The role of lithospheric heterogeneities in the location of the Cenozoic intraplate deformation of Iberia from finite element modeling

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[1] The Cenozoic evolution of the Central System basement uplift and of the continental Duero and Madrid basins on the Iberian Peninsula occurred in a compressive context triggered by the Alpine orogeny. Although several models of deformation have been suggested, the factors that led to the development of these have not been analyzed to date. We present here finite element modeling of the intraplate deformation of the Iberian lithosphere during the Cenozoic. This modeling has taken into account the role played by (1) the tectonic shortening originating from the Pyrenean and Betic collisions, (2) the weight of Tertiary continental sediments, and (3) thermomechanical heterogeneities present in the lithosphere. The numerical model represents a lithospheric section, with a creep-type mechanical behavior, perpendicular to the Central System, and the two sedimentary basins. Both initial conditions and the results obtained have been specified based on the available geophysical and geological data. This study has enabled us to simulate the present lithospheric structure of the central Iberian peninsula: (a) the formation of the Central System relief and its horizontal shortening, (b) the basement folding in the Duero and Madrid basins, (c) the development of the root of the mountain range, and (d) the depth of the lithosphere-asthenosphere border in the interior of Iberia. The deformations occurred through a large-scale folding and thickening mechanism, by the shortening of a lithosphere that was initially thinner, and with lateral contrasts in temperatures and composition.

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1. Introduction

[2] In the past few years a lot of research has been undertaken into the geodynamic processes that control continental topography [Burg and Podladchikov, 2000; Cloetingh *et al.*, 2005; Vietor and Oncken, 2005; Célérrier *et al.*, 2005]. Geological and geophysical observations show that this activity is not only located at the plate boundaries but also that plate interiors can be subject to intense deformation. In order to build models that can afford a better understanding of the intraplate dynamics, it has become normal to use numerical tools that integrate and evaluate a great deal of multidisciplinary information on the behavior of the lithosphere. In this way it has been shown that lithospheric shortening in convergent geodynamic settings is accommodated through the homogeneous thickening of the lithosphere and the development of folds and faults [Cloetingh *et al.*, 1999, 2002; Burg and Podladchikov, 2000; Gerbault, 2000;

Guest *et al.*, 2007; Burg and Schmalholz, 2008]. Analytical [Biot, 1961; Lambeck, 1983; Zuber, 1987; Schmalholz *et al.*, 2002] and analog studies [Shemenda, 1992; Martinod and Davy, 1994; Burg *et al.*, 1994; Sokoutis *et al.*, 2005] have also been investigating the lithosphere deformation by means of large-scale folding in oceanic and continental areas.

[3] The topography of the Central System mountains and the basements of the Duero and Madrid continental flexural basins were structured during the Cenozoic in the center of Iberia (Figure 1). Although they have been intensely studied, there is still an interesting ongoing debate as to the origin of these deformations and their relationship to the plate boundaries [Ribeiro *et al.*, 1990; Vegas *et al.*, 1990; de Vicente *et al.*, 1996; Gómez-Ortiz *et al.*, 2005; de Vicente *et al.*, 2007; de Vicente and Vegas, 2009]. The Central System uplift is attributed to different tectonic mechanisms: brittle deformation in the upper crust with ductile thickening in the lower crust [Vegas *et al.*, 1990; de Vicente *et al.*, 1996], thickening of the middle crust related to intracrustal detachments [Ribeiro *et al.*, 1990; Gómez-Ortiz *et al.*, 2005], and lithospheric folding resulting from horizontal shortening [Cloetingh *et al.*, 2002; de Vicente *et al.*, 2007; de Vicente and Vegas, 2009]. These processes have been attributed to the tectonic stresses transmitted toward the interior from the active borders of the Iberian microcontinent. These are known as the Pyrenean and/or Betic compressions that have

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Figure 1. (a) Location of the main Cenozoic tectonic units of the Iberian Peninsula. The black dotted rectangle shows the position of Figure 1b, and the black dotted and dash line shows the location of the gravity profile of Figure 1c. The division of the Iberian Massif is depicted with yellow lines: C, Cantabrian Zone; AOS, West Asturian-Leonese Zone; OS, Olla de Sapo Domain; CI, Central Iberian Zone. (b) Physiography of the Spanish Central System, the Duero basin and the Madrid basin (SRTM-90 m). The NW-SE black line indicates the orientation of the modeled pre-Cenozoic lithospheric profile. A, Almanzor peak (2592 m); P, Peñalara peak (2430 m). (c) The 2+1/2D gravity model for central Iberia modified from *de Vicente et al.* [2007]: top is the calculated gravity (accumulated error 1.50 mGal) and bottom is the density model.

resulted from the interactions of the Eurasian, Iberian and African plates from the Late Cretaceous until today. Nevertheless, some authors also give a significant role to the tectonothermal events that have affected the Iberian lithosphere from the Paleozoic [*Vergés and Fernández, 2006*].

