 1. For instance, the pressure and temperature at 40 km reach values of 1.1 GPa and 474 °C, respectively.

Table 3 shows that there are significant differences (0.05 – 0.2 Mg m⁻³) in the calculated density values obtained from the various empirical relationships. This leads to a wide range of possible densities corresponding to the observed seismic velocities. Therefore, the choice of a proper velocity-density relationship is one of the major problems related to developing combined gravity and seismic models. Moreover, it has been demonstrated by many studies (e.g. Christensen & Mooney 1995; Rudnick & Fountain 1995; Hacker & Abers 2004) that seismic velocities and densities are significantly dependent on P/T conditions, water content, rock composition and metamorphic grade. The silica and garnet content is particularly significant (Kozlovskaya *et al.* 2004b and references therein; Tassara 2006). With this in mind, two other approaches were used, both of which allow the P -wave velocity and density values of given rocks to be calculated based on their chemical or mineralogical composition (Sobolev and Babeyko personal communication, 2004; Hacker & Abers 2004). The results obtained (Table 4) show that rocks of different composition might have the same or similar densities but quite different velocities, and vice versa. Therefore, it is very important to account for composition when densities are determined from seismic velocities. However, while it is possible to compile mineralogical/chemical data for rocks typical of smaller regions (several tens of kilometers studied in 3-D, or several hundreds of kilometres studied in 2-D), for large regions, such as the one studied here (~600 km × 600 km), it is more difficult. This is due to the complex nature of large regions and to the availability of published data, which might be limited.

For the purposes of this study, many different densities were tested during the modelling. Additionally, it is important to keep in mind that the velocity models also constrain the geometry of the density model. Thus, the densities used should be derived as a trade-off between the seismic velocities, the depths of seismic interfaces from the velocity models, and gravity anomalies. Ideally, other independent geophysical observations should be used while converting seismic velocities to densities. In this study, the converted densities (Table 5) were tested against the Moho geometry along the receiver-

function profile at ~39°S, which shows the Moho at 40 km depth below the main volcanic arc (Yuan *et al.* 2006). With the receiver-function Moho fixed, the densities of the middle and lower crust that would best match the Bouguer gravity anomalies along this profile were determined. The assigned densities of 2.85 and 3.1 Mg m⁻³ correspond to the average velocities of 6.5 and 7.0 km s⁻¹ for the middle and lower crust, respectively, which is in agreement with the seismic data (Table 5).

Velocities in the upper crust are mainly influenced by the presence of cracks and fractures, possibly filled with fluids, that decrease the observed seismic velocities (Schön 1996 and references therein). Therefore, if a standard density-velocity relationship is applied, the densities obtained might be underestimated (Kozlovskaya *et al.* 2004a). Additionally, the approach of Sobolev & Babeyko (1994) is not valid for upper-crustal rocks with velocities below 6.05 km s⁻¹. Hence, direct density measurements of selected samples from the forearc region of the study area were carried out in order to better constrain the upper-crustal densities in the forearc region (Glodny personal communication, 2004; Tašárová 2004). However, very few samples were measured and, therefore, the results cannot be easily generalized for the forearc region along its entire length. Nevertheless, the direct density measurements complete the density determination since they provide further insight into more local-scale upper-crustal structures (<10 km long, e.g. the Nahuelbuta granites) that are not resolved very well in the seismic profiles.

6 STRUCTURES OF THE DENSITY MODEL

6.1 Forearc

The part of the continental crust closest to the trench, the Coastal Cordillera, comprises a Palaeozoic accretionary complex (Figs 2 and 6). Its density and geometry influence the forearc gravity field the

Table 5. Observed velocities along the SPOC and ISSA profiles are shown with corresponding densities calculated based on the methods of Sobolev & Babeyko (1994) (S&B) and Christensen & Mooney (1995) (C&M).

	SPOC (km s ⁻¹)	ISSA (km s ⁻¹)	Calculated density (S&B) (Mg m ⁻³)	Calculated density (C&M) (Mg m ⁻³)	Employed density (Mg m ⁻³)
Upper crust (forearc)	4.0–6.4	4.5–6.4	–2.79	1.6–2.85	2.4–2.72
Upper crust (arc)	5.8–6.2	5.8–6.2	–2.71	2.64–2.79	2.7
Middle crust	6.5–6.7	~6.6	2.85–2.95	2.89–2.97	2.85
Lower crust (forearc)	6.7–7.2	6.7–7.2	2.95–3.2	2.97–3.15	3.1 (3.15–3.2) ^a
Lower crust (arc)	6.9–7.2	6.8–7.0	3.02–3.23	3.00–3.15	3.1
Oceanic crust	6.6–7.2	6.8–7.2	–	2.94–3.15	2.95–3.1
Upper mantle (forearc)	≥7.2	≥7.2	–	≥3.15	3.23

Notes: The calculated values are compared with the densities employed in the 3-D model. The densities from the Sobolev and Babeyko approach were computed assuming lower temperatures for the forearc and higher for the volcanic arc. Secondly, a non-linear relationship of Christensen & Mooney (1995) for all rock types (including the upper mantle) was applied (see also Table 3). The velocities and densities of the oceanic crust correspond to depths of ~20 to 45 km. The relationship of Sobolev & Babeyko (1994) is limited to continental crustal rocks with velocities greater than 6.05 km s⁻¹. Therefore, the velocities lower than that of upper mantle and oceanic velocities were not converted into densities and the missing values are denoted by hyphens.

^aThe densities of the forearc high-density body.

most. The Palaeozoic accretionary complex consists of the Western Series high P/T rocks that dominantly comprise metabasites and metasediments, but also contain metapelites, metapsammites and serpentinites (Willner *et al.* 2004; Glodny *et al.* 2005). The accretionary complex north of ~38.5°S also includes the low P/T Eastern Series rocks (metasediments, metapelites and metapsammites), which completely lack metabasites. Thus, the density of the northern part of the Palaeozoic accretionary complex should be lower than the southern part. The density values obtained for the Palaeozoic accretionary complex from direct density measurements are ~2.75 Mg m⁻³ in the northern and ~2.8 Mg m⁻³ in the middle and southern segments (Tašárová 2004; Glodny, personal communication, 2004). These densities are derived assuming that the accretionary complex in the middle and southern segments contains 20 per cent more metabasites than the northern part. However, the selected measured samples are very limited and unlikely to characterize an area extending over ~600 km. In addition, because of the lack of geological and geophysical constraining data, it is impossible to accurately estimate the composition of the deeper parts of the Palaeozoic accretionary complex in the southern two segments. For this reason and in order to maintain the simplicity of the model, a constant density of 2.85 Mg m⁻³ was used for the entire Palaeozoic accretionary complex (north and south of 38.5°S). In the northern segment, this value is consistent with the seismic and gravity data. The forearc north of 38.5°S, where there is a pronounced gravity high, requires higher densities than the values inferred from the direct density measurements (Fig. 7). Additionally, the four seismic profiles available from the study area are all located in the northern segment. They provide the only control on the deeper part of the Palaeozoic accretionary complex and none of them has revealed a low-velocity feature within the forearc. The *P*-wave velocities in the Palaeozoic accretionary complex change from ~6–6.2 km s⁻¹ at depths of 0–10 km to ~6.7–7.1 km s⁻¹ in its lower parts at 10–30 km depth (Lüth *et al.* 2003; Krawczyk *et al.* 2006). According to the Sobolev & Babeyko (1994) approach, the densities corresponding to the average *P*-wave velocities of 6.1 and 6.9 km s⁻¹ in the upper and lower parts of the crust are 2.66 and 3.04 Mg m⁻³ for colder regions, respectively (Table 3). Since the accretionary complex is represented in the model by only one body, it was assigned

a density of 2.85 Mg m⁻³. This value corresponds to an average *P*-wave velocity of 6.5 km s⁻¹ at 20 km depth, which is in agreement with the seismic data.

