Contents lists available at ScienceDirect

### **Tectonophysics**

journal homepage: www.elsevier.com/locate/tecto

# From the North-Iberian Margin to the Alboran Basin: A lithosphere geo-transect across the Iberian Plate



TECTONOPHYSICS

A. Carballo<sup>a</sup>, M. Fernandez<sup>a</sup>, I. Jiménez-Munt<sup>a</sup>, M. Torne<sup>a</sup>, J. Vergés<sup>a</sup>, M. Melchiorre<sup>a</sup>, D. Pedreira<sup>b</sup>, J.C. Afonso<sup>c</sup>, D. Garcia-Castellanos<sup>a</sup>, J. Díaz<sup>a</sup>, A. Villaseñor<sup>a</sup>, J.A. Pulgar<sup>b</sup>, L. Quintana<sup>b</sup>

<sup>a</sup> Group of Dynamics of the Lithosphere (GDL), Institute of Earth Sciences Jaume Almera (ICTJA-CSIC), Lluís Solé i Sabarís s/n, 08028 Barcelona, Spain

<sup>b</sup> Departamento de Geologia, Universidad de Oviedo, Oviedo, Spain

<sup>c</sup> GEMOC-CCFS, Department of Earth and Planetary Sciences, Macquarie University North Ryde NSW, 2109, Australia

#### ARTICLE INFO

Article history: Received 3 December 2014 Received in revised form 23 June 2015 Accepted 16 July 2015 Available online 29 July 2015

Keywords: Iberian lithosphere Cantabrian Mountains Betics Gravity Mantle petrology Thermal modeling

#### ABSTRACT

A ~1000-km-long lithospheric transect running from the North-Iberian Margin to the Alboran Basin (W-Mediterranean) is investigated. The main goal is to image the changes in the crustal and upper mantle structure occurring in: i) the North-Iberian margin, whose deformation in Alpine times gave rise to the uplift of the Cantabrian Mountains related to Iberia-Eurasia incipient subduction; ii) the Spanish Meseta, characterized by the presence of Cenozoic basins on top of a Variscan basement with weak Alpine deformation in the Central System, and localized Neogene-Quaternary deep volcanism; and iii) the Betic-Alboran system related to Africa-Iberia collision and the roll-back of the Ligurian-Tethyan domain. The modeling approach, combines potential fields, elevation, thermal, seismic, and petrological data under a self-consistent scheme. The crustal structure is mainly constrained by seismic data whereas the upper mantle is constrained by tomographic models. The results highlight the lateral variations in the topography of the lithosphere-asthenosphere boundary (LAB), suggesting a strong lithospheric mantle strain below the Cantabrian and Betic mountain belts. The LAB depth ranges from 180 km beneath the Cantabrian Mountains to 135-110 km beneath Iberia Meseta deepening again to values of 160 km beneath the Betic Cordillera. The Central System, with a mean elevation of 1300 m, has a negligible signature on the LAB depth. We have considered four lithospheric mantle compositions: a predominantly average Phanerozoic in the continental mainland, two more fertile compositions in the Alboran Sea and in the Calatrava Volcanic Province, and a hydrated uppermost mantle in the North-Iberian Margin. These compositional differences allowed us to reproduce the main trends of the geophysical observables as well as the inferred P- and S-wave seismic velocities from tomography models and seismic experiments available in the study transect. The high mean topography of Iberia can be partly consistent with a low-velocity/high-temperature/low-density layer in the sublithospheric mantle.

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#### 1. Introduction

The present lithospheric structure of the Iberian Peninsula (hereinafter referred as Iberia) developed from the interplay of consecutive geodynamic processes related to: i) the Early Jurassic opening of the Betic-Rif corridor between the Central Atlantic and the Ligurian-Tethys oceans, ii) the Late Jurassic opening of the North Atlantic along the western margin of Iberia, iii) the Early Cretaceous opening of the Bay of Biscay and Pyrenean basins, and iv) the Late Cretaceous to Recent convergence of Africa and Europe, which triggered the total consumption of the Ligurian-Tethys Ocean. As a consequence of this protracted convergence phase and the acting horizontal stresses, both the north and south plate boundaries (Bay of Biscay–Pyrenees and Azores– Gibraltar, respectively) and the interior of the Iberian plate compressed resulting in orogenic belts and significant intraplate deformation (e.g., Dercourt et al., 1986; Banks and Warburton, 1991; Casas Sainz and Faccenna, 2001; Cloetingh et al., 2002; Vergés and Fernàndez, 2006; De Vicente et al., 2008).

From latest Cretaceous to Oligocene times shortening accommodated mainly along the Bay of Biscay–Pyrenees plate boundary, generating crustal-scale thrusts in the Pyrenees–Cantabrian Mountains but also in the inner part of the plate (Iberian Chain and Central System), as well as in the Betic–Rif and Atlas systems (Vergés and Fernàndez, 2006). Although deformation in the northern part of Iberia lasted until the late Oligocene–Early Miocene, from Oligocene times onward, the convergence between Africa and Eurasia was mostly accommodated along the southern and eastern Iberian margins, giving rise to the formation of the Betic–Rif orogenic system and the Valencia Trough–Balearic Promontory–Algerian Basin system, both related to subduction and further slab retreating of the Ligurian–Tethys domains (e.g., Torne et al., 1996; Gueguen et al., 1998; Vergés and Sàbat, 1999; Rosenbaum and

E-mail addresses: albertocarballo.g@gmail.com, acarballo@ictja.csic.es (A. Carballo).

## Lister, 2002; Faccena et al., 2004; Spakman and Wortel, 2004; Vergés and Fernàndez, 2012).

As a result of this long history, Iberia is characterized by different geological units (Vera, 2004; Fig. 1). Western Iberia corresponds to the Variscan Iberian Massif (VIM), which is mainly made up of Precambrian to Paleozoic metasediments and granitic rocks. In turn, the eastern part of Iberia is formed by thick Mesozoic sedimentary sequences inverted during the Alpine orogeny, which may incorporate Paleozoic basement, and the associated foreland basins. The eastern and southern margins of Iberia were affected by coeval extension and compression during the Neogene, giving rise to intramontane basins and the present Iberian-Mediterranean margin. Considerable volcanic activity, with magmas characterized by calc-alkaline and alkaline affinities, developed from Early Miocene to Pliocene and Quaternary times along the Mediterranean border, affecting onshore and offshore areas (e.g., Ancochea and Nixon, 1987; Cebriá and Lopez-Ruiz, 1995; Martí et al., 1992; Torres Roldán et al., 1986; Turner et al., 1999; Villaseca et al., 2010).

The superposition of all these tectonothermal episodes resulted in large lateral variations of the crust and lithospheric mantle thickness (see Torne et al., 2015–in this volume and Mancilla and Diaz, 2015–in this volume, for recent compilations and new results). In the last decades, several studies have been conducted to unravel the lithospheric structure of Iberia and its margins using a thermal approach (e.g., Ayarza et al., 2004; Fernàndez et al., 2004; Frizon de Lamotte et al., 2004; Fullea et al., 2007; Palomeras et al., 2011; Roca et al., 2004; Tejero and Ruiz, 2002; Torne et al., 2000; Torne et al., 2015–in this volume; Zeyen and Fernandez, 1994). Recent lithospheric models incorporate a quantified thermal and petrophysical characterization of the upper mantle, down to 400 km depth, in the Betic-Rif-Atlas system

(Fullea et al., 2010), across the NE Iberia and Algerian margins (Carballo et al., 2015), and across the Cantabrian Mountains and North-Iberia Margin (Pedreira et al., 2015). Up to now, however, no attempts have been made to integrate a geophysical–petrological model crossing the Iberian Peninsula from N to S and consistent with its tectonothermal evolution.

In this work, we present the crust and upper mantle structure along a 1065 km-long transect, crossing the entire Iberian plate in N-S direction, by using a modified version of the LitMod-2D code (Afonso et al., 2008). The transect starts in the Bay of Biscay and crosses the Cantabrian Mountains, the Duero Basin, the Central System, the Tagus (or Tajo) Basin, and the Betic Cordillera (hereinafter Betics), ending in the Alboran Basin (Fig. 1). The study incorporates a large amount of recently acquired seismological data from the Topolberia Project (this volume) and is aimed to: (1) investigate lithospheric structure variations in Iberia from the Cantabrian margin in the north to the Alboran Sea in the south, combining geophysical data (elevation, gravity, geoid, heat flow, and seismic), and petrological and geochemical constraints; (2) estimate potential variations in the composition of the upper mantle, compatible with seismic data and tomography models; and (3) investigate the effects of sublithospheric mantle perturbations of thermal and/ or compositional origin imaged by tomography models.