[4] This paper examines the question of these intraplate Cenozoic deformations through modeling different sections of the Iberian lithosphere with a plane strain approach (ANSYS software). In the model we integrate the available data on the thermomechanical structure of the center of the Iberian Peninsula. Our main goals are two: (a) to analyze the factors that gave rise to these deformations far from the plate boundaries, and (b) to simulate a structure for the crust and the lithosphere equivalent to today's. With regard to the first, we will look at the effects of the tectonic shortening that resulted from the Eurasia-Africa convergence, the lithostatic

load of the tertiary sediments and the thermal and compositional lateral heterogeneities of the Iberian lithosphere. As for our second goal, a general picture of the topography and horizontal shortening of the Central System is obtained. This includes the geometry of the Variscan basement flexure of the Duero and Madrid basins, the crustal root below the intraplate range and the current thicknesses of the different layers of the central Iberian lithosphere.

2. Geological and Geophysical Setting

2.1. The Central System and the Duero and Madrid Cenozoic Basins

[5] The Central System is a crustal pop-up with an uplifted Variscan basement. It shows a double vergence and thick-skin tectonics and has had a complex tectonic evolution

during the whole Cenozoic [Vegas *et al.*, 1990; Ribeiro *et al.*, 1990; de Vicente *et al.*, 1996; Andeweg *et al.*, 1999; de Vicente *et al.*, 2007]. Its deformation is asymmetric with a series of imbricated thrusts verging toward the Duero basin at its northern border and a single major thrust in the south with a tectonic transport toward the Madrid basin that accumulates much of the deformation at the southern border (Figure 1a). The intraplate basement uplift stretches for over 500 km to the Atlantic coast in Portugal, with an orientation ranging between E-W and NE-SW and an average width of 80 km. Today the mountain chain is over 2000 m, rising to a maximum height of 2592 m in the Sierra de Gredos and 2430 m in the Sierra de Guadarrama. These peaks show a vertical throw of ~2000 m and ~1400 m respectively over the Madrid Basin (Figure 1b). Mountain alignments are separated by small pop-downs and triangle zone like structures in which thin pre-tectonic Paleogene and Mesozoic sediments have been preserved. However, the Central System is not an ancient inverted Mesozoic rift [de Vicente *et al.*, 2009b; de Vicente and Vegas, 2009] and the deformation location must have a different origin.

[6] The Madrid and Duero basins show a parallel sedimentary development until the Eocene-Oligocene when they became separate basins with the uplift of the Central System [Santisteban *et al.*, 1996; Armenteros *et al.*, 2002; Alonso-Zarza *et al.*, 2004; Herrero *et al.*, 2010]. During the Tertiary they behave as flexural foreland basins and their border are mainly delimited by large crustal thrusts.

2.2. Cenozoic Deformations

[7] Kinematic reconstructions for Europe, Iberia and Africa during the alpine orogeny indicate that their relative movements had a general N-S and NW-SE orientation, with very slow E-W displacements during the Miocene [Rosenbaum *et al.*, 2002a, 2002b]. The emplacement of the Pyrenean thrusts began at the end of the Cretaceous and continued until the Oligocene in the eastern Pyrenees and until the Early Miocene in the central and western Pyrenees [Vergés *et al.*, 1995; Anadón and Roca, 1996]. In the Cantabrian Mountains, the prolongation of the Pyrenean orogen toward the west, tectonic activity started in the Paleocene-Eocene and continued until the Neogene [Boillot and Malod, 1988; Alonso *et al.*, 1996], and even migrated to the peninsular northwest in the late Oligocene where lasted into the late Miocene [Martín-González and Heredia, 2011]. Current seismicity is mainly concentrated in the Pyrenees and in the NW corner [Goula *et al.*, 1999; Martínez-Díaz *et al.*, 2006; Ruiz *et al.*, 2006; de Vicente *et al.*, 2007; Stich *et al.*, 2010]. The convergence processes in the south of the Iberia micro-continent started in the Late Cretaceous-Paleogene and had a deformation peak in the Early Late Miocene [Sanz de Galdeano, 1990; Martín-Algarra and Vera, 2004; Soto *et al.*, 2008]. Regional seismicity in the south is diffuse, so that is not possible to clearly define the plate limit between Iberia and Africa [Buforn *et al.*, 1988; Galindo-Zaldívar *et al.*, 1993; Stich *et al.*, 2005].