The Palaeozoic accretionary complex north of 38.5°S includes the granitoids of the Nahuelbuta Mountains (Fig. 5), which are characterized by a lower density (the modelled value is 2.72 Mg m⁻³). Beneath this intrusive body, which occurs only in the northern segment at 36°S–39°S, a high-density structure has been modelled in order to reproduce the observed gravity field (at short wavelength). This body is also included only in the northern segment and it can be explained as a mafic lower crustal residue associated with the intruded granitoids of the Nahuelbuta Mountains. However, based on the *P*-wave velocities (7.0–7.3 km s⁻¹) and the assigned densities (3.15–3.2 Mg m⁻³), this structure could also represent serpentinized mantle, as interpreted in the tomography model (Bohm 2004) (see Section 7.2).

The upper crust of the forearc region also includes the Longitudinal Valley that was modelled as a sedimentary basin and whose sediment infill locally exceeds 3 km (Jordan *et al.* 2001). The *P*-wave velocities characteristic for the Longitudinal Valley reach values of ~5.5 km s⁻¹ along all of the SPOC and ISSA profiles, except for the SPOC northernmost profile, where the velocities reach a maximum of ~4 km s⁻¹ (Lüth *et al.* 2004; Krawczyk *et al.* 2006). The calculated densities for sedimentary rocks (Gardner *et al.* 1974; Wang 2000) corresponding to these velocities reach values of 2.4–2.65 Mg m⁻³ for the lower and higher *P*-wave velocities, respectively. The bulk density should, however, also represent the volcanic rocks within the Longitudinal Valley that between 36°S to 39°S constitute equal or greater volumes than the sedimentary rocks (Jordan *et al.* 2001), and hence a higher density of 2.6 Mg m⁻³ was used. The thickness of the sediments in the Longitudinal Valley, based on the gravity model, is 1–4 km.

6.2 Arc and backarc

The continental crust, apart from the Palaeozoic accretionary complex representing the Coastal Cordillera in the forearc, is modelled by three layers—the upper, middle and lower crust. During the forward density modelling, the crustal layers were divided into many