#### 2. Geological setting

For simplicity, in this section we describe the geological characteristics of the main morphotectonic units relevant to the modeled transect, which is divided in three segments. The reader is referred to Gibbons and Moreno (2002), Vera, (2004), and Vergés and Fernàndez (2006),



Fig. 1. Geological map of the study area showing the principal geological domains of Iberia (modified after Vergés and Fernàndez, 2006). Thick gray line locates the modeled transect. CZ, Cantabrian Zone; WALZ, West Asturian–Leonese Zone; CR, Ciudad Rodrigo Basin; CCR, Catalan Coastal Ranges; GB, Guadalquivir Basin.

and references therein for a more general description of the geology of the Iberian Peninsula.

#### 2.1. Northern segment – the Cantabrian Mountains

The Cantabrian Mountains represent the westward continuation of the Pyrenees along the northern coast of Iberia. They are characterized by south-directed thrusts and associated folds, involving the Paleozoic basement along most of the belt. North-verging structures are present in the easternmost part of the Cantabrian Mountains, continuing to the west along the North-Iberian margin. The estimated crustal shortening including the offshore North-verging wedge is about 100 km according to Gallastegui (2000) and Gallastegui et al. (2002) and Pedreira et al. (2015). Before this Alpine orogenic episode, several deformation stages affected the area since the Late Palaeozoic (Barnolas et al., 2004; Pérez-Estaún and Bea, 2004). The oldest event - Late Devonian to latest Carboniferous - is the Variscan orogeny, which resulted from the continental collision between Laurussia and Gondwana (e.g., Nance et al., 2012; Pérez Estaún et al., 1991). After the Variscan orogeny, a rifting episode took place during the Late Paleozoic-Early Mesozoic, giving rise to localized Permo-Triassic basins. This event was followed by an important period of crustal extension related to the opening of the North Atlantic Ocean and the Bay of Biscay during Late Jurassic and Early Cretaceous times. Crustal extension culminated in sea-floor spreading along the western margin of Iberia in the mid to Late Cretaceous, finishing in the Campanian (e.g., Fernández-Viejo and Gallastegui, 2005; Sibuet and Collette, 1991; Srivastava et al., 1990). During the Eocene and the Oligocene, N-S to NW-SE compression affected the North Iberian (Cantabrian) passive continental margin, whose geometry and inherited structures conditioned the formation of a double crustal indentation and the uplift of the Cantabrian Mountains (e.g., Alonso et al., 1996; Ferrer et al., 2008; Pulgar et al., 1995). The Cantabrian margin lower crust is underthrusted below a detachment level and protrudes into the Iberian crust forcing the northdirected subduction of its lower half (Gallastegui, 2000; Pedreira et al., 2003, 2007; Pulgar et al., 1995).

#### 2.2. Central segment – The Central System and adjacent foreland basins

The Central System is a linear ENE–WSW chain extending along 300 km in the central part of the Iberian Peninsula. It was uplifted as a crustal pop-up during Oligocene–Miocene bounded by NE–SW-trending reverse faults (Banks and Warburton, 1991; de Vicente and Muñoz-Martín, 2013; Ribeiro et al., 1990; Vegas et al., 1990). The uplifted basement corresponds to large Early Carboniferous–Permian granitoids cropping out in the western part (Bea et al., 2004; Fúster and Villaseca, 1987; Gómez-Ortiz et al., 2005), and low-grade Lower Palaeozoic metasediments in the eastern part (De Vicente et al., 2004; Gómez-Ortiz et al., 2005). The Central System is bounded by two large sedimentary basins in the central part of Iberia: the Duero Basin to the NW and the Tagus Basin to the SE.

The Duero Basin began to form during the Late Cretaceous as a continental shelf of the Cantabrian Margin. During the Eocene, the northern part evolved as a foreland basin linked to the formation of the Cantabrian Mountains (Alonso et al., 1996), whereas the southern border acted as the foreland basin of the Central System during the early Oligocene–Miocene (Capote et al., 2002). Depocenters are related to the active borders of the basin, the sedimentary thickness reaching maximum values of more than 2700 m in the north (Gallastegui, 2000) and ~3000 m in the south (Gómez-Ortiz et al., 2005). The Tagus Basin is a foreland basin related to the interaction between the Alpine tectonic uplift of the Central System and the transpressive regime along the Iberian Chain. The basin is filled with Late Eocene to Late Miocene continental deposits with maximum thickness of 3500 m along its NW border (Gómez-Ortiz et al., 2005). Close to the southern border of the Tagus Basin, the Calatrava Volcanic Province (CVP) extends over an area of 5500 km<sup>2</sup> (Ancochea and Ibarrola, 1982). In this region, Late Miocene to Quaternary magmatism with intraplate (OIB) affinity often carries xenoliths that provide direct information on the composition of the mantle. Geochemical and isotopic analyses of ultramafic xenoliths show a wide range of compositional heterogeneities due to: i) tholeiitic melts or recycled subducted oceanic crust, and ii) metasomatism of alkaline melts similar to the hosting lavas (e.g., Bianchini et al., 2010; Cebriá and Lopez-Ruiz, 1995; Villaseca et al., 2010).

#### 2.3. Southern segment – The Betic Cordillera and Alboran Basin

The southern margin of Iberia underwent a complex tectonic evolution about which there is a very active debate (e.g., Carminati et al., 2012; Casciello et al., 2015; Faccena et al., 2004; Rosenbaum and Lister, 2002; Spakman and Wortel, 2004; van Hinsbergen et al., 2014; Vergés and Fernàndez, 2012). According to Vergés and Fernàndez (2012) and Casciello et al. (2015), the Betic-Rif orogen was formed as a consequence of SE-dipping subduction of the Ligurian-Tethys lithosphere beneath Africa from Late Cretaceous to middle Oligocene, followed by a fast NW and W retreating slab roll-back. In the Algerian domain, the subduction of the Ligurian-Tethys lithosphere dipped towards the NW, and the SE- and E-directed retreating produced the opening of the Valencia Trough, the Algerian Basin, and the Tyrrhenian Sea. Many of the other models assume a unique subduction polarity of the Ligurian-Tethys lithosphere, directed to the NW and claiming for a S-, SE- and SW-directed slab retreating with a trench rotation of  $\sim 180^{\circ}$ in the Betic segment (e.g., Faccena et al., 2004; Rosenbaum and Lister, 2002; Spakman and Wortel, 2004; van Hinsbergen et al., 2014).

Independently on the preferred model, the Betic orogen comprises several major tectonic domains: a) the Guadalquivir foreland basin; b) the ENE-WSW trending External Betics, separated in Prebetic and Subbetic domains; c) the Gibraltar Flysch Units around the western side of the thrust belt; and d) the metamorphic complexes of the Internal Betics. The NW and W retreat of the Ligurian-Tethys slab produced the thrusting of the Internal Betics domain over the Iberian and Maghrebian passive continental margins. In this context, the Alboran Basin formed as a back-arc basin characterized by a thin continental crust in its western part, evolving eastwards to a thin continental crust modified by arc magmatism, a magmatic-arc crust, and finally an oceanic crust in its easternmost part (Booth-Rea et al., 2007; Torne et al., 2000). The extensional evolution of the Alboran basin was accompanied by broadly distributed volcanism ranging from early Oligocene (backarc) tholeiitic through calc-alkaline series to middle-late Miocene tholeiitic through calc-alkaline and alkaline series related to subduction and lithosphere thinning, respectively (e.g., Duggen et al., 2004).

#### 3. Methodology

The methodology used in this work is based on the *LitMod-2D* code (Afonso et al., 2008). This software combines petrological and geophysical modeling of the lithosphere and sub-lithospheric upper mantle, within an internally-consistent thermodynamic–geophysical framework. The model follows a forward scheme in which the crustal geometry is constrained from previous seismic studies, and the lithosphere mantle geometry and composition are changed according to the geodynamic setting, xenolith data and tomography models until the best fit is reached.

Each mantle body in *LitMod-2D* is characterized by its main oxides composition (in weight%) within the Na<sub>2</sub>O–CaO–FeO–MgO–Al<sub>2</sub>O<sub>3</sub>–SiO<sub>2</sub> (NCFMAS) system. Stable mineral assemblages are determined through self-consistent thermodynamic calculations using the code Perplex, which minimizes free energy for a given combination of oxides as a function of pressure and temperature (Connolly, 2005). Thus, physical properties of each mineral and of the bulk mantle (density, thermal

expansion coefficient, elastic parameters, and thermal conductivity) depend not only on temperature, but also on pressure, composition, and phase changes. In this work, we use an augmented-modified version of Holland and Powell (1998) (revised in 2002) thermodynamic database, which takes into account all major phases relevant for the continental crust (Afonso and Zlotnik, 2011). Anelastic effects are computed based on the experimental results and model of Jackson et al. (2002). In this study we selected a grain size of 5 mm for the mantle, in agreement with previous studies (Afonso et al., 2008 and references therein).