[8] The deformations in the Central System seem to have been caused by approximately NW-SE and N-S compressive stresses originated at the northern and southern Iberian plate borders during the Pyrenean collision (Paleogene) in particular and, to a lesser extent, the Betic (Neogene)

collision [Casas Sainz and Faccenna, 2001; de Bruijne and Andriessen, 2002; de Vicente *et al.*, 2007; de Vicente and Vegas, 2009]. Apatite fission track analysis shows four accelerated cooling events across the range [de Bruijne and Andriessen, 2000; de Bruijne and Andriessen, 2002]: Middle-upper Eocene, Lower-upper Oligocene, Lower Miocene and Upper Miocene-present-day. Plio-Quaternary neotectonic activity has been identified in the Madrid Basin, although the instrumental seismicity is low in the area [de Vicente *et al.*, 1996, 2007; Herraiz *et al.*, 2000]. The rate of Cenozoic shortening in the Guadarrama sector of the Central System is estimated to be ~20% [de Vicente *et al.*, 1996], and the current vertical throw between the Madrid and Duero basin basements and the topography of the Central System is ~6 km.

2.3. Structure, Composition and Thermal Flow of the Iberian Intraplate Lithosphere

[9] The lithosphere-asthenosphere limit of the Iberian Peninsula is located at a depth of almost 100 km [Banda *et al.*, 1981; Fernández *et al.*, 1998; Tejero and Ruiz, 2002; Spakman and Wortel, 2004]. The average thickness of the crust is around 30 km with different structural levels [Banda *et al.*, 1981; Suriñach and Vegas, 1988; ILIHA DSS Group, 1993]: An upper crust of 11–14 km, a middle crust of 7–12 km and a lower crust of 7–9 km. Below the Central System, however, the crustal thickness increases to 34–36 km as a result of convex bending in the upper crust and concave thickening in the lower crust (crustal pop-up structure) related to the alpine compression (Figure 1c) [Suriñach and Vegas, 1988; Vegas *et al.*, 1990; de Vicente *et al.*, 2004; Gómez-Ortiz *et al.*, 2005; de Vicente *et al.*, 2007]. The crust is partially covered by the Tertiary sediments of the Madrid and Duero basins. In the Duero basin, these reach a thickness of 2000 m in the western part of the Cantabrian thrusting front, whereas in areas near the Central System sedimentary thickness is over 2500 m [Alonso *et al.*, 1996; Pulgar *et al.*, 1996; Fernández-Viejo *et al.*, 2000; de Vicente *et al.*, 2004; Gómez-Ortiz *et al.*, 2005; de Vicente *et al.*, 2007]. In the Madrid basin, the sedimentary record has a maximum thickness of ~3400 m east of the Sierra de Guadarrama, which includes ~200 m of Cretaceous marine sediments [Racero, 1988; Querol, 1989; Gómez-Ortiz *et al.*, 2005; de Vicente *et al.*, 2007].

[10] The Tertiary continental sedimentary strata are siliclastic when connected to alluvial and fluvial processes on the edges and evaporitic in central areas [Mediavilla *et al.*, 1996; Santisteban *et al.*, 1996; Alonso-Zarza *et al.*, 2004]. The upper crust of the Central System consists of granodiorites, monzogranites, and leucogranites, which intrude in Precambrian-Palaeozoic metamorphic rocks that belong to what is called the Central Iberian Zone of the Iberian Variscan Massif [Villaseca *et al.*, 1999; Bea *et al.*, 2004]. The western part of the basement of the Central System contains large outcrops of granitoids, most of them generated during the Early Carboniferous-Permian period, forming one of Western Europe's main Variscan batholiths [Füster and Villaseca, 1987; Bea *et al.*, 2004]. The basement of the Duero and Madrid basins mainly consists of slates and quartzites, although plutonic rocks have also been identified, which mostly belong to the Western Asturian-Leonese Zone

Figure 2. Characteristics of the finite element mesh of the intraplate Iberian lithosphere (shown in Figure 1b): (a) laterally homogeneous and (b) laterally heterogeneous (vertical scale equals horizontal scale). The loads applied in the analysis of the deformation and the interfaces in which isostatic forces are estimated are also represented (black dashed lines).

and Central Iberian Zone [Banda *et al.*, 1981; Racero, 1988; Querol, 1989; Gómez-Ortiz *et al.*, 2005; Suarez Alba, 2007]. The composition of the middle crust is similar to that of the outcropping intrusive rocks [Banda *et al.*, 1981; Villaseca *et al.*, 1999], whereas the lower crust is felsic granulite [Villaseca *et al.*, 1999; Julià and Mejía, 2004]. Although no clear peridotite xenoliths have been found in the Central System, peridotite is presumed to make up the composition of the mantle [Bea *et al.*, 1999, 2004; Orejana *et al.*, 2006].

[11] The surface heat flow estimations in the central peninsula, drawn from a reduced number of measurements, oscillate between $\sim 60\text{--}80\text{ mW m}^{-2}$, with maximum values found near the Central System and minimum ones in the Duero and Madrid basins [Fernández *et al.*, 1998]. The high heat flow in the intraplate chain comes from a high concentration of radioactive elements in the granitic rocks of the crust [Fernández *et al.*, 1998; Villaseca *et al.*, 1999].