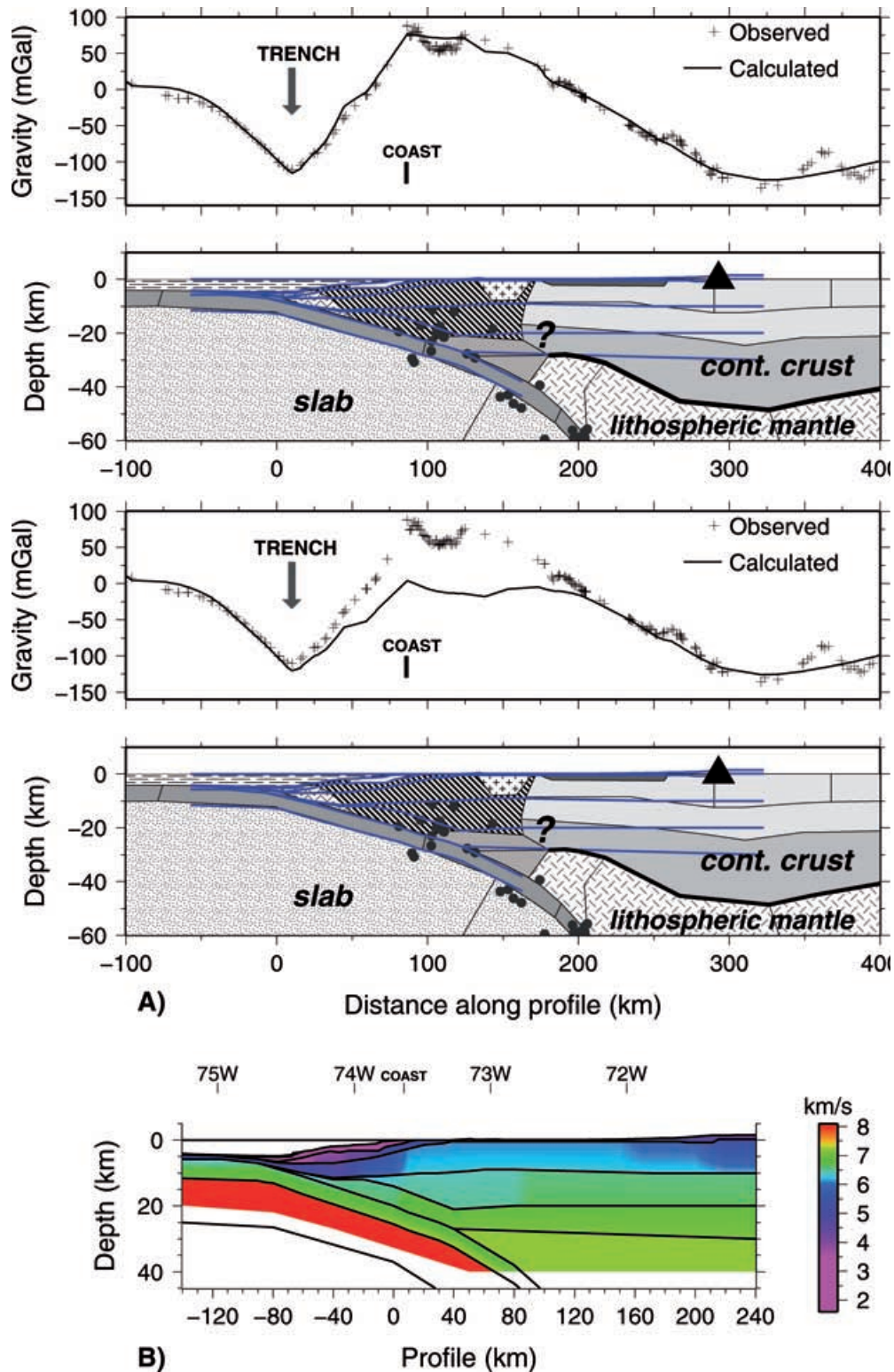


Figure 7. (A) Detailed image of the forearc structures along the same profile as shown in Fig. 6. The gravity effect with a reduced density of 2.75 Mg m^{-3} in the accretionary complex, as determined from the geochemical data and the direct density measurements, is shown for comparison in the lower image. The slab is constrained by seismic data (blue lines, SPOC middle profile shown in B) and seismicity (black dots; Bohm *et al.* 2002). The figure shows that the mass deficit associated with the low-density accretionary complex cannot be compensated by a shallower slab ($\sim 10 \text{ km}$ required). Therefore, the accretionary complex in the density model was assigned a density of 2.85 Mg m^{-3} (top image), which matches both gravity and seismic data. The continental Moho is marked by a thick black line, but does not continue further to the slab due to the uncertainties in the nature of the unidentified high-density body. The position of the volcanic arc is indicated by the black triangle. See legend in Fig. 6. (B) SPOC middle seismic profile (Lüth personal communication, 2007). The last velocity isoline, 7.2 km s^{-1} , was assumed by Krawczyk *et al.* (2006) to represent the continental Moho. However, the continental Moho was not imaged in the seismic profile. This explains the large misfit under the volcanic arc between the gravity-derived Moho and the interpreted Moho in the seismic model.

blocks in both N–S and W–E directions to allow the introduction of along- and across-strike density variations. However, in order to keep the model simple, lateral density variations in the middle and lower crust were in the final model not included. In reality, while heterogeneity is likely, there is insufficient data to better constrain the density variations in the middle and lower crustal layers. The determination of the densities in the middle and lower crust is described in the previous section. The upper crust includes some along- and across-strike changes, the nature of which can be constrained by surface geology (Melnick & Echtler 2006). The upper-crustal density in the northern segment is 2.7 Mg m^{-3} underneath the volcanic arc, which is consistent with the seismic models (Table 5). In the backarc, the density decreases to 2.68 Mg m^{-3} . In the middle and southern segment the densities associated with the volcanic arc increase to 2.75 and $2.8\text{--}2.82 \text{ Mg m}^{-3}$, respectively. This increase is attributed to the magmatic additions towards the south (e.g. fig. 30.2 of Melnick & Echtler 2006). Nevertheless, factors such as porosity, cracks and fluids in the upper crust, were not considered in this large-scale lithospheric model.

6.3 Continental mantle

Below the Moho and until a depth of about 180 km, the lithosphere is divided into four blocks with densities ranging from 3.23 to 3.32 Mg m^{-3} . The minimum lithospheric density is located under the forearc and main volcanic arc. This value reflects the possible hydration/serpentinization of the forearc mantle. The assigned density of 3.23 Mg m^{-3} would correspond to a mixture of ~ 20 per cent fully serpentinized mantle with ~ 80 per cent of normal mantle, if the fully serpentinized mantle and normal mantle have densities of 2.7 and 3.35 Mg m^{-3} (Hacker & Abers 2004), respectively. The high temperatures expected for the volcanic arc region would also reduce seismic velocities and hence densities. The density of the lithosphere increases towards the backarc from 3.28 to 3.3 Mg m^{-3} . Below the amalgamated colder backarc blocks to the east, the lithospheric density reaches 3.3 Mg m^{-3} at depths of $40\text{--}80$ km and 3.32 Mg m^{-3} at depths of $\sim 80\text{--}180$ km. These densities are consistent with values typical for mantle periodites (lherzolite, harzburgite) (Hacker *et al.* 2003; Hacker & Abers 2004). However, the densities were modified according to the observed gravity data and with respect to the reference mantle density of $\sim 3.35 \text{ Mg m}^{-3}$. The depth of the lithosphere/asthenosphere boundary (LAB) changes from about 80 km under the volcanic arc to ~ 180 km under the easternmost part of the region. However, unlike in other studies (Tassara *et al.* 2006), the LAB is not constrained thermally, nor by seismic data since there is insufficient heat flow data south of $\sim 30^\circ\text{S}$ (e.g. Hamza & Muñoz 1996; Artemieva 2006 and references therein) and seismic models extending to greater depths do not exist.

The asthenosphere is modelled in two parts. The part closer to the subducting slab at depths between 80 and 130 km represents a deeper portion of the mantle wedge with high temperatures (above $\sim 1200^\circ\text{C}$) and partial melts (e.g. Stern 2002; Hacker *et al.* 2003; Abers *et al.* 2006). A low density of 3.25 Mg m^{-3} was estimated based on the formula of Lachenbruch & Morgan (1990), assuming the density and temperature in the lithospheric mantle to be $\sim 3.3 \text{ Mg m}^{-3}$ and $\sim 800^\circ\text{C}$, respectively. The rest of the asthenosphere at depths of $180\text{--}300$ km is characterized by a density of 3.34 Mg m^{-3} , a bit lower than the reference model mantle density of $\sim 3.35 \text{ Mg m}^{-3}$.

6.4 Oceanic plate

The depth of the ocean floor is constrained by the bathymetry grid of Zapata (2001) and the ocean water density is set to 1.03 Mg m^{-3} . The bathymetry measured during the SPOC experiment shows that the trench at latitudes $36^\circ\text{S}\text{--}40^\circ\text{S}$ shallows gradually from 5000 m north of 37°S to 4200 m in the south (Reichert & Schreckenberger 2002). It also becomes broader, increasing in width from ~ 50 km in the north to ~ 110 km at $\sim 40^\circ\text{S}$. The trench is filled with a wedge of sediments increasing in thickness from west to east and estimated to be between 1.5 and 2.8 km thick (Reichert & Schreckenberger 2002). The thickness of the trench sediments derived from the density model is on average 2–2.5 km and, together with the active frontally accreted sediments, they are characterized by a density of 2.6 Mg m^{-3} . The offshore forearc is dominated by sedimentary basins (e.g. Mordovojich 1981; Reichert & Schreckenberger 2002) that are included in the model and are also necessary to match the observed gravity anomaly at short wavelengths. All three offshore basins, Itata, Arauco and Valdivia (Fig. 2), were modelled and are characterized by a low density of 2.4 Mg m^{-3} . According to the indirect approach of Wang (2000), this density corresponds to the *P*-wave seismic velocity of $\sim 4 \text{ km s}^{-1}$. Applying the empirical relationship of Gardner *et al.* (1974), such a density corresponds to a *P*-wave velocity of 3.6 km s^{-1} . Hence, the assigned density is in accordance with the seismic data, which in the offshore forearc are characterized by low values of $\sim 4 \text{ km s}^{-1}$ (Lüth *et al.* 2003; Krawczyk *et al.* 2006). The thickness of the sediments in the basins inferred from the gravity model is 1.5–3 km.