The heat transport equation is solved by using the finite elements method in steady-state with the following boundary conditions: 0 °C at the surface, 1330 °C at the LAB, and no heat flow across the lateral boundaries of the model. Below the LAB, the algorithm considers a 40 km-thick thermal buffer with a temperature of 1400 °C at its base. This is done to avoid unrealistic discontinuities between a conductive thermal gradient within the lithospheric mantle and an adiabatic thermal gradient within the asthenosphere. The temperature at the base of the model (400 km) is set to 1520 °C. The temperature gradient below the thermal buffer layer is restricted to 0.35 < dT/dz < 0.50 °C/km, otherwise the temperature at 400 km depth is modified accordingly (see Afonso et al., 2008 for further details). Then, gravity, geoid, elevation, surface heat flow and P- and S-seismic velocities are computed and compared with observations.

Some modifications have been incorporated to the LitMod-2D code to improve the mantle thermal conductivity calculations and to incorporate sub-lithospheric mantle anomalies. First, we add the radiative contribution to lattice thermal conductivity, as described in Grose and Afonso (2013). Second, we consider thermal, compositional, or thermo-compositional anomalies relative to the surrounding sublithospheric mantle to explain seismic velocity anomalies imaged by tomography models. In the case of thermal anomalies, the code assigns to the anomalous zone(s) the same composition as given to the asthenosphere (usually Primitive Upper Mantle, PUM), and recalculates the relevant physical parameters (density, seismic velocity, phase changes, and thermal conductivity) at P and T+ $\Delta$ T conditions,  $\Delta$ T being the prescribed temperature anomaly relative to the surrounding mantle. When the anomaly is compositional, the code calculates the relevant physical parameters at the T-P conditions, considering the prescribed chemical composition. Thermo-compositional anomalies can be related to lithospheric mantle bodies that have been detached and sunk into the asthenosphere and therefore, have a different temperature and composition than the surrounding asthenosphere. These sub-lithospheric anomalies can be coupled, when the vertical stresses induced by their buoyancy due to density anomalies are transmitted to surface elevation, or decoupled, when density anomalies are not transmitted to surface elevation. Therefore, decoupled anomalies do not affect calculated isostatic topography but they do on gravity and geoid calculations.

The temptative lithospheric structure is used to calculate the lateral lithostatic pressure variation at the compensation depth (400 km). There are deviations from the local isostasy model that we compensate with a flexural isostatic model to evaluate whether they can result in significant topography misfits. Vertical loads relevant to flexure are computed from the lateral changes in lithostatic pressure at the base of the model (400 km depth) resulting from the derived crust and upper mantle structure. Note that if a constant pressure at the base of the transect over the entire profile is obtained, it implies that the section is under perfect local isostasy. To calculate the deflection from the vertical load distribution, we use an updated version of the code tAo (Garcia-Castellanos, 2007). The deflection or vertical displacement of each column, associated with the vertical load, is calculated for a prescribed elastic thickness Te distribution. Larger Te values smooth out longer wavelengths of these vertical displacements, reducing the misfit between the model and the observed topography. Using a Te that is consistent with the tectonothermal age of the lithosphere (Watts, 2001) allows reducing the misfit below uncertainty and then the regional isostasy can explain the topographic profile and the misfit of the local isostatic model. See Jiménez-Munt et al. (2010) for details about flexural calculation.

#### 4. Geophysical data

#### 4.1. Surface geophysical data

Surface geophysical observables as elevation, Bouguer anomaly, geoid height, and heat flow (Fig. 2) were used to constrain the resulting crustal and upper mantle structure along the selected geo-transect. Elevation data come from ETOPO1 (Amante and Eakins, 2009), a global elevation model of the Earth surface of  $1 \times 1$ -min arc resolution. Along the profile, elevation is over 2000 m in the Cantabrian Mountains, about 1500-2000 m in the Betic Cordillera and ~1300 m in the Central System. The Duero and Tagus basins are flat areas with average elevations of ~800 m and ~600 m, respectively, while in the Bay of Biscay (Cantabrian Sea) and the Alboran Sea are -4500 m and -1900 m, respectively. Bouguer anomalies onshore were obtained from a recent compilation of gravity data in Iberia (Ayala, 2013) in the framework of the Topo-Iberia Project. Offshore free-air gravity anomalies were obtained from the global satellite altimetry data model V16.1 (Sandwell and Smith, 1997, updated 2007) and Bouguer anomalies were computed using the FA2BOUG code (Fullea et al., 2008). Maximum Bouguer anomaly values are achieved in the Bay of Biscay and the Alboran Sea, reaching 300 and 100 mGal, respectively, whereas minimum values of less than -100 mGal are found in the Central System and the Betic Cordillera. Geoid height is taken from EGM2008 (Pavlis et al., 2008) and filtered up to degree and order 8, to retain only a residual geoid anomaly that reflects the density distribution within the lithosphere and the uppermost sub-lithospheric mantle (~400 km depth). Most of Iberia shows geoid height values between 8 and 11 m, with maximum values along the profile in the Betic Cordillera and the Tagus Basin. The minimum geoid values are in the oceanic basins, reaching -5 m in the Bay of Biscay. Surface heat flow data were obtained from Fernàndez et al. (1998) and Marzán Blas (2000). Along the profile, the heat flow onshore Iberia varies between 45 and 65 mW/m<sup>2</sup>, exceeding 100 mW/m<sup>2</sup> in the Alboran Sea. To account for lateral variations perpendicular to the strike of the profile, elevation, gravity and geoid data have been projected onto the profile within a strip of 25 km half-width, whereas for heat flow data we used a 50 km half-width strip.

#### 4.2. Crustal geometry from previous studies

The crustal geometry along the geo-transect is based on previous works including seismic surveys, geological cross-sections and combined geological-geophysical interpretations. The northern segment, from 0 to 500 km distance, coincides with the model presented by Pedreira et al. (2015), which is based on the same methodology. A difference with our model is that Pedreira and coauthors calculate the physical parameters of the lower crust beneath the Duero Basin and the Cantabrian Mountains using a thermodynamic approach. These authors consider a dry felsic–granulite composition (Villaseca et al., 1999) beneath the southern half of the Duero Basin and a more intermediate composition (50%–50% mixture between felsic–granulite and the global average composition of Rudnick and Gao (2003)) with 2% wt. H<sub>2</sub>O northwards. Pedreira et al. 2015) also distinguish two alternative models in which, depending on the assumed pre-compression structure, the lower crustal root beneath the Cantabrian Mountains reaches maximum depths of ~60 km or ~90 km.

In contrast, southwards of the Cantabrian Mountains, we consider a lower crust with homogeneous density (and thermal properties) along the whole profile. Beneath the Cantabrian Mountains and for depths deeper than 30 km, we have adopted the same intermediate-hydrated lower crust composition considered by Pedreira et al. (2015), using also the thermodynamic approach to calculate the physical parameters.



**Fig. 2.** Geophysical data. a) Elevation map from ETOPO1 Global Data Base (V9.1) (http://www.ngdc.noaa.gov/mgg/fliers/01mgg04.html). b) Bouguer anomaly. Onshore, in the Iberian Peninsula, the Bouguer anomaly comes from Ayala (2013). Offshore and Africa Bouguer anomaly has been calculated from the free air anomaly by Sandwell and Smith (1997) using Fullea et al. (2008). Contour interval is 20 mGal. c) Geoid anomaly map from EGM2008 Global Model (Pavlis et al., 2008). Long wavelengths components (>5000 km) have been removed by subtracting spherical harmonics up to degree and order 8 from the total geoid. Contour interval is 1 m. d) Surface heat flow measurements from Fernàndez et al. (1998), Marzán Blas (2000) and International Heat Flow Commission global data set for Algeria (http://www.heatflow.und.edu/index2.html). Colors of solid circles indicate heat flow values.

Therefore, the hydrated lower crust is progressively densified with depth and ultimately eclogitized beneath the Cantabrian Mountains (body 12 in Table 2). For simplicity, we only consider the shallower crustal root option (~60 km depth), since this is the preferred model by Pedreira et al. (2015). Similar Moho-depth values are locally found in several Alpine orogens like Eastern Alps (60 km; Cassinis, 2006), Zagros (56–69 km; Paul et al., 2010), Pyrenees (55–70 km; Chevrot et al., 2015), and Rif (>55 km, Mancilla and Diaz, 2015–in this volume). See also Artemieva and Thybo (2013) for a European Moho compilation. Another difference with Pedreira's model is the thickness of the lower crust beneath the southern half of the Duero Basin, where we consider a value of 9 km, according to Suriñach and Vegas (1988), instead of 7 km, which was an extrapolation based on the observed thickness along a seismic refraction profile located further north (Profile 5, Fernández-Viejo et al., 2000).