3. Modeling Approach

[12] The thermal and mechanical equations related to the numerical modeling have been solved with the software ANSYS (Swanson Analysis System, Inc.) based on the finite element method [Zienkiewicz and Taylor, 1994a, 1994b]. The mesh contains 10,030 elements (9,180 plane strain quadrilateral elements of eight nodes and 850 line elements of three nodes) which are interconnected by 27,989 nodes (Figure 2). The element size is finer toward the center of the model in order to get a more detailed solution. The quadrilateral elements have thermal and creep capabilities whereas the line elements are used across the non-vertical material interfaces to include the restoring force of isostasy. The temperature distribution has been calculated first by means of

a thermal analysis in steady state conditions because as there is no information about the temperature in the late Mesozoic we cannot include the temperature variation with time. The solved temperatures are then used in the structural analysis to derive nonlinear rheologies. The displacements are determined under static loading conditions, subdividing the applied load into a series of incremental steps. Accurate solutions have been obtained using a time step size between 1,000 and 100,000 years. The specific geometry, thermo-mechanical properties and boundary conditions are explained below.

[13] The finite element model that has been designed to study the Cenozoic deformations of the Iberian interior takes into account the geological and geophysical characteristics previously described. Its geometry represents a section of the Iberian lithosphere before the Cenozoic convergence that could extend from the Cantabrian-Pyrenean thrust-fault front to the Betic detachments (530 km long) in a transverse direction to the nucleation zone of the Cenozoic shortening in the Sierra de Guadarrama and parallel to the alpine compressive stresses (Figures 1b and 2). The topographic surface has been represented by a horizontal line because during the Upper Paleocene Cretaceous, Iberian topography was very gentle and close to sea level [Martín-Serrano *et al.*, 1996; de Vicente *et al.*, 2007]. Regarding its lithospheric structure, it has been assumed that the characteristics were similar to those of the present [Banda *et al.*, 1981; Suredach and Vegas, 1988; ILIHA DSS Group, 1993]: a 13-km upper crust, a 9-km middle crust, an 8-km lower crust, and a 70-km lithospheric mantle.

[14] The mechanical behavior of each layer fits the power law creep dislocation and the thermal analysis is based on descriptions of stable geotherms, such that heat is transported by conduction, and the radiogenic heat production shows

Table 1. Geometry, Loads, and Properties (Rheological and Thermal Parameters) of the Lithospheric Units Differentiated in the Deformation Model of the Iberian Lithosphere^a

	Thickness ^b (km)	Density ^c (kg/m ³)	Heat Production ^d (μWm ⁻³)	Elastic Parameters ^e		Creep Parameters ^f				
				Young's Modulus (Pa)	Poisson's Ratio	Lithology	A (Mpa ⁻ⁿ s ⁻¹)	n	E (kJ mol ⁻¹)	Conductivity ^g (W m ⁻¹ K ⁻¹)
Upper crust										
Central System	13	2670	3.3 exp ^(-y/12)	75·10 ⁹	0.26	dry granite	1.8·10 ⁻⁹	3.2	123	2.5
	13	2670	3.3 exp ^(-y/12)	75·10 ⁹	0.26	wet granite	2.0·10 ⁻⁴	1.9	137	2.5
Basins	13	2780	Duero: 1.8 exp ^(-y/12) Madrid: 2.5 exp ^(-y/12)	75·10 ⁹	0.26	dry quartzite	6.7·10 ⁻⁶	2.4	156	2.5
	13	2780	Duero: 1.8 exp ^(-y/12) Madrid: 2.5 exp ^(-y/12)	75·10 ⁹	0.26	wet quartzite	3.2·10 ⁻⁴	2.3	154	2.5
Middle crust	9	2800	Exponential decay from upper crust	91·10 ⁹	0.27	quartzdiorite	1.3·10 ⁻³	2.4	219	2.5
Lower crust	8	2900	0.8 exp ^(-y/12)	114·10 ⁹	0.26	felsic granulite	8.0·10 ⁻³	3.1	243	2.1
Lithospheric mantle	70	3300	0.02 exp ^(-y/12)	172·10 ⁹	0.28	dry peridotite	2.5·10 ⁴	3.5	532	3.4
	70	3300	0.02 exp ^(-y/12)	172·10 ⁹	0.28	wet peridotite	2.0·10 ³	4.0	471	3.4

^aA 3.5·10⁻⁵ K⁻¹ value has been used for the thermal volumetric expansion coefficient [Tejero and Ruiz, 2002]. A, pre-exponent constant; n, power law exponent; E, activation energy.

^bValues taken from Tejero and Ruiz [2002].

^cValues taken from Gómez-Ortiz et al. [2005].

^dValues taken from Fernández et al. [1998], Villaseca et al. [1999], and Tejero and Ruiz [2002].

^eValues calculated from the seismic-wave velocities [Banda et al., 1981] and the densities [Gómez-Ortiz et al., 2005] (see section 3).

^fValues taken from Ranalli [1995].