The subducting slab is represented by oceanic crust and lithospheric mantle (Fig. 6). The slab consists of several separate blocks in order to incorporate the changing density values of the crust and mantle with depth. The oceanic crust is represented by a one-layer body with a thickness of $\sim 6\text{--}7$ km, as imaged in seismic data (Lüth *et al.* 2004; Krawczyk *et al.* 2006). It is divided down-dip into five blocks, each having a constant density value. A bulk density of 2.9 Mg m^{-3} is assigned to the crust before the plate enters the trench. This density increases step-wise to 3.5 Mg m^{-3} at depths of $200\text{--}300$ km. In general, this density change is related to the mineralogical reactions (basalt/gabbro–blueschist–eclogite transformation) occurring within the oceanic crust with changing *P/T* conditions (Peacock 1993; Stern 2002). Similarly, the mantle lithosphere is divided into several blocks. The density of the oceanic lithosphere increases with depth from 3.28 Mg m^{-3} (3.26 Mg m^{-3} in the south) at the beginning of subduction to 3.41 Mg m^{-3} at a depth of 300 km. This increase reflects the dehydration of the slab (e.g. Stern 2002). The density of the oceanic lithospheric mantle, as well as its thickness, is age-dependent (e.g. Turcotte & Schubert 2002). The thickness of the modelled lithospheric mantle is ~ 90 km in the northern part, where the subducting plate is older (~ 35 Myr) and decreases to ~ 50 km south of 40°S , where the slab is younger (<20 Myr). In addition, based on the modelling results, a density reduction is associated with the changing age and thickness. This reduction occurs south of 40°S where the density of the incoming oceanic lithosphere is 0.02 Mg m^{-3} less than in the north. The density values for the oceanic crust and mantle were estimated using the method of Hacker & Abers (2004). The determination of the densities within the subducting slab could, however, be significantly improved if thermal and corresponding petrological models were available (e.g. Hacker *et al.* 2003; Peacock *et al.* 2004). Since no thermal and petrological models are available for the region studied here, the employed densities only roughly represent the composition of the subducting

slab. Both the oceanic and continental asthenosphere have a density of 3.34 Mg m^{-3} .

7 RESULTS AND DISCUSSION

7.1 Crustal thickness

The modelled continental crust is on average 40–45 km thick, with local extremes of 53–55 km (maximum), and ~25 km (minimum) (Fig. 8). The maximum crustal thickness occurs beneath the Cordillera del Viento massif in the northern segment, where the Bouguer anomaly reaches its minimum values of –160 to –180 mGal, and the elevation is higher than elsewhere (Figs 1, 2 and 4). Similar values of crustal thickness below the volcanic arc were imaged in the local earthquake tomography model (Bohm 2004). The crustal thickness below the volcanic arc decreases gradually from 45 to 50 km in the north to ~35 km in the south. Where the refraction seismic data provide control on the continental Moho (along the SPOC north and ISSA profiles), there is good agreement with the gravity-derived Moho (Krawczyk *et al.* 2006).

The crust beneath the Longitudinal Valley is characterized by thinning relative to the volcanic arc and is 33–38 km thick in the northern segment between 36°S and 39°S. Likewise, thinning of ~10 km relative to the volcanic arc in the northern segment is also

indicated by the results of the local earthquake tomography model (Bohm 2004). In the middle segment, where the Longitudinal Valley is disrupted, the crustal thickness in the forearc region reaches values of 38–45 km, which is similar to the crustal thickness underneath the volcanic arc in this segment. Hence, there is no pronounced thinning of this part of the forearc relative to the volcanic arc. In the southern segment, where there are no seismic data available, the crustal thinning may be supported by and associated with the extension observed in the study area (Jordan *et al.* 2001 and references therein). The observed gravity anomaly of 50–70 mGal in this segment, located above the western half of the Longitudinal Valley, is reproduced in the density model by the minimum crustal thickness of 25–28 km.

If the high density body underneath the Coastal Cordillera (Figs 6 and 7) in the northern segment is assumed to be serpentinized mantle (see following section), then the minimum crustal thickness would occur here and would reach ~22 km. Similarly thin crust in the forearc of a subduction zone has also been observed in central Japan (Kamiya & Kobayashi 2000).

The backarc area is characterized by a normal crustal thickness of ~40 km just behind the volcanic arc, thinning towards the east to ~30 km. This thinning reflects the more positive Bouguer anomaly observed close to the Atlantic coast (Fig. 4). However, the backarc area is completely unconstrained. In addition, the local upper-crustal