To constrain the crustal structure in the central segment, we used the Toledo-Salamanca wide-angle seismic survey (Suriñach and Vegas, 1988), the ALCUDIA deep seismic reflection profile (Martínez Poyatos et al., 2012), the ALCUDIA2 Wide-Angle Seismic Reflection Transect (Ehsan et al., 2015-in this volume) and several geological cross-sections (e.g., Banks and Warburton, 1991; Casas Sainz and Faccenna, 2001; Casas-Sainz and De Vicente, 2009; Quintana et al., 2015-in this volume). In the southern segment, we used seismic data from different experiments (e.g., Banda et al., 1993; Carbonell et al., 1997; Comas et al., 1995; Gallart et al., 1995) and geological cross-sections (e.g., Banks and Warburton, 1991; Berástegui et al., 1998; Frizon de Lamotte et al., 2004; Michard et al., 2002; Platt et al., 2003 and Ruiz-Constan et al., 2012). In addition to these data, we have also used for the whole transect the compilation of seismic data in Iberia by Díaz and Gallart (2009), and the recent receiver function data obtained in the Topolberia project from the deployment of seismic stations (IberArray) covering the whole Iberian Peninsula (e.g., Mancilla et al., 2012, 2015-in this volume).

#### 4.3. Physical properties

The physical properties assigned to the sediments and the crustal layers are summarized in Table 1. Most of the interior of the Iberian Peninsula, from the External Betics to the Cantabrian Mountains, shows a horizontal layered crustal structure (Fig. 3). However, in the western Central System, and according to different authors, we have considered an upper crustal structure with different lithology dominated by monzogranitic and leucogranitic rocks (e.g., Gómez-Ortiz et al., 2005; Jiménez-Díaz et al., 2012; Martín-Velázquez and de Vicente, 2012). In consequence, we have considered a lower density and higher radiogenic heat production in the Central System than in the surrounding upper crust (Table 1). Along the geo-transect, radiogenic heat production has been taken from a global compilation carried out by Vilà

#### Table 1

Parameters used in the model for the different crustal bodies. a and b values are dependent
on the chemical composition and pressure-temperature conditions (see text and Table 2)

Bodies	Crust and sediments model	Density (kg/m <sup>3</sup> )	H. Ρ. (μW/m <sup>3</sup> )	T. C (W/(K · m))	
1	Tertiary and Quaternary	2350	1.2	2.5	
2	Mesozoic sediments	2550	1.2	2.5	
3	Accretionary prism	2550	1.5	2.5	
4	Upper crust	2750	1.65	2.4	
5	Middle crust	2850	0.5	2.1	
6	Lower crust (Iberia)	2950	0.2	2	
7	Lower crust (Cantabrian Margin)	2900	0.33	2	
8	Upper Crust Central System	2670	2	2.4	
9	Lower Crust (Bay of Biscay)	3120	0.15	2	
10	Neogene Extended Continental	2900	0.5	2.5	
	Crust of Alboran				
11	Betics domain	2750+6*Z	1	2.5	
		(km)			
12	Subducted N-Iberian lower crust	a	b	0.02	

et al. (2010). Thermal conductivity values for crustal bodies were taken from previous studies (Fernàndez et al., 1998). Density values have been derived from previous studies and from available seismic velocities, and fall within the velocity-density envelopes defined by Barton (1986), Christensen and Mooney (1995) and Hamilton (1978). In the northern segment of the geo-transect we have taken the physical properties from Pedreira et al. (2015), with only very minor modifications in some bodies to allow a better integration with the rest of the data along the transect. In the mantle and in the subducting lower crust beneath the Cantabrian Mountains, mineral assemblages and physical properties are calculated by the thermodynamic module in *LitMod-2D* from the prescribed chemical composition and pressure-temperature conditions (Table 2). The adopted crustal structure is shown in Fig. 3.

#### 4.4. Mantle characterization and LAB depth from previous studies

The characterization of the mantle along the profile comes mainly from available Pn- and Sn-velocities (i.e. for the uppermost mantle), global and regional tomography models, and geochemical and isotopic analyses on xenolith samples. Pn data have been compiled from different experiments (e.g., Díaz et al., 2013; Díaz and Gallart, 2009; and DSSWG-Alboran, 1978), showing average values of 8.0-8.1 km/s in Iberia mainland with maximum values of ~8.2 km/s in the southern Variscan Iberian Massif (e.g., Díaz and Gallart, 2009; Díaz et al., 2013). Minimum Pn values of  $\sim$ 7.75  $\pm$  0.15 km/s are found in the Alboran Basin (Díaz and Gallart, 2009 and DSSWG-Alboran, 1978). Sn values have been taken from regional tomography in the Euro-Mediterranean region (Díaz et al., 2013) resulting in average velocities of ~4.6 km/s, with maximum values of 4.65 km/s near the Central System. In the northern part of the profile, beneath the North-Iberian Margin, the uppermost mantle show anomalous low Vp velocities, with values of 7.7–7.9 km/s detected in seismic refraction/wide-angle reflection profiles (Fernández-Viejo et al., 1998; Pedreira et al., 2015; Ruiz, 2007).

Deeper P-wave mantle velocities along the modeled profile are obtained from a global travel-time tomography model using the same method described in Bijwaard et al. (1998), incorporating additional earthquakes from 1995 to 2002 and arrival times (Villaseñor et al., 2003). In total, more than 14 million arrival times from 300,000 earthquakes were reprocessed using the EHB methodology (Engdahl et al., 1998). The ray paths corresponding to these new arrival times sample, mainly, the uppermost mantle with a resolution of  $0.5^{\circ} \times 0.5^{\circ}$  in horizontal and 25-50 km in depth. Fig. 4 shows the depth distribution of the Pwave velocity anomaly relative to the ak-135 model (Kennett et al., 1995). Positive anomalies, exceeding 1%, correspond to the thinned crust and comparatively colder lithosphere of the North-Iberian Margin, and to the subducting Ligurian-Tethys slab beneath the Betics-Alboran region. Negative anomalies, exceeding -2%, are located beneath the Alboran Basin and the Calatrava Volcanic Province (CVP), extending down to 200 km depth. The tomography section also shows a zone of slow P-wave velocity (-0.3% to -1%) extending from the northern end of the profile to the CVP, and from 230 km depth to the bottom of the model. A local conspicuous negative anomaly is imaged beneath the Cantabrian Mountains, which could be related to crustal thickening due to shortening. From the Cantabrian Mountains to the CVP, the lithospheric mantle does not show significant velocity anomalies.

S-wave mantle velocities are taken from a European shear wave velocity model derived from inversion of seismic shear and surface waveforms (Legendre et al., 2012). These authors show velocity anomalies relative to a regional reference model for different layers with bottom depths of 80, 110, 150 and 200 km, respectively. S-wave velocities show relatively low values ( $\sim -3\%$ ) down to 110 km depth. No significant anomalies are recorded at deeper levels, with the exception of the CVP where a conspicuous negative anomaly, exceeding -3%, from 110 to 200 km depth is identified, the maximum intensity being at 110–150 km.





Fig. 3. Crustal structure corresponding to the best fitting model. Different bodies have been modeled using geological cross-sections (overlaid) and seismic experiments (see text for details). Physical properties of crustal bodies are reported in Tables.

Xenolith data along the modeled profile are scarce and restricted to the Central System and CVP. The analyses reported by Orejana et al. (2006) on three pyroxenite xenolith samples from the Central System give compositions of 13-18% MgO, 11-19% CaO and 9-15% Al<sub>2</sub>O<sub>3</sub>, with Mg# of 76. Unfortunately, these compositions, together with P-T estimates, indicate an origin related to metasomatism of depleted mantle or cumulates at the base of the crust, and are far to be representative of the bulk mantle composition (Orejana et al., 2006). In the Calatrava Volcanic Province, geochemical evidences in terms of major and trace elements of mineral phases, as well as whole rock and isotopic analyses on spinel lherzolite xenoliths, show two metasomatizing events affecting the mantle beneath this region: one is related to a subduction component agent reflecting clinopyroxenes LREE, Pb, Th and U enrichments and more radiogenic whole rock compositions, while the second is related to an alkaline melt, similar to the host lavas (MREE enrichments in the clinopyroxenes and HIMU signature of whole rock Sr-Nd systematics) (e.g., Bianchini et al., 2010; Cebriá and Lopez-Ruiz, 1995; Villaseca et al., 2010).

The geometry of the LAB beneath the Iberian Plate has been the subject of several studies integrating different geophysical data. In the northern segment of the profile the results obtained by Pedreira et al. (2015), who used a similar crustal structure, show a maximum LAB depth of ~170 km beneath the Cantabrian Mountains, shallowing to 125–145 km below the Duero Basin. Along the central segment, LAB-depth estimates from 1D thermal modeling show values of 95– 100 km beneath the Duero and Tagus basins (Jiménez-Díaz et al., 2012), which are slightly lower than those obtained by Fernàndez et al. (1998, 2004), who proposed values of 110–120 km. In the southern segment, Torne et al. (2000) based on 3D gravity modeling combined with heat flow and elevation data, proposed lithospheric thickness values ranging from 140 km beneath the Betics to ~50 km in the East Alboran Basin. More recently, Fullea et al. (2010) proposed LAB-depth values of ~160 km beneath the Betics and 50–70 km beneath the East Alboran Basin based on a 3D geophysical–petrological model.