^gConductivity is the thermal parameter. Values taken from Fernández et al. [1998] and Tejero and Ruiz [2002].

exponential decay with depth [Ranalli, 1995; Turcotte and Schubert, 2002]:

$$\sigma_d = \left(\frac{\dot{\epsilon}}{A}\right)^{(1/n)} \exp^{(Q/nRT)}$$

$$k \frac{d^2 T}{dy^2} = -S$$

$$S(y) = S_0 \exp^{(-y/D)}$$

where σ_d is the differential stress, $\dot{\epsilon}$ is the strain rate, A and n are the material constants, Q is the activation enthalpy for creep, R is the gas constant, T is the temperature, k is the thermal conductivity, S is the heat production, S_0 is the heat production at the top of the layer, y is the depth, and D is the exponential decay constant. The thermomechanical properties have been defined taking into account different studies on the Central System (Table 1) [Banda et al., 1981; Fernández et al., 1998; Villaseca et al., 1999; Tejero and Ruiz, 2002; Gómez-Ortiz et al., 2005]. The elastic parameters (Young's modulus, E , and Poisson's ratio, ν) have been calculated based on seismic-wave velocities [Banda et al., 1981] and on densities [Gómez-Ortiz et al., 2005] of the peninsular center according to these equations [Jaeger and Cook, 1969; Schön, 2004]:

$$\nu = \frac{(\theta - 2)}{2(\theta - 1)}$$

$$E = \rho V_P^2 \left(\frac{(1 + \nu)(1 - 2\nu)}{(1 - \nu)} \right)$$

where $\theta = (V_P/V_S)^2$, V_P and V_S are respectively the P and S wave velocities, and ρ is the density. To calculate the

lithostatic load of the Duero and Madrid basins and correctly account for the isostatic effect, it was also necessary to define the density of the Tertiary sediments (2300 kg m⁻³) and of the asthenospheric mantle (3250 kg m⁻³).

[15] Three homogeneous lithospheres are considered (Figure 2a and Table 1): the Duero type, the Central System type, and the Madrid type, with different densities and parameters for the creep in the upper crust and thermal structure. A heterogeneous lithosphere is also considered (Figure 2b and Table 1), in which from NW to SE, a Duero-type lithosphere gives way to a Central System type and to a Madrid type. This differentiation highlights the zoning of materials from the Iberian Massif caused by the Variscan Orogeny (Figure 1a): metamorphic-sedimentary rocks from the Western Asturian-Leonese Zone in the north (Duero basin), metamorphic and plutonic rocks from the Central Iberian Zone in the south (Madrid basin), in addition to the intense plutonism characteristic of the Central Iberian Zone (Central System). The northern border of the Central System has a dip that is slightly less pronounced than at the southern edge, in line with results from gravimetric models [Gómez-Ortiz et al., 2005; de Vicente et al., 2007]. The dry and wet behavior of the upper crust and the lithospheric mantle has also been analyzed (Table 1).

[16] The deformation of the lithosphere during the Cenozoic has been estimated considering (Figure 2): (a) the lithostatic load of the current thickness of Cenozoic sediments from the Duero and Madrid basins and (b) the tectonic shortening on the laterals of the section, which can simulate the thrusting from the edges of the Iberian micro-continent so as to shorten the length of the model by 10%. Likewise, the isostatic displacements resulting from density contrasts between the various lithospheric layers have been included, as at the base of the lithosphere to simulate lithosphere-asthenosphere interactions [Wu, 2004]. Two situations have been evaluated, one in which only the horizontal shortening has been included and a second in which sediment weight

Figure 3. (a) Temperature in the crust-mantle limit. (b) Surface heat flow. D, Duero-type lithosphere; M, Madrid-type lithosphere; CS, Central System-type lithosphere; H, heterogeneous lithosphere. The black dashed line shows the current surface heat flow from *Fernández et al.* [1998].

and tectonics are introduced simultaneously. The complex Cenozoic evolution has been simplified, so as shortening and sedimentary load have been gradually applied over 60 Ma. Under these conditions, the average convergence velocity (~ 0.9 mm/yr) is lower than the estimates at the boundaries of Iberia in the periods of maximum deformation (e.g., 6 mm/yr in the Eastern Pyrenees during early and middle Eocene [Vergés *et al.*, 1995]), and the convergence velocities between plates (e.g., 8 mm/yr in Iberia during lower Oligocene to lower Miocene [Rosenbaum *et al.*, 2002a]). However, low horizontal velocities (< 1 mm/yr) are obtained in the center and east of the Iberian Peninsula from numerical models of deformation in the current Eurasian-African boundary [Jiménez-Munt and Negredo, 2003]. On the other hand, taking into account the length of the lithosphere that has been modeled, the average strain rate is $5.2 \times 10^{-17} \text{ s}^{-1}$, and current estimates in the vicinity of the Iberian Peninsula ranges from 10^{-16} to 10^{-17} s^{-1} [Jiménez-Munt and Negredo, 2003]. The thermal structure has been modeled based on the radiogenic heat production (the exponential decay constant reaching a 12 km value), a 15°C temperature at the surface, and a 1350°C temperature at the base [O'Reilly *et al.*, 2001; Turcotte and Schubert, 2002; Tejero and Ruiz, 2002].