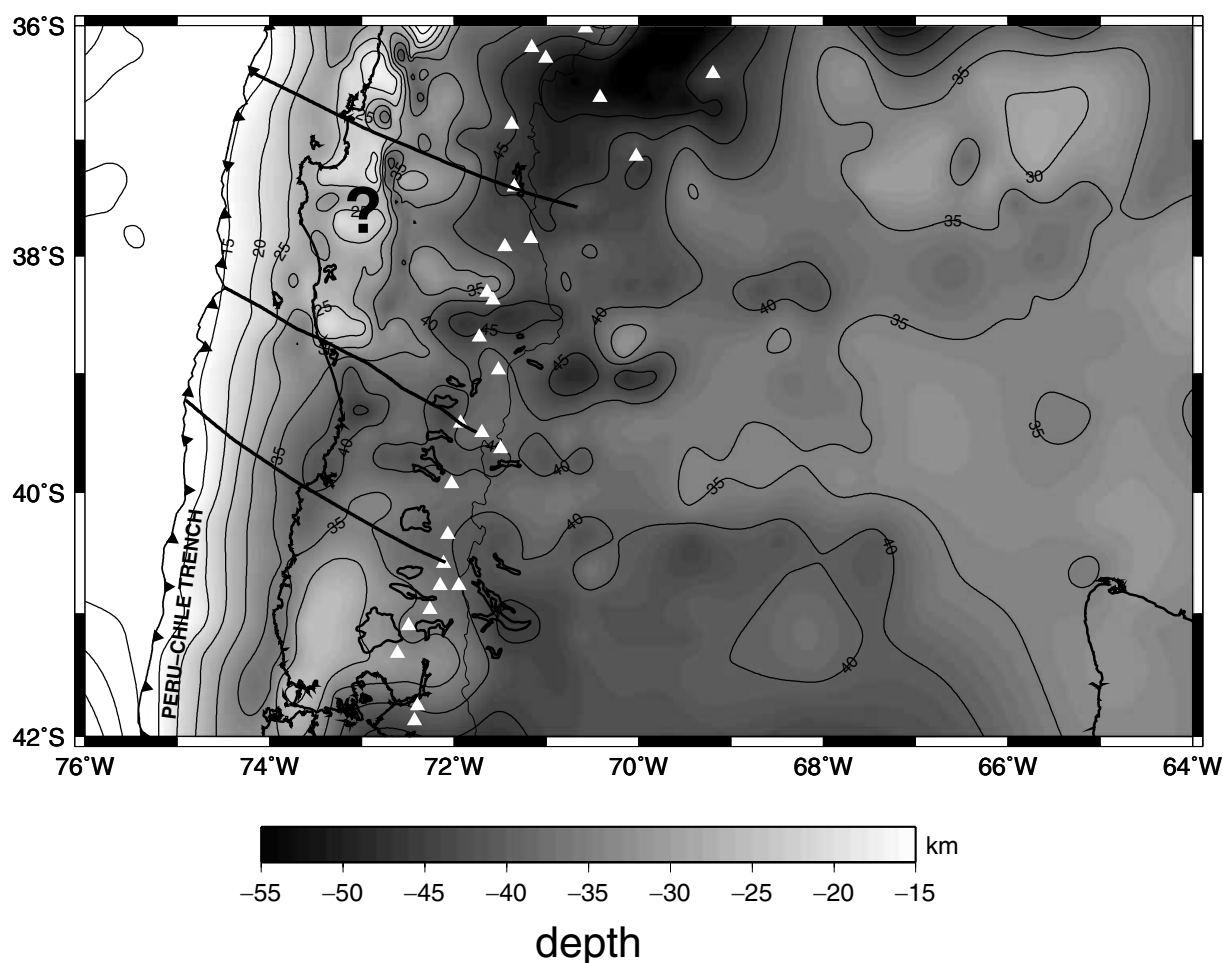


Figure 8. The depth of the continental and oceanic Moho derived from the gravity modelling. The segments identified in the study area are indicated by oblique black lines and the volcanic arc is denoted by white triangles. The crustal thickness in the forearc region of the northern segment, where the unidentified high-density body was modelled at the base of the accretionary complex, is uncertain. This is due to the uncertainty in the nature of this body.

structures, such as the Neogene backarc basins (Fig. 2) were not considered and the crust was modelled with a normal density distribution consistent with the global average (Christensen & Mooney 1995).

7.2 Serpentinized mantle

Based on the observed seismic velocities, the upper part of the mantle wedge was predicted to be typically ~15 per cent serpentinized with a bound water content of 4–6 per cent (Carlson & Miller 2003). However, Hacker *et al.* (2003) report localized hydration/serpentinization of mantle wedges that reaches 60–80 per cent. In the gravity model, the upper mantle underneath the forearc with a density of 3.23 Mg m^{-3} at depths of ~40–80 km (minimum 25 km in the southern segment) represents the hydrated/serpentinized mantle. The employed density values imply ~20 per cent serpentinized mantle, which is in accordance with the estimates given above. Serpentinized/hydrated mantle wedges have been revealed worldwide. For example, the mantle wedge in northern Chile is serpentinized/hydrated at depths of 50–100 km (ANCORP 1999; Graeber & Asch 1999), in Cascadia at depths of 30–60 km (Bostock *et al.* 2002; Blakely *et al.* 2005) and also in Costa Rica (e.g. Sallarès *et al.* 2001).

In the northern segment of the studied region, a high-density body comprising two layers with densities of 3.15 and 3.2 Mg m^{-3} , had to be included at depths of 20–40 km. This body is located at the base of the Palaeozoic accretionary complex (Figs 6 and 7). Initially, this structure was assumed to represent basally accreted sediments, but the high densities required to match the observed short-wavelength gravity signal do not support this hypothesis. Moreover, the high-density body partly overlaps with the zone of reduced P -wave velocities (~6.8–7.5 km s^{-1}) at 30–50 km depths in the local earthquake tomography model, which were interpreted to represent serpentinized mantle wedge (Bohm 2004). If the high-density body is assumed to be serpentinized mantle, the continental Moho from the gravity model would meet the subducting slab ~10–15 km shallower than interpreted in the wide angle reflection/refraction and tomography models. Moreover, the densities of $3.15 - 3.2 \text{ Mg m}^{-3}$ and P -wave velocities of $6.8 - 7.5 \text{ km s}^{-1}$, can represent both mafic lower crust and serpentinized mantle (Krawczyk *et al.* 2006 and references therein). To distinguish serpentinite from crustal material, the P - and S -wave velocity structures, as well as the V_p/V_s (or Poisson's) ratio, must be further investigated, since serpentinite tends to have a high V_p/V_s (and Poisson's) ratio (Kamiya & Kobayashi 2000). However, the estimated V_p/V_s ratio of 1.8–1.84 (Bohm 2004) is typical for both mafic crust and 20–30 per cent serpentinized mantle (Table 4).

The formation of serpentinite in the forearc mantle seems to control the downdip limit of subduction thrust earthquakes (Peacock & Hyndman 1999 and references therein), and thus the seismicity in the zone of serpentinized mantle should be reduced. In the subduction zone of central Japan, a low-velocity zone was revealed at similar depths of 20–45 km, with P -wave velocities reaching ~6.9 km s^{-1} , S -wave velocities of ~3.42 km s^{-1} and Poisson ratio of ~0.33 (Kamiya & Kobayashi 2000). The region characterized by a high Poisson ratio correlates very well with the zone in which the seismicity along the upper boundary of the slab is significantly reduced. In southern Chile, however, Bruhn (2003) shows that thrust earthquakes from the ISSA earthquake catalogue occurs to a depth of 50 km, whereas the zone of reduced P -wave velocities in the tomography model, interpreted as serpentinized mantle, is at

depths of 30–50 km. Thus, the high-density body could just represent normal lower crust rather than serpentinized mantle. Also, the unidentified body could be a combination of mafic crustal material (density of 3.15 Mg m^{-3} and P -wave velocity of ~7.1 km s^{-1}) in the upper part and hydrated/serpentinized mantle (density of 3.2 Mg m^{-3} and P -wave velocities of 7.2–7.3 km s^{-1}), characterized by 20–30 per cent serpentinization, in the lower part.

7.3 Negative forearc gravity anomaly

The most significant feature in the onshore forearc gravity field is the negative gravity anomaly in the middle Valdivia-Liquiñe segment that contrasts with the segments north and south of it (Fig. 4), and also with almost the entire margin. Moreover, a positive gravity anomaly along the continental margin within a subduction zone is a characteristic feature globally (Wells *et al.* 2003). Based on the density modelling results, there are two possible explanations for the missing positive gravity anomaly. Either the continental crust in the forearc comprises lower density material, or the position of the oceanic plate beneath the forearc region is deeper than in the northern segment. Geochemical data indicate a southward increase in the density of the forearc accretionary complex (Glodny *et al.* 2006 and references therein). Therefore, differences observed in the gravity field cannot, from a geological point of view, be explained by low-density crustal material in the continental forearc. Therefore, the variation in the gravity anomalies along the margin is best explained by varying the position of the slab below the forearc.

Examining the depth to the top of the slab along all 28 cross-sections of the density model at distances of 100 and 130 km away from the trench reveals a southward deepening of the slab (Table 6). The slab is shallowest in the northern Arauco-Lonquimay segment at 37°S–39°S (constrained by seismic data), reaches its greatest depth beneath the middle Valdivia-Liquiñe segment and shallows slightly beneath the southern Bahía-Mansa-Osorno segment (both segments unconstrained). In general, at a distance of 130 km from the trench, the slab reaches depths of 32–37 km in the northern segment, 37–42 km in the middle and ~40 km in the southern segments (Fig. 9).