None of the previously referred models accounted for the presence of a conspicuous positive seismic velocity anomaly beneath the Betics imaged by seismic tomography and extending down to ~600 km depth (e.g., Bezada et al., 2013; Bijwaard and Spakman, 2000; Garcia-Castellanos and Villaseñor, 2011; Monna et al., 2013; Piromallo and Morelli, 2003; Spakman and Wortel, 2004). Such anomaly has been interpreted as a retreating lithospheric slab, which is probably affected by a lateral tear propagating westwards (Garcia-Castellanos and Villaseñor, 2011; Vergés and Fernàndez, 2012).

#### 5. Results

Along the modeled profile we have distinguished three lithospheric mantle types differing in composition according to the geological domains (Fig. 5). The overall composition of the lithospheric mantle along the profile falls in the lherzolitic field. Thereby, we have considered a lherzolite average (Table 2) (Tc\_1 in Griffin et al., 2009) as the lithospheric mantle composition prevailing in the Iberian Peninsula and the North-Iberian Margin. In the region of the Calatrava Volcanic Province, we changed slightly the composition towards a more fertile mantle, according to available geochemical data on xenolith samples (Table 2) (Villaseca et al., 2010). In the East Alboran Basin we have considered a PUM composition according to Fullea et al. (2010) (Table 2). The North-Iberian Margin is characterized by low Pn velocities and then, we have followed the composition model proposed by Pedreira et al. (2015), where a thin layer (5-15 km thickness) of hydrated uppermost mantle is considered (body 13, Fig. 5e). The lithospheric slab below the Betics is modeled assuming a colder body with a temperature anomaly of  $\Delta T = -200$  °C and with the same composition than the

Table 2

Chemical compositions of mantle bodies (% wt) used in the model. Body 12 corresponds to the subducted lower crust beneath the Cantabrian Margin.

Bodies	Composition	SiO <sub>2</sub>	$Al_2O_3$	FeO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> 0	H <sub>2</sub> O
12	Subducted N-Iberian lower crust	56.89	16.81	7.88	5.28	5.47	2.56	1.97	2
13	Serpentinized Tc_1	44.31	3.48	7.96	39.63	3.09	0.24	0.03	1
14	Tc_1 (Av. Tecton)	44.64	3.51	8.02	39.93	3.11	0.24		
15	CVP (Calatrava_65290)	44.51	3.76	8.75	37.89	3.28	0.36		
16	Pm_1 (PUM)	45.00	4.5	8.10	37.8	3.6	0.36		



Anomaly  $V_p \%$ 

Fig. 4. P-wave tomography model along the profile (Villaseñor et al., 2003). P-wave tomography anomaly is relative to the ak135 reference model. Contour lines every 1%.

Iberian lithosphere (Tc\_1). The slab is un-welded to the overlaying lithospheric mantle to simulate the lateral tearing.

The composition of the sub-lithospheric mantle is assumed to correspond to PUM, according to McDonough and Sun (1995). To account for the low velocity layer imaged by tomography from 260 km depth to the bottom of the model, we have increased the temperature of this layer by 65 °C along the transect, and by 80 °C beneath the Calatrava Volcanic Province (Fig. 5e). Note that the geometry of the sublithospheric bodies is slightly narrower than the seismic anomalies derived from tomographic models in order to reproduce the average percentage ( $\%\Delta Vp$ ) value.

The best fit model is illustrated in Fig. 5. The calculated response of gravity, geoid and elevation match very well the observed regional trends with some local misfits. Gravity and geoid show a root mean square error (rmse) of 9.2 mGal and 0.48 m, respectively. We have considered that the sub-lithospheric bodies are mechanically coupled to the lithosphere, such that the vertical stresses related to density changes within the asthenosphere are transmitted to the Earth's surface. Therefore, the density anomalies related to sub-lithospheric bodies affect not only the calculated geoid and gravity anomalies, but also the elevation. In the next section we discuss how an eventual mechanical decoupling of these anomalous sub-lithospheric bodies could affect the calculated elevation. The calculated elevation, under the assumption of local isostasy, shows local misfits of up to 1000 m in the North-Iberian Margin and of few hundred meters in the Betics and the Alboran Basin (rmse - 360 m). However, when flexural rigidity of the lithosphere is considered, the calculated elevation fits well with observations (rmse = 90 m). We have considered a variable elastic thickness, changing from Te = 30 km in the northern segment of the profile (e.g., Pedreira et al., 2015), to Te = 20 km in the central segment (e.g., Jiménez-Díaz et al., 2012; Ruiz, 2007; Gómez-Ortiz et al., 2005), and Te = 10 km in the southern segment (e.g., Garcia-Castellanos et al., 2002). Because the relevant lateral pressure variations at the compensation level are of relative short wavelength (<100 km), the associated vertical motions are significantly smoothed by flexural isostatic compensation. The calculated surface heat flow (SHF) is also in agreement with the general trend of measured values. Highest calculated SHF values correspond to the thinned lithosphere of the Alboran Basin, although a noticeable underestimation relative to the measured values is observed. In the Iberian mainland, the calculated SHF decreases from ~60 mW/m<sup>2</sup> in the Tagus Basin to 55 mW/m<sup>2</sup> in the Duero Basin, and to ~50 mW/m<sup>2</sup> in the Cantabrian Mountains and the North-Iberian Margin.

The resulting depth of the lithosphere–asthenosphere boundary shows conspicuous lateral variations, ranging from minimum values of around 60 km beneath the Alboran Basin, to more than 160 km beneath the Betics. The central part of the Iberia shows a LAB-depth of 135 km decreasing to ~130 km below the Duero and Tagus basins and to ~110 km in the CVP. The Cantabrian Mountains are characterized by a LAB-depth of ~180 km shallowing towards the North-Iberian Margin to values of 130–140 km.

#### 5.1. Temperature and density distribution

The calculated steady-state temperature distribution is shown in Fig. 6. Within the crust, the isotherms are roughly flat, showing a slight upward deflection beneath the Southern Betics, related to the pronounced lithospheric thinning in the Alboran Basin. In contrast, the isotherms show a downward deflection in the North-Iberian Margin, related to the thin crust and low heat production. The Moho temperature varies from ~200 °C in the Bay of Biscay, to about 650 °C in the Iberian Meseta and the Calatrava Volcanic Province, increasing to ~750 °C beneath the Betics and the North-Iberian Margin, where the Moho depth reaches values of 40 km and 60 km, respectively. In the central part of the Alboran Basin the Moho temperature decreases to values of 500 °C. The temperature distribution in the sub-lithospheric mantle shows the perturbations relative to the surrounding adiabatic thermal gradient prescribed in the subducting slab underneath the Betics (-200 °C), and the sub-lithospheric layer located below 260 km depth that rises to less than 140 km depth beneath the CVP  $(+65 \text{ and } +80 \degree \text{C}, \text{ respectively}).$ 

The density distribution of subcrustal domains was calculated according to mantle composition and thermodynamic formulation, and it is shown in Fig. 7. In the northern segment, the density at the crustmantle boundary (CMB) varies between  $\sim$  3320 kg/m<sup>3</sup> beneath the Cantabrian Mountains to  $\sim$  3230 kg/m<sup>3</sup> in the hydrated mantle of the North-Iberian Margin. In central Iberia the uppermost mantle density is of 3300 kg/m<sup>3</sup>, decreasing to 3290 kg/m<sup>3</sup> beneath the Betics and to 3250 kg/m<sup>3</sup> beneath the Alboran Basin. The lithospheric mantle of the northern and central segments of the profile is almost isopycnic, with average densities of ~3350 kg/m<sup>3</sup> and ~3325 kg/m<sup>3</sup>, respectively. Beneath the Betics, the lithospheric mantle density increases with depth from 3290 kg/m<sup>3</sup> below the Moho to 3380 kg/m<sup>3</sup> at the LAB, indicating that in this region the pressure effects on density prevail on the temperature effects due to the high Moho temperature and the consequent low thermal gradient within the lithospheric mantle. Below the Alboran Basin, the lithospheric mantle is also almost isopycnic with a density of ~3250 kg/m<sup>3</sup>. Maximum lithospheric mantle densities of ~3360-3380 kg/m<sup>3</sup> are reached at the base of the lithosphere, beneath the thickened regions of the Betics and the Cantabrian Mountains, respectively.



Fig. 5. Model results (red lines), measured data with the standard deviation of data projected onto the profile within a strip of 25 km half-width (blue dots and vertical bars): (a) surface heat flow; (b) Bouguer anomaly; (c) geoid height; (d) elevation: in red, calculated topography assuming local isostasy; in green, calculated elevation in the case of effective elastic thickness for the different Te (30 km, 20 and 10 km) used along the profile (see text for details); (e) lithospheric structure. Crustal structure like in Fig. 3. Numbers in panel (e) denote mantle bodies summarized in Table 2. Dataset sources as in Fig. 2.