4. Thermal Models

[17] To analyze the thermal structure of the four types of lithospheres with different rheologies, the temperature at the crust-mantle limit and the surface heat flow have been represented (Figure 3). Both parameters remain constant through the entire section when it is homogeneous, but they change laterally if the modeled lithosphere is heterogeneous. The temperature range in the Moho is $\sim 600^\circ\text{C}$, and the surface heat flow is $\sim 65 \text{ mW m}^{-2}$. For both parameters, the minimum values are found in the Duero-type lithosphere (Western Asturian-Leonese Zone), whereas the maximum values are reached in the Central System-type lithosphere (Central Iberian Zone). The Madrid-type lithosphere, in the Central Iberian Zone but with a much lower volume of plutons, presents intermediate characteristics. Although these results are consistent with thermal studies conducted in the area [Fernández *et al.*, 1998; Tejero and Ruiz, 2002;

Fernández *et al.*, 2008], several discrepancies can be observed, due to differences in the lithospheric structures and thermal parameters that have been used.

5. Deformation Models

[18] This section analyzes the basic factors that could have controlled the deformation of the distinct units of the Iberian lithosphere during the Cenozoic and led to its current configuration: rheology (composition and temperature variations), loads (tectonic shortening and Tertiary basin weights), and geometry (dip of the Central System margins and thickness of the lithospheric layers). Specifically, four series of models have been designed, differentiated by their thermo-mechanical properties, and which have been resolved in two ways: by shortening the model by 53 km horizontally to simulate collisions between Europe-Iberia and Iberia-Africa (T models) and by further including the load of the Duero and Madrid basin sediments (TS models). In this way, we can determine if the basin depth is caused by tectonic shortening or it also needs the weight of the sediments. The horizontal shortening in the Central System after deformation has been calculated in each model by means of the classic strain equation $\epsilon = (x - x')/x$ [Turcotte and Schubert, 2002] where x is the initial length at surface (that is 100 km, Figure 2) and x' is the final length of the range measured in the horizontal and expressed as a percentage.

5.1. Homogeneous Composition and Temperature (Homo Series)

[19] The three types of homogeneous lithospheres have been studied (Duero, Madrid, and Central System types), for both dry and wet conditions in the crust and in the mantle.

5.1.1. T-Homo Models

[20] The deformation occurs as a result of a homogeneous thickening of the lithosphere down to a ~ 120 km average depth. The average horizontal shortening found in the Central System area is 10%. The topography rises to 500–700 m and the Moho goes down to ~ 3500 , ~ 4000 m deep (Figure 4a). The highest values are obtained in Central System-type lithospheres, with a less dense, warmer, granitic upper crust. Regarding the effect of the wet and dry rheologies on the

Figure 4. Vertical displacements of the surface and of the crust-mantle limit of the six lithospheres with homogeneous composition and temperature after being deformed by: (a) tectonic shortening (T-homo model) and (b) tectonic shortening and sedimentary load (TS-homo model). CS, Central System sector; CS-dd, Central System-type lithosphere with dry upper crust and lithospheric mantle; CS-ww, Central System-type lithosphere with wet upper crust and lithospheric mantle; DB-dd, Duero-type lithosphere with wet upper crust and lithospheric mantle; MB-dd, Madrid-type lithosphere with dry upper crust and lithospheric mantle; MB-ww, Madrid-type lithosphere with wet upper crust and lithospheric mantle.

deformation, the major increases are produced with the granite dry parameters and the quartzite wet parameters.

5.1.2. TS-Homo Models

[21] In this case, a flexural harmonic folding develops, owing to the bending of the entire crust from the weight of the Duero and Madrid basins (Figure 4b), remaining the average 10% shortening in the Central System sector. The base of the lithosphere, which is located at a -121 km average depth, shows soft undulations (600 m average amplitude). The topographic surfaces located in positive heights would reproduce the elevations of a mountainous landscape with elevations below 1000 m. On the contrary, areas with negative heights represent the basement beneath the Tertiary sediments, in which the vertical displacement increases from the laterals of the model to the mountain range, where the depocenters would be located. In the crust-mantle limits, the height differences exceed 5000 m but they have shapes similar to topographic surfaces.

5.2. Homogeneous Composition and Heterogeneous Temperature (Hete-T^a Series)

[22] A lithosphere deformation with a uniform composition but with a heterogeneous thermal structure (colder in the Duero basin, warmer in the Central System, and intermediate in the Madrid basin) has been modeled. Four different situations have been considered, by combining an upper crust composed of granite (Central System type) or quartzite (basins type) with a dry or wet rheology in the upper crust and lithospheric mantle.