When analysing the forearc gravity anomaly, it is important to keep in mind that the slab beneath the coastline in the middle segment is much deeper than in the northern segment because of the greater separation between the trench and coastline (Figs 1, 4 and 9). At latitudes between 36°S and 38°S, the trench/coastline separation is about 90–100 km, with a minimum separation of 75 km at the Arauco Peninsula (~37°S). The depth to the slab below the coast here is 22–30 km. The coastline retreats in the middle Valdivia-Liquiñe segment (at 39°–40°S), where it reaches a maximum distance from the trench of 130–140 km and the downgoing slab underneath the coastline is at its greatest depth of 36–42 km. In the southern Bahía-Mansa-Osorno segment (south of 40°S), the

Table 6. Depths to the slab below the forearc

Profile at latitude	Depth (km) 100 km from trench	Depth (km) 130 km from trench
37.3°S	–25	–33
38.8°S	–29	–36.5
39.4°S	–30	–38
39.6°S	–32	–40
40.8°S	–30	–41
41.3°S	–30	–39

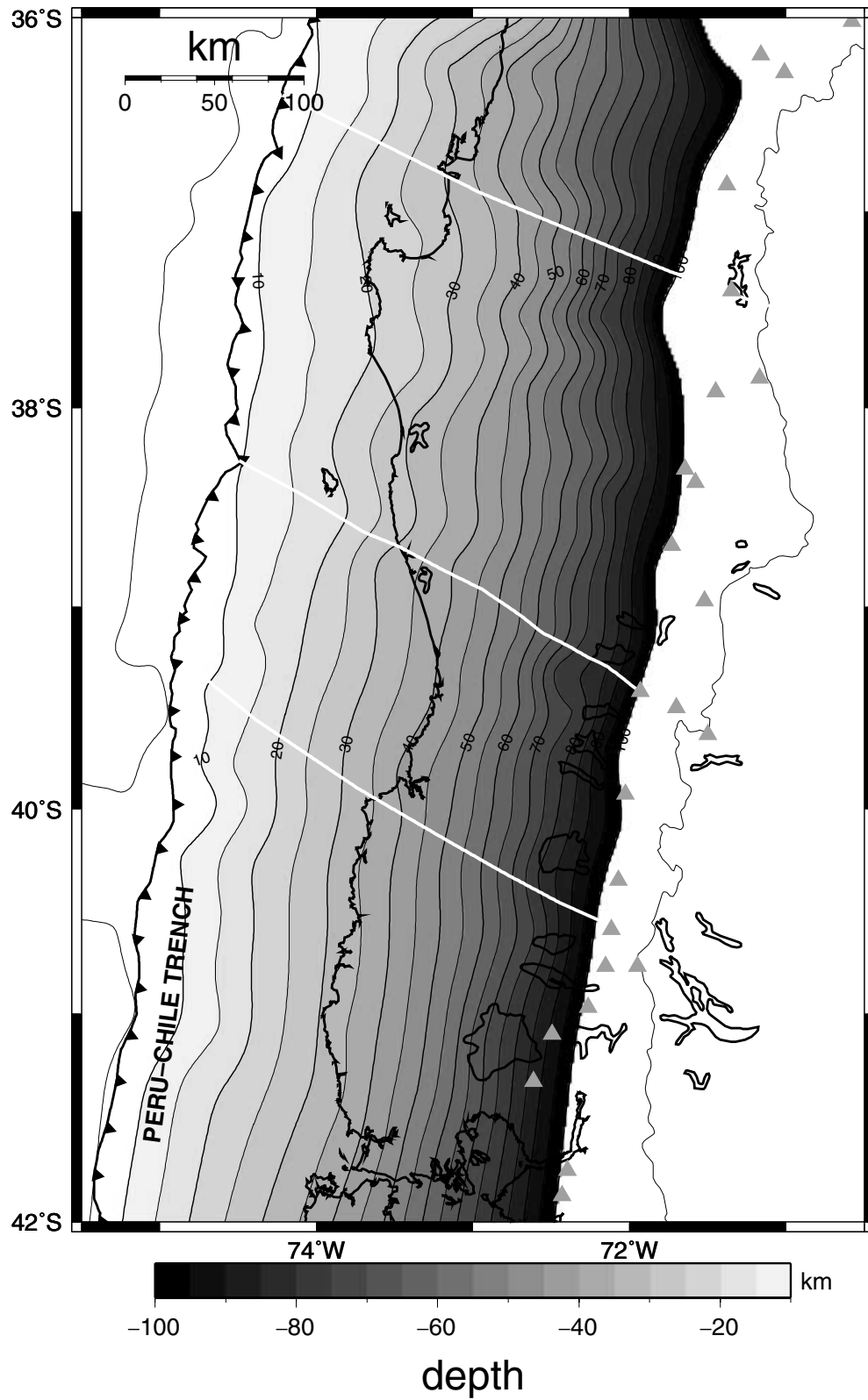


Figure 9. The depth to the slab. The resulting modelled depth to the slab is shown from the trench to the volcanic arc (grey triangles). The three segments are indicated by the oblique white lines. The slab is shallowest below the Arauco Peninsula ($\sim 37.5^\circ\text{S}$), and deepest below the forearc of the middle segment between 39° and 40°S , while the density of the accretionary complex remains constant along the margin (2.85 Mg m^{-3}).

distance between the coastline and the trench narrows to about 110 km and the slab becomes shallower (33–35 km depth).

7.4 Uncertainties in the model

The overall difference between measured gravity values and those determined from the density model is 9.9 mGal, which is similar to the error estimated for the Bouguer anomalies. The correlation coefficient between the observed and modelled fields is 0.98. Less than 5 per cent of the modelled anomaly has a deviation with respect to the observed data greater than the estimated Bouguer gravity data error. The accuracy of the 3-D density model presented is quantified in terms of how well it reproduces the observed gravity anomalies. However, this certainly does not say anything about the validity of the modelled structures, particularly in the areas where no other geophysical data are available. Due to the non-uniqueness of the interpretation of potential fields, other models with different structure and density distribution could also reproduce the observed anomalies with the same accuracy. For the southern segment of the studied region, two end-member models were constructed (Section 8). Both modelled fields have the same deviation with respect to the observed data, but the modelled structures differ significantly. The difference in the modelled Moho is ~ 10 km and the difference in the densities of both the middle and lower crust is 0.05 Mg m^{-3} . Thus, the validity of the modelled structures and a decision on whether they well represent the studied region cannot be proved by analysing the fit of the modelled and measured fields. The density model, particularly in the unconstrained regions, identifies and quantifies the possible sources of the anomalies observed (Section 8), but it should be seen as one possible representation of the lithospheric structures.

In the regions constrained by other geophysical results, the uncertainties associated with the modelled structures can be assumed based on the accuracy of the other results. The estimated errors in the velocities of the wide angle reflection/refraction seismic models are thought to be between ± 0.1 – 0.2 km s^{-1} down to about 20 km depth (Lüth and Mechie, personal communication, 2006). This is because first arrival seismic data cover the structure down to about this depth (figs 8.3–8.5 of Krawczyk *et al.* 2006). Below ~ 20 km depth the velocity errors could be up to about $\pm 0.3 \text{ km s}^{-1}$. This is due to the fact that below ~ 20 km depth, there are (almost) no first arrival seismic data covering the structure, only secondary reflected arrivals. The uncertainty of 0.1 km s^{-1} in the seismic velocities would, according to the empirical velocity–density relationships, produce a difference in the calculated densities in the order of 0.03 – 0.05 Mg m^{-3} (Table 3). However, the density used for a particular structure does not correspond to a single velocity value (Section 5). Instead, a body with a constant density (e.g. the lower crust) is consistent with a wide range of the observed velocities.