The imposed temperature anomalies in the sub-lithospheric mantle bodies translate into density anomalies of  $-6 \text{ km/m}^3$  (perturbation of +60 °C) and  $-9 \text{ kg/m}^3$  (perturbation of +80 °C), approximately. Within the slab, thermal and composition anomalies result in a

positive density contrast of ~10 kg/m<sup>3</sup> at the top of the slab, vanishing with depth up to the olivine–wadsleyite phase transition, with a density increase of 70 kg/m<sup>3</sup>. Note that lowering the average temperature of the slab by 200 °C produces the rising of the Ol-Wd



Fig. 6. Calculated temperature distribution along the modeled transect and mantle domains with different chemical composition (see Table 2). Contour lines every 100 °C.

transition from around 410 km to 390 km depth. See also Discussion and Fig. 13.

#### 5.2. Velocity results

The depth-distribution of the calculated P-wave velocities within the mantle is shown in Fig. 8a. Most of the central Iberia shows values of 8.05–8.08 km/s, increasing to > 8.1 km/s in the Cantabrian Mountains and the North-Iberian Margin, with the exception in the last case of the uppermost, hydrated part, which show calculated velocities of ~7.9 km/s. The CVP shows lithosphere mantle velocities of ~8 km/s, whereas beneath the Betics, velocities increase with depth from 8.0 to 8.15 km/s. The Alboran Basin shows the lowest calculated Vp values, ranging from 7.9 km/s, close to the Moho, to 7.6 km/s at the LAB. At sublithospheric levels, velocities are affected by the presence of bodies with prescribed anomalous temperature and composition. Therefore, the positive thermal anomalies of 65–80 °C result in a P-wave velocity decrease of ~0.05–0.08 km/s. Within the slab, the combined effect of negative thermal anomaly (-200 °C) and composition implies an increase of ~0.15 km/s (see also Fig. 12). Fig. 8b shows the calculated Vp anomalies in (%) relative to a reference model consisting of a 130-km thick lithosphere and 30-km thick crust with a lithosphere mantle composition corresponding to Tc\_1. Both, the global tomography (Fig. 4) and the synthetic tomography (Fig. 8b) models show very similar anomalies despite some minor misfits in their amplitude and shape.

Calculated S-wave velocities (Fig. 9) show a similar pattern than P-waves, with maximum values within the lithospheric mantle



Fig. 7. Calculated mantle density distribution along the modeled transect. Contour lines every 20 kg  $m^{-3}$ .



Fig. 8. a) Calculated P-wave seismic velocities in the upper mantle down to 410 km; b) synthetic P-wave tomography from our model. In this case velocity anomalies are relative to the synthetic reference model consisting of a 130-km thick lithosphere and 30-km thick crust with a lithosphere mantle composition corresponding to Tc\_1. Contour lines every 1%. Compare with Fig. 4.

of > 4.6 km/s beneath the North-Iberian Margin and minimum values of 4.2–4.4 km/s beneath the Alboran Basin. A low velocity zone with Vs < 4.45 km/s, extending from some km above the LAB to 220 km depth, is clearly depicted along the profile except beneath the Cantabrian Mountains and the Betic Cordillera. At sub-lithospheric levels, the positive thermal anomaly of 65–80 °C translates into a Vs decrease of 0.1–0.12 km/s, whereas the combined compositional and negative thermal anomaly of -200 °C in the subducting slab implies a Vs increase of 0.08 km/s. Superposed on the calculated S-wave velocities in Fig. 9, we plotted the Vs values obtained at different depth levels from the regional tomography model of Legendre et al. (2012). In this case, we can only compare qualitatively the lateral velocity variations down to 200 km

depth. Although our results reproduce maximum lithosphere mantle velocities beneath the North-Iberian Margin and the Betic Cordillera, and minimum velocities in the CVP, there is a major discrepancy in the predicted velocity-depth variation. The main qualitative difference is that, according to Legendre et al. (2012), the velocity tends to increase with depth following the ak-135 reference model and therefore, these authors do not image the sub-lithospheric low velocity zone except in the CVP. However, recent studies carried out in Iberia and the westernmost Mediterranean region, based on ambient noise and ballistic finite-frequency Rayleigh wave tomography, show an almost continuous low velocity zone except beneath the Cantabrian (not sampled) and Betic orogens (Palomeras et al., 2014).



**Fig. 9.** Calculated S-wave seismic velocities in the upper mantle down to 410 km depth. Contour lines every 0.05 km/s. Superimposed are the S-wave velocities inferred from the regional tomography model by Legendre et al. (2012) at different depth levels. S-values have been calculated according to the reference velocities of 4.38 km/s (at 80 and 110 km), 4.39 km/s (at 150 km) and 4.45 km/s (at 200 km) (see Legendre et al., 2012).

Fig. 10 compares the calculated P- and S-wave velocities in the uppermost lithospheric mantle with Pn and Sn estimates from different seismic experiments taking into account the reported uncertainties in the experimental determinations. The calculated Pn and Sn values are the averaged velocities within the first 15 km of the uppermost mantle and the associated standard deviation. Our Pn velocity results fit with



Fig. 10. a) Pn velocities along the modeled transect. Red line indicates calculated Pn values. Circles with error bars denote Pn-values from seismic experiments (see legend); b) Sn velocities along the modeled transect. Red line indicates calculated Sn values. Symbols with error bars denote Sn-values from (Díaz et al., 2013).

most of the available data except in the Betics and the Alboran Basin, where Díaz and Gallart (2009) have proposed values of 8.2  $\pm$  0.05 km/s 7.75  $\pm$  0.15 km/s, respectively. The calculated Sn velocity values show a similar trend than available data (Díaz et al., 2013) although with slightly lower magnitudes.

#### 6. Discussion

The integrated petrological-geophysical modeling approach used in this study has some intrinsic limitations that should be considered in the subsequent sections. The assumption of thermal steady-state with no advective transport is strictly valid for old tectono-thermal regions (>100 my), as it is the case of the Variscan part of the geo-transect. This assumption is less valid in the Cantabrian Mountains and the Betic-Alboran system, where tectonic deformation ended more recently. However, the presented results are constrained by the simultaneous fitting of a set of "instantaneous" density-dependent observables such as gravity, geoid and elevation, and then results must be considered as a snapshot of the present-day density distribution related to active tectonic processes. A transient thermal model could improve the calculated temperature and density distribution, but the vertically averaged density anomaly must remain similar to yield similar topography. In contrast, accounting for transient effects requires a more sophisticated numerical approach and, more importantly, a deep knowledge of the past and ongoing geodynamic processes and particularly the timing of those processes.

The assumption of local isostasy implies that the lithosphere does not retain shear vertical stresses and it is a particular case of regional isostasy with Te = 0. This assumption is valid for young and weak lithospheres, as can be the case of the Betic-Alboran region ( $Te \approx 10$  km), but not along the rest of the profile, where the lithosphere is rheologically stronger with a  $Te \ge 20$  km. The resulting density distribution after the best fit with the observed elevation will depend on the considered Te, such that the larger Te, the larger allowed misfit in elevation.

Sub-lithospheric mantle anomalies inferred from tomography models are related to either temperature or chemical variations beneath the LAB. The resulting sub-lithospheric density anomalies will tend to generate creeping flow in the asthenosphere (Stokes flow) and viscous stresses. How these stresses are transmitted through the lithosphere and modify the calculated isostatic elevation (dynamic topography contribution) depends on the vertical viscosity distribution within the crust and upper mantle. The end-members of this dynamic topography contribution are calculated by considering a sub-lithospheric mantle that is either fully isostatically coupled to the crust or fully decoupled from it.

Finally, the physical properties of the mantle bodies are determined by their chemical composition and the P–T conditions. Density depends largely on composition whereas seismic velocities are very sensitive to temperature conditions. However, identifying mantle density and seismic velocities with bulk composition is not straightforward because of the lack of uniqueness. Afonso et al. (2013a,b), based on a non-linear 3D multi-observable probabilistic (Bayesian) inversion approach, show that a wide range of compositions can, equally well, explain multiple geophysical data. Hence, deep temperature anomalies  $\leq 150$  °C and compositional anomalies  $\Delta Mg\# < 3$  are not simultaneously resolvable, being the bulk Al<sub>2</sub>O<sub>3</sub> content a better compositional indicator than Mg#. In consequence, the considered mantle chemical compositions are compatible with the geophysical observables, but we cannot decide whether these compositions are unique.

#### 6.1. LAB topography

The calculated topography of the lithosphere–asthenosphere boundary relies on the considered crustal structure (geometry and parameters), on mantle images from seismic tomography studies, and on the compositions of the lithospheric and sub-lithospheric mantle. The modeled crustal structure is based on available data from seismic experiments and integrated geophysical models carried out close to the studied transect. Similarly, several works have investigated the lithospheric structure of Iberia using different methodological approaches. Here, we focus on the comparison of our proposed LAB-depth along the studied transect with those proposed from previous studies (Fig. 11).