5.2.1. T-Hete-T^a Models

[23] The lithosphere layers undergo a non-harmonic folding (Figure 5a), which entails an 11% shortening in the Central System sector. The base of the lithosphere experiences a concave bending within a -116 km to -124 km depth interval. In the central part of the surface of the model,

a gentle antiform develops, which is delimited by two small synforms, whereas a synform is created in the Moho, with a crustal root shape and average amplitude of 2 km. There are no great differences between the various compositions of the crust, except in the central part of the model with dry quartzite. However, the lateral temperature variations condition vertical displacements in the three sectors in such a way that their magnitudes are higher with higher geotherms (Central System > Madrid Basin > Duero Basin).

5.2.2. TS-Hete-T^a Models

[24] The topography caused by tectonic deformation in the T-hete-T^a models changes because of the flexural response (bending) to the sedimentary loads, although the resulting shortening in the Central System is equivalent. The amplitude of the two synforms increases in the topographic surface (Figure 5b), outlining the basements of the two basins similarly to that of the TS-homo models (Figure 4b). This deformation can also be found in the Moho under the basins, and only the model with dry quartzite develops an antiform in the central part of this limit. The three remaining lithospheres show some very gentle synforms. The base of the lithosphere is also structured into a large flexural synform.

5.3. Heterogeneous Composition and Homogeneous Temperature (Hete-Comp Series)

[25] In these models, the upper crust composition varies laterally (quartzitic in the laterals and granitic in the central part), but the temperature remains uniform. Only the two extreme thermal structures have been evaluated (Duero type and Central System type). Combining these models with dry and wet conditions of the upper crust, it results in four different initial situations.

5.3.1. T-Hete-Comp Models

[26] With a composition contrast, a non-harmonic folding also occurs, but it shows two clearly different deformation

Figure 5. Vertical displacements of the surface and of the crust-mantle limit of four lithospheres with a uniform composition but thermally heterogeneous after being deformed by: (a) tectonic shortening (T-hete- T^a model) and (b) tectonic shortening and sedimentary load (TS-hete- T^a model). Vertical displacements of the surface and of the crust-mantle limit of four lithospheres with a laterally heterogeneous composition and homogeneous temperature after being deformed by (c) tectonic shortening (T-hete-comp model) and (d) tectonic shortening and sedimentary load (TS-hete-comp model). CS, Central System; DB, Duero Basin; dd, dry upper crust and lithospheric mantle; ww, wet upper crust and lithospheric mantle.

styles according to the wet or dry rheological conditions (Figure 5c). The thermal factor only introduces slight variations in the displacement magnitude, displacements being somewhat more important when the thermal structure is warmer. In dry conditions, the horizontal shortening increases by 19% in the central sector, which generates a very abrupt intraplate shaping consisting of two elevated zones (3284 m average maximum height) delimiting a depressed zone equivalent to an intramountain basin (1203 m average minimum height). Its crust-mantle limit shows a concave bending with 2.2 km average amplitude. In wet conditions, the surface topography is similar to the two models with heterogeneous temperature (T-hete- T^a models, Figure 5a), although the crust-mantle limit presents gentle undulations. The base of the lithosphere moves to a -120 km average depth and develops a ~ 2.0 km synform with dry parameters or slight undulations (~ 0.6 km amplitude) with wet parameters.

5.3.2. TS-Hete-Comp Models

[27] When a sedimentary load is introduced, bending of the basins occurs again but with a greater magnitude for dry models (Figure 5c). At the crust-mantle limit, as was the case with the TS-hete- T^a models (Figure 5b), the crustal root diminishes or the antiform increases.

5.4. Heterogeneous Composition and Temperature (Hete Series)

[28] In this last series, we have deformed a lithosphere whose composition and temperature vary from the NW to SE of the profile (Figures 2b and 3 and Table 1). Two intermediate rheological situations have been taken into consideration: a dry upper crust with a wet lithospheric mantle and a wet upper crust with a dry lithospheric mantle. Once the rheology and load effects on deformation were studied, two other series of models were conducted to estimate the importance of the lithospheric initial geometry.

Figure 6. Vertical displacements of the surface and of the crust-mantle limit of four lithospheres with heterogeneous composition and temperatures after being deformed by (a) tectonic shortening (T-hete model), (b) tectonic shortening and sedimentary load (TS-hete model), (c) tectonic shortening and sedimentary load, from a lithosphere in which the dip of the edges of the Central System has been modified (dip-TS-hete model; the dd and ww curves of the previous TS-hete model are also included in this graph), and (d) tectonic shortening and sedimentary load, from a thinned lithosphere (thickness-TS-hete model). CS, Central System; dd, dry upper crust and lithospheric mantle; dw, lithosphere with dry upper crust and wet lithospheric mantle; wd, lithosphere with wet upper crust and dry lithospheric mantle; ww, wet upper crust and lithospheric mantle. The black dashed lines show the crustal thickness (1), the Duero basin sedimentary thickness (2), the Madrid basin sedimentary thickness (3) and the topography (4) from the gravity model in the work by *de Vicente et al.* [2007] (Figure 1c).