The errors in the depths to the major boundaries in the seismic models are thought to be about 10 per cent throughout. This means, for example, that the oceanic Moho at about 40 km depth has an error of ± 4 km. These error estimates are similar to those found by Patzwahl *et al.* (1999) for a very similar type of seismic experiment in north Chile. In addition, the depth error associated with hypocentre location is estimated to be ± 3 km (Bohm, personal communication, 2006), and the error in the depth to the Moho based on the receiver-function analysis is assumed to be ± 5 km (Yuan, personal communication, 2006). Thus, the uncertainties associated with the most important boundaries in the density model, such as the depth to the slab and the continental Moho, are also thought to have an error of ± 5 km in the constrained regions, which takes into account also the uncertainties of the Bouguer anomaly data. The error of

± 5 km in the depth to the slab and Moho would cause a difference in the mean density of about $\pm 0.05 \text{ Mg m}^{-3}$ (Fig. 10). Elsewhere, the depth to major boundaries depends on the densities used and, since both are unconstrained, it is impossible to estimate absolute depth errors.

8 DEFINING THE ANOMALY SOURCES

The results obtained from the density modelling are, due to the inherent non-uniqueness associated with potential field interpretation, ambiguous, particularly in the regions not constrained by other independent geophysical, geological and petrological data. Hence, the density modelling and interpretation of the gravity field cannot, in principal, resolve more than one unknown parameter. However, the density modelling can define possible sources of the observed gravity anomalies, quantify them in terms of size (depth and extent) and density, support other models and hypotheses and provide alternative predictions that can then be investigated using other methods. The following sections describe such application of the density model presented.

8.1 Upper plate density versus depth to the slab

Based on the 3-D modelling, the forearc gravity field results from the interaction of the lower (oceanic) and upper (continental) plates. The position of the Nazca plate and its density have a significant impact on the gravity anomalies. However, the depth to the slab is dependent upon the structures and densities of the upper plate. Therefore, the major differences in shape and magnitude of the gravity anomaly along the coastline are controlled and influenced by both the forearc Palaeozoic accretionary complex and the position of the oceanic plate.

In the northern segment, where the depth to the slab is constrained by seismic and seismological data, the density modelling predicts the mean density of the Palaeozoic accretionary complex, lying directly above the slab, to be 2.85 Mg m^{-3} . Additionally, an unidentified body with densities of 3.15 and 3.2 Mg m^{-3} reproduces the pronounced positive gravity anomaly observed here (Section 7.2). In the middle and southern segments, where there is no control on the slab position from other geophysical observations, the density model gives only a rough estimate of slab depth because the results are highly dependent on upper-plate structures. The estimates of slab depth would be improved by more detailed information on the composition of the accretionary complex, which means that the density of the Palaeozoic accretionary complex is a very significant issue. Surface geological observations predict a north-to-south density increase, but there is little published information on the rock mineralogical composition that could provide more accurate constraints. Despite this, the results of the 3-D density modelling show the following. If a constant density is used for the Palaeozoic accretionary complex in the entire model, then in order to reproduce the negative gravity anomaly the oceanic plate in the middle segment needs to be ~ 5 km deeper than in the northern segment. Modifying the density in the upper plate changes the position of the slab by about 5 km per 0.05 Mg m^{-3} density difference. This means that when the Palaeozoic accretionary complex is less dense, the oceanic plate is shallower. Conversely, if the Palaeozoic accretionary complex is denser, the oceanic plate is necessarily deeper. Therefore, if a north-to-south density increase was included in the density model, the resulting difference in the depth to the slab between the northern and the two segments to the south would be even greater than 5 km (Fig. 10).

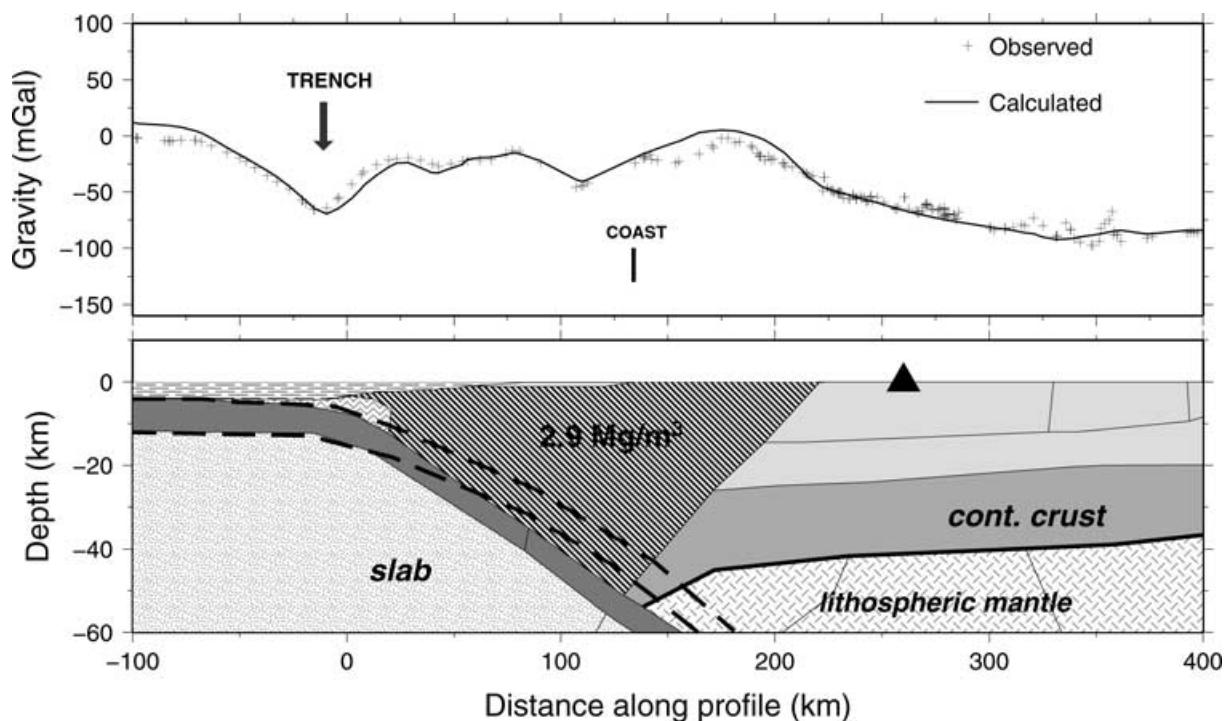


Figure 10. The ambiguous situation in the middle segment where there are no seismic constraints (along the cross-section at $\sim 39.3^\circ\text{S}$, Profile B in Fig. 5). A north-to-south density increase in the accretionary complex rocks, as indicated by the geological data, was tested and a density of 2.9 Mg m^{-3} was used. The mass excess associated with this density increase must be compensated by locating the slab even deeper than in the final model in which a constant density of 2.85 Mg m^{-3} was assumed for the accretionary complex (dashed lines indicate slab position from the final model). The thick black line denotes the continental Moho. The position of the volcanic arc is indicated by the black triangle. See legend in Fig. 6.