In the northern segment of the transect, we used a very similar model to that of Pedreira et al. (2015). Consequently, the LAB depths obtained by these authors are very similar to the depths attained in our model: 10-15 km deeper under the North-Iberian Margin and southern Duero Basin regions whereas is about 20 km shallower beneath the crustal root of the Cantabrian Mountains (Fig. 11). These differences in the LAB-depth can be mainly attributed to small differences in the calculated mantle thermal conductivity and in the geometry and density of the lower crust. Mantle thermal conductivity in Pedreira et al. (2015) is based on Hofmeister (1999), while in the present work we are considering the radiative contribution described in Grose and Afonso (2013). In addition, Pedreira et al. (2015) consider a 7-9 km thick lower crust beneath the Duero Basin and used a density depending on composition and P-T conditions. In our case, we used a homogeneous lower crust along the whole profile (except for the crustal root beneath the Cantabrian Mountains) with a constant density of 2950 kg/m<sup>3</sup> and a thickness of  $\sim$ 9 km.

Interestingly, our results show a gentle increase in the LAB depth from ~130 km beneath the Duero and Tagus basins to ~135 km beneath the Central System. Similar lithospheric thickening values have been proposed from thermo-mechanic numerical experiments as resulting from the application of horizontal stresses and shortening to the northern and southern margins of the Iberian Peninsula (Martín-Velázquez and De Vicente, 2012).

A similar integrated geophysical-petrological approach was also used to image the 3D lithospheric structure in the Atlantic-Mediterranean transition region by Fullea et al. (2010). In their model, the LAB depth varies from <70 km in the Alboran Basin to 150–170 km beneath the Betics and ~110 km in the Iberian Massif and thus, it is consistently deeper than in our model (Fig. 11). These differences are related to: i) the crustal structure, which in the case of Fullea et al. (2010) was simpler; ii) the above mentioned calculation of the mantle thermal conductivity; and iii) the presence of the detached/torn lithospheric slab beneath the Betics, which is not considered in Fullea et al. (2010).

Many of the studies carried out in Iberia concerning the lithosphere structure and LAB-depth are based on a 'pure thermal approach'. Following this approach, Fernàndez et al. (1998) and Jiménez-Díaz et al. (2012) proposed lithospheric thickness values of about  $100 \pm 10$  km beneath the Duero and Tagus basins, and the Central System. Both studies were based on 1D modeling of surface heat flow and elevation data. Integrated 2D lithospheric modeling combining heat flow, gravity, geoid, and elevation data gave a slightly thicker lithosphere below the Tagus Basin, with a LAB depth of 120 km (Fernàndez et al., 2004). A 3D modeling of the lithosphere of the Alboran Basin and surroundings, combining elevation, surface heat flow and gravity data, yielded LABdepth values of ~60 km in the central part of the basin, and >130 km beneath the Betics (Torne et al., 2000). Finally, a recent 1D study combining elevation, geoid data, and thermal analysis, further constrained by 3D gravity modeling over the whole Iberian Peninsula, shows very similar LAB-depth values, except in the northern and southern margins (Torne et al., 2015-in this volume). These margins are characterized by strong crust and lithospheric mantle deformation affecting structures far beyond the shoreline and therefore not considered in the referred study. An inherent limitation of the thermal approach is that the density of the lithospheric mantle is assumed to be only temperature dependent, and that the underlying asthenosphere has a constant density everywhere. Consequently, the density within the lithospheric mantle decreases with depth according to the thermal expansion coefficient and therefore, the LAB cannot be effectively constrained by seismic or



Fig. 11. Calculated topographies of the lithosphere-asthenosphere boundary from different models.

petrological models. Nevertheless, although the absolute LAB-depth values can differ from a geophysical-petrological approach, the main trends will be comparable as far as the predominant effect on lateral density variations is related to temperature rather than to pressure and/or composition (see Tunini et al., 2014 for a thorough discussion). We must finally note that our methodology is also subjected to uncertainties in the determination of the depth to the geotherm defining the LAB of the order of  $\pm$  10–15% (Afonso et al., 2013a,b).

#### 6.2. Lithospheric mantle composition

The lithospheric mantle, along the modeled transect, shows lateral changes in composition. We have considered a homogeneous composition in the Iberian mainland, with the exception of the Calatrava Volcanic Province, according to the availability of geochemical analyses on xenoliths and/or volcanic samples. The dominant composition corresponds to the Average Tecton garnet sub-continental lithospheric mantle (Tc\_1) (Griffin et al., 2009), which has also been proposed by Pedreira et al. (2015) and Fullea et al. (2010) along the northern and southern segments of the transect, respectively. Carballo et al. (2015) used a slightly different composition for the lithospheric mantle beneath the Pyrenees and NE-Iberia, based on the average composition of the lherzolites from the Pyrenean massif of Lherz (Pr6, Griffin et al., 2009). However, as demonstrated by Carballo et al. (2015) and Pedreira et al. (2015), using compositions Tc\_1 or Pr\_6 give almost undistinguishable results. Carballo et al. (2015) also tested the four different compositions proposed for 'tecton' lithospheres (formed or modified at <1 Ga) from global studies (Griffin et al., 2009). These authors concluded that both, Tc\_1 and Tc\_2, could be compatible with the geophysical observables with small changes in the structure of the lithospheric mantle. The density difference at given P-T conditions attains maximum values of 6 kg/m<sup>3</sup> between these two lithospheric mantle compositions. In contrast, Tecton spinel peridotitic compositions Tc\_3 and Tc\_4 from Griffin et al. (2009) yield too low lithospheric mantle densities  $(-10 \text{ kg/m}^3 \text{ and } -12 \text{ kg/m}^3 \text{ relative to Tc_1, respectively})$ that integrated over the lithospheric mantle results in noticeable changes either in the LAB-depth or in the geometry and/or density of the crust.

The Calatrava Volcanic Province shows recent volcanism defining a second compositional domain in the lithospheric mantle along the profile. Spinel lherzolite xenoliths from the Cenozoic CVP provide a direct sampling of the shallow lithospheric mantle of beneath this area (Villaseca et al., 2010). Calatrava xenoliths are lherzolites estimated to originate from depths of 35–50 km. They are richer in clinopyroxene and poorer in orthopyroxene than other Iberian mantle xenolith suites, being slightly to moderately depleted relative to primordial mantle. The absence of harzburgites in the studied suites of the CVP reflects the very low degrees of partial melting (F) that have affected this portion of mantle (<10%) (Villaseca et al., 2010); low F-values are inferred also by the trace element contents of clinopyroxenes and by the low cr# [calculated as Cr/(Cr+Al) of the spinels, allowing to speculate that the mantle beneath CVP has a composition very close to that of a fertile mantle. A composition of a mantle, depleted by a very low partial melting event (1.5%), estimated by the HREE contents of the clinopyroxenes was chosen to characterize the CVP mantle (Villaseca et al., 2010). The change in the lithospheric mantle composition in the CVP results in a shallower LAB since the density contrast with respect to  $Tc_1$  is about ~11 kg/m<sup>3</sup>.

In the lithospheric mantle beneath the Alboran Basin we have considered a PUM composition (McDonough and Sun, 1995), as proposed by Fullea et al. (2010). This composition was also considered by Carballo et al. (2015) for the Algerian Basin. Compared to Tc\_1, a PUM composition allows for a higher density and a lower P-wave velocity at the uppermost mantle levels. This lower velocity is related to the higher content in plagioclase of PUM relative to Tc\_1, which is particularly noticeable at the Pg stability field. Finally, in the North-Iberian Margin we have also modified the mantle composition by considering a hydrated uppermost mantle, with 1% water content, following Pedreira et al. (2015). Mantle hydration results in alteration of peridotites and the consequent reduction of P-wave velocities as measured in seismic experiments.

#### 6.3. Thermal and compositional sub-lithospheric anomalies

Seismic velocity anomalies at sub-lithospheric levels, imaged from tomography models, suggest the presence of bodies with different temperature and/or composition than in the surrounding PUM sublithospheric mantle. The positive velocity anomaly, exceeding  $1\% \Delta Vp$ 

beneath the Betics, we interpreted as a detached/torn lithospheric slab with a temperature 200 °C lower than the surrounding sublithospheric mantle, and a Tc\_1 composition corresponding to the Iberian mantle lithosphere. The negative velocity anomaly, exceeding -2% $\Delta$ Vp beneath the CVP, and extending along the central and northern segments of the profile down to 260 km depth, is interpreted as a positive thermal anomaly with a  $\Delta$ T = 65–80 °C and the same composition as the sub-lithospheric mantle.