8.2 The crustal thinning south of 40°S

As interpreted in the geological model of Echtler (personal communication, 2002), crustal thinning beneath the Longitudinal Valley is likely in the southern segment. Since there are no other geophysical data available, this hypothesis was tested against the gravity data. The Bouguer anomaly in the southern segment is characterized by high values centred above the western part of the Longitudinal Valley (Fig. 4). The 3-D density modelling shows that a crust thinned to $\sim 25\text{--}30 \text{ km}$, with an average crustal density of 2.92 Mg m^{-3} , is one possible source of the observed gravity high. Similarly to the estimations of the depth to the slab, the predictions of the crustal thickness, although without additional constraints from seismic data, would be enhanced by more detailed information on the composition of the crust because the extent of the modelled structures is always dependent on their densities. For instance, if the density of the middle and lower crust are increased by 0.05 Mg m^{-3} (resulting in an average crustal density of 2.96 Mg m^{-3}), $\sim 40 \text{ km}$ thick crust can also reproduce the observed anomaly (Fig. 11). Since there is no detailed information about the composition of the middle and lower crust, based on the density modelling alone it cannot be concluded which possibility is more realistic. However, the model with crustal thinning is preferred, as it is supported by the tectonic regime of this area which involves extension to form the Longitudinal Valley (Jordan *et al.* 2001).

9 CONCLUSIONS

Gravity modelling was performed for the southern part of the South American convergent margin. The resulting density model integrates all available geophysical and geological data and provides

the first 3-D structural image of the Southern Andes between 36°S and 42°S . The structure of the model and the assigned densities satisfy seismic and gravity data. Also, due to the significant misfits found during the original gravity data analysis, new profile data were collected and the whole database reprocessed. This improvement in the original gravity database consequently increases the reliability of the parameters obtained from the modelling.

The most striking feature of the gravity field is the lack of the coast-parallel, positive gravity anomaly in the middle segment, which is in contrast to the rest of the margin. Geological data indicate a southward density increase, not a decrease, within the forearc accretionary complex and, therefore, the observed gravity field cannot be reproduced by the structures of the upper plate. Hence, the anomalous negative gravity in the forearc region is explained by a greater depth to the top of the slab underneath the forearc.

Based on global studies, the regions of negative gravity anomaly in the forearc correlate with zones of high seismic moment release where mega-thrust earthquakes are likely to occur (Song & Simons 2003). Additionally, high coseismic slip (asperities) tend to occur beneath the prominent gravity lows (Wells *et al.* 2003). The anomalous middle segment of the study area partly overlaps with the rupture area of the largest ever recorded Valdivia earthquake of 1960 (e.g. Herron 1981 and references therein, Cifuentes 1989). Additionally, the topography in the onshore forearc of the middle and southern segments remains quite low ($\sim 100 \text{ m}$) despite the fact that the subducting plate becomes younger and hence more buoyant towards the south. Therefore, the correlation between the negative gravity anomaly, low forearc topography, deeper slab (i.e. thicker forearc) and a high seismic moment release indicates a higher degree of plate coupling beneath the middle segment of the study

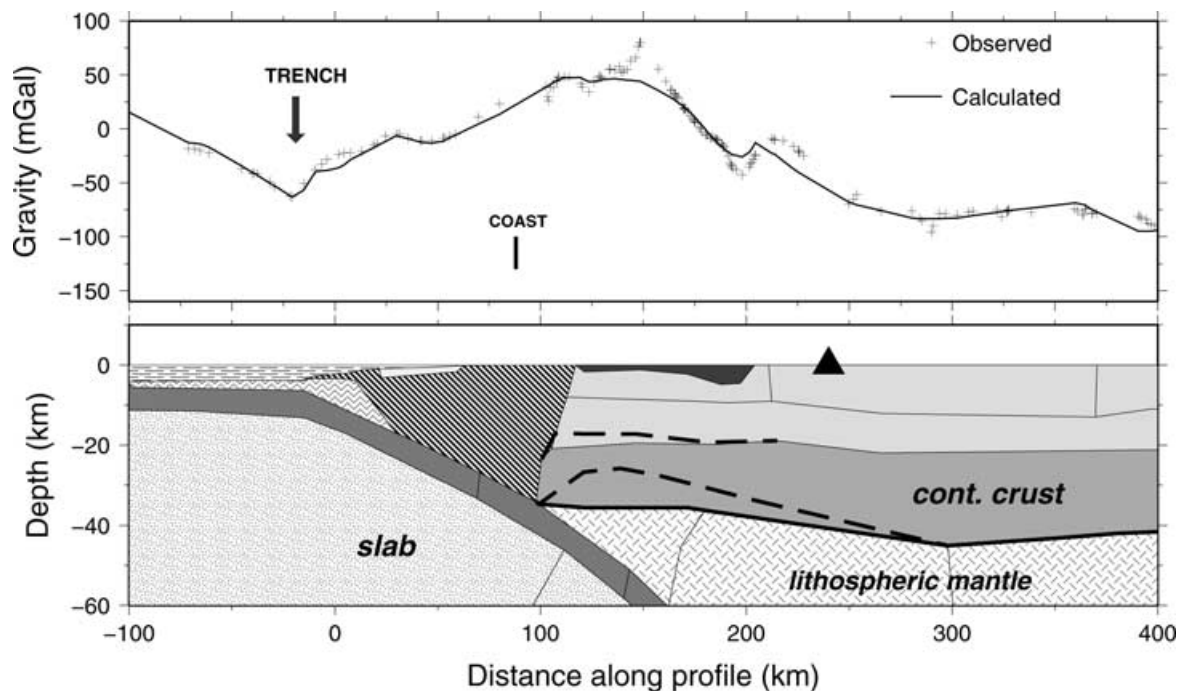


Figure 11. An alternative model that reproduces the gravity high south of 40°S, shown along the profile at 40.8°S (profile C in Fig. 5). The gravity high in the southern segment can also be reproduced by thicker crust below the Longitudinal Valley if the densities of the middle and lower crust are increased by $\sim 0.05 \text{ Mg m}^{-3}$. The Moho and the intracrustal boundary, as they are modelled in the final model, are indicated by the dashed lines. The position of the volcanic arc is indicated by the black triangle. See legend in Fig. 6.

area (Hackney *et al.* 2006). Thus the interpretation and 3-D density modelling of the subduction zone gravity field bring greater insight into the subsurface structures and provide important information for further studies of the seismogenic zone.

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