Fig. 12 shows the depth variation of Vp along selected vertical profiles to illustrate the influence of temperature, phase changes and chemical composition. The resulting graphics clearly show the Vp increase at ~40 km depth related to the plagioclase-spinel transition amounting <0.05 km/s, except in the Cantabrian Mountains where Pg-Sp transition superposes to the change from hydrated to dry mantle. In this case, the increase in Vp amounts 0.32 km/s. The Vp increase related to the compositional change at the LAB is also very conspicuous, especially in regions with thin and 'hot' lithospheric mantle (Fig. 12b and c). The Vp decrease related to anomalous high temperature in the sublithospheric mantle (Fig. 12a, b, c) amounts ~0.06 km/s for  $\Delta T =$  65 °C and ~0.09 km/s for  $\Delta T = 80$  °C (Fig. 12c). The combined thermal and compositional effect of the lithospheric slab implies a Vp increase of ~0.125 km/s between 200 km and 280 km depth, being of ~0.1 km/s at higher depths due to the high-pressure orthopyroxene phase change. Finally, the P–T conditions within the lithospheric slab cause the rise of the exothermic olivine–wadsleyite mineral phase transition to 385 km depth and then, a Vp increase of ~0.31 km/s (Fig. 12d).

Changes in chemical composition, phase transitions, and P–T conditions, translate into density changes as shown in Fig. 13. The density increase related to compositional change through the LAB, from Tc\_1 to PUM, amounts ~20 kg/m<sup>3</sup>; whereas that related to the olivine–



Velocity profiles (Vp)

**Fig. 12**. Variation of P-wave velocities with depth, calculated along the profile (at 200, 400, 750 and 900 km) for the preferred model with and without the thermal asthenospheric anomaly. Arrows denote the lithosphere–asthenosphere boundary (LAB) and the following phase transitions: Pg–Sp, plagioclase–spinel; Opx–Hpx, orthopyroxene–high-pressure magnesian pyroxene.



Fig. 13. Variation of mantle density with depth, calculated along the profile (at 200, 400, 750 and 900 km) for the preferred model with and without the asthenospheric anomaly.

wadsleyite transition within the subducting slab, usually at 410 km depth, amounts ~70 kg/m<sup>3</sup>. The positive sub-lithospheric thermal anomaly implies a decrease in density of 6 kg/m<sup>3</sup> in the northern and central segment of the profile, and of 10 kg/m<sup>3</sup> beneath the CVP. The combined effect of temperature and composition in the lithospheric slab produces a density increase of 8 kg/m<sup>3</sup> at its top, vanishing with depth until the olivine–wadsleyite phase transition.

We have calculated the contribution of the sub-lithospheric anomalous bodies on elevation when vertical stresses related to asthenosphere density changes are not transmitted to the Earth's surface (decoupled mode). In this case, topography equals to that calculated without anomalous bodies, though their influence on potential fields would remain. Fig. 14 illustrates the difference between elevation calculated considering coupled and decoupled anomalous sub-lithospheric bodies. The topography perturbation caused by these bodies could be partly attenuated by viscous deformation within the low-viscosity asthenosphere, and partly by regional isostasy due to lithosphere rigidity. The lower the wavelength of the anomaly is the higher the attenuation of the topography perturbation. Therefore, an accurate estimation of the role of sub-lithospheric mass-anomalies on topography requires dynamic calculations and accurate constitutive equations. According to our results, the maximum topography perturbation related to the higher temperature of the asthenospheric body is around +340 m along the northern and central segments of the profile, rising to +500 m in the Calatrava Volcanic Province (Fig. 14). The topography contribution related to the slab is of about -300 m, due to the combined effect of its lower temperature, which vanishes with depth, and to the olivine–wadsleyite phase transition occurring near the bottom of the slab. Interestingly, trying to fit the decoupled elevation makes the fitting the measured gravity and geoid difficult. This deep and long wavelength density anomaly, imaged by tomography models as a relatively low velocity layer, could be the responsible for the high mean elevation of Iberia, which exceeds 600 m above sea level.

The proposed contribution of the sublithospheric mantle thermal anomalies on topography is a small part of the total contribution of the lithospheric structure. According to Lachenbruch and Morgan (1990), the elevation of an unloaded asthenospheric column is of -3500 m, meaning that the overlying lithosphere contributes to the final elevation by about 4.1 km for an average lberian topography of 600 m. Therefore, the sub-lithospheric mantle anomalies contribute between 7.1% (in the Central System) and 11.6% (in the CVP) to the observed elevation if vertical stresses are totally transmitted to the surface (dynamic coupling).

Despite the sublithospheric mantle anomalies, it is clear that the major contribution to the present-day Iberian relief is due to the horizontal stresses arising from the protracted Late Cretaceous to Recent convergence between Iberia and Africa. Horizontal stresses acted mainly on the plate boundaries but also propagated to the plate's interior causing deformation of the crust and the lithospheric mantle. Therefore, the superposition of different length-scale deformations in the upper



Fig. 14. Difference between the elevations calculated considering coupled and decoupled anomalous sub-lithospheric bodies. The topography perturbation varies from +340 m in the northern and central segments of the profile, rising to +500 m and decreasing to about -300 m in the Betics.

crust, mid-lower crust, and lithospheric mantle resulted in lateral variations of the lithosphere buoyancy and the corresponding isostatic response (e.g., Cloetingh et al., 2002; Fernández-Lozano et al., 2011; Martín-Velázquez and De Vicente, 2012; Torne et al., 2015–in this volume; Vergés and Fernàndez, 2006).

#### 7. Conclusions

Joint modeling of gravity, topography, geoid, heat flow, and mantle seismic velocity data, using geological and crustal seismic data as constraints, allowed us to propose a new 2D lithospheric model across the Iberian Peninsula and its north and south margins. Different geophysical observables are simultaneously fitted, thus reducing the uncertainties associated with the modeling of these observables alone or in pairs. The results obtained from the numerical experiments allow us to draw the following concluding remarks:

- The resulting lithospheric structure shows large variations in crust and lithospheric mantle thicknesses. The LAB depth beneath the North-Iberian Margin is 130–140 km increasing to 180 km under the Cantabrian Mountains. In the central part of Iberia the LAB gently deepens from 130 km beneath the Duero and Tagus basins to 135 km beneath the Central System, reaching only ~100 km depth further south, around the Calatrava Volcanic Province. Southwards, the LAB depth increases to about 160 km below the Betics and decreases sharply to around 60 km in the Alboran Basin.
- Along the modeled geotransect the lithospheric mantle must be heterogeneous in composition. We have distinguished four chemical compositions, which are compatible with geophysical observations in terms of density, seismic velocities and geodynamic setting. In the Cantabrian margin (Bay of Biscay) we consider a hydrated (1 wt.% H<sub>2</sub>O) Phanerozoic to Neo-Proterozoic or 'tecton' (Tc\_1) uppermost mantle whereas in the continental mainland we use a predominant 'tecton' average (Tc\_1) mantle composition. A more fertile lithospheric mantle close to PUM is considered to underlie the Calatrava Volcanic Province, while beneath the Alboran Basin, we use a PUM composition like in the underlying asthenosphere.
- Beneath the Betic Cordillera, we consider a detached/torn lithospheric slab characterized by a 'tecton' composition (Tc\_1), and a temperature

anomaly of -200 °C relative to the surrounding sub-lithospheric mantle, which is interpreted as corresponding to the Ligurian-Tethys slab. Along the northern and central segments of the profile, we have considered a sub-lithospheric thermal anomaly of +65 °C extending from 260 km of depth to the bottom of the model. In the Calatrava Volcanic Province, this anomaly might reach +80 °C in the uppermost sub-lithospheric mantle at depths between 240 and 120 km.

- Topography perturbations related to the low velocity-high temperature asthenospheric anomaly may amount up to + 340 m along the northern and central segments of the Iberian plate along the study profile, and to + 500 m in the Calatrava Volcanic Province. The detached/torn lithospheric slab beneath the Betic Cordillera generates a topography perturbation of 300 m, due to its higher average density as a consequence of its lower temperature and the rising of the olivine–wadsleyite transition. Topography perturbations of sublithospheric origin require a dynamic coupling between the lithosphere and the sublithospheric anomalous bodies.
- Despite the clear tectonic origin of the Iberian relief, the high mean topography of the Iberian Peninsula can be partly related to the buoyancy induced by the low-velocity /high-temperature/low-density layer from 260- to 400-km depth consistent with tomography models. Mechanical coupling of these density anomalies with the overlying lithosphere seem to be necessary to simultaneously reproduce the topography and potential field data over the area.

#### Acknowledgments

We are indebted to two anonymous reviewers and to the Associated Editor A. Azor for their helpful comments and suggestions on the previous versions of the manuscript. The presented work has been supported by Topo-Iberia Consolider-Ingenio CSD2006-0004, GASAM/TopoMed (CGL2008-03474-E/BTE/07-TOPO-EUROPE-FP-006), TECLA (CGL2011-26670) funded by the Spanish Government, and PYRTEC-IP2 (SV-PA-10-03, funded by the Government of Asturias/ESF TOPO-EUROPE Programme) projects. AC benefitted from a JAE-PreCP grant from CSIC. JCA acknowledges the support from ARC Grant DP120102372. This is

contribution 617 from the ARC Centre of Excellence for Core to Crust Fluid Systems (http://www.ccfs.mq.edu.au) and 1004 in the GEMOC Key Centre (http://www.gemoc.mq.edu.au).

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