5.4.1. T-Hete Models

[29] Two deformation styles can be observed again, conditioned this time by the dry or wet rheology of the upper crust. The shortening in the central sector amounts to 22% with dry crusts and 12% with wet crusts. The resulting mountain uplift, its crustal root, and the lithospheric base experience their most important displacements with dry crusts (Figures 6a and 7a). If we compare the results of this tectonic-load folding with those of previous series (Figures 5a–5c), we can observe that thermal contrasts mainly affect the Moho topography and the lithosphere, whereas lateral composition variations in the upper crust condition the topography of the surface of the model.

5.4.2. TS-Hete Models

[30] General observations repeat themselves when the sedimentary load is included (Figures 6b and 7b): (a) the deformation style is conditioned by upper-crust creep and (b) the deformation in the surface and in the Moho depends on compositional and temperature contrasts, respectively. The shortening values on the intraplate range surface are also similar, and the major differences can be found in the lower land elevations, the deeper subsidence below the basins, and the decrease of crustal root amplitude.

5.4.3. Dip-TS-Hete Models

[31] An initial geometry curved on the Central System borders is included, similar to the present geometry [*Racero*,

Figure 7. Final geometry of the heterogeneous lithosphere (with crust and mantle in dry conditions) after 60 Ma deformation: (a) tectonic shortening (T-hete model), (b) tectonic shortening and sedimentary load (TS-hete model), (c) tectonic shortening and sedimentary load in a lithosphere with variable dips along the north and south border of the Central System (dip-TS-hete model), and (d) tectonic shortening and sedimentary load in a thinned lithosphere (thickness-TS-hete model). The gray lines indicate the initial geometry.

1988; Querol, 1989; Gómez-Ortiz *et al.*, 2005; de Vicente *et al.*, 2007]. Only cases with a dry or wet rheology, both in the upper crust and the lithospheric mantle, were modeled. They are represented together with the two previous heterogeneous models. The deformation style is maintained and only vertical displacements slightly decrease (Figures 6c and 7c).

5.4.4. Thickness-TS-Hete Models

[32] Because of the thickening during deformation (3–7 km), the final thickness of the crust and lithosphere prove

to be excessive compared to present estimations (Figure 6b). Thus, to adjust the deformed geometry, the initial thickness of the crust and the lithosphere was reduced to 27 and 85 km respectively. Only two situations have been analyzed in which the crust and lithospheric mantle have parameters of creep, either dry or wet. The topographic surfaces are almost entirely adjusted to the T-hete models final geometries, and only central topography show slightly lower elevations (Figure 6d). However, in the crust-mantle limit, a general decrease of the vertical displacements and crustal

root amplitude occurs. The average depth of the lithosphere-asthenosphere limit also diminishes (-103 km, Figure 7d).

6. Discussion

6.1. Effect of Temperature and Composition on Intraplate Deformation

[33] The dislocation creep activates thermally in such a way that a rise in temperature entails minor strength to the deformation [Ranalli and Murphy, 1987; Burov and Diament, 1995]. In this regard, differences fewer than 100 m in the vertical displacements have been observed in lithospheres without rheology contrasts (compare DB-dd curves with MB-dd or DB-ww curves with MB-ww in T-homo series, Figure 4a). However, when introducing lateral thermal variations, the major deformation is concentrated in the warmest central part, and the differences among sectors rise up to several kilometers (T-hete-T^a, Figure 5a). The thermal weakening of the Central System sector can be linked to the substantial production of crustal radioactive heat resulting from intense Late Palaeozoic magmatism, which would be enough to localize the deformation. Moreover, these models confirm that small lateral contrasts in geothermal distribution can influence deformation as a crucial factor [Fernández and Ranalli, 1997; Burov, 2003]. Thus, during shortening, deformation develops as a result of a homogeneous thickening mechanism or pure shear in those lithospheres that are completely homogeneous (homo series, Figure 4a). In contrast, the lateral thermal variations effectively contribute to the large-scale folding of the crust (hete-T^a series, Figure 5a). Similar results have been described in models of intraplate deformations in the Flinders Ranges in Southern Australia [Célérier et al., 2005].

[34] The rheological behavior of the lithosphere is also strongly controlled by its composition [Ranalli and Murphy, 1987; Kohlstedt et al., 1995; Afonso and Ranalli, 2004]. Its effect is similar to that of temperature concerning the deformation of lithospheres that are laterally homogeneous or heterogeneous. If we compare the behavior among models with various upper crust compositions, we can again identify an oscillation of a few hundred meters in the magnitude of displacements, those being somewhat more important in granitic crusts (Figures 4a, 4b, and 5a). However, if the lithospheres contain lateral composition contrasts, the range of values between the different sectors can increase by an order of magnitude (hete-comp series, Figure 5c). Introducing these compositional variations also makes possible the general folding of the lithosphere layers by horizontal shortening.

[35] The deformation style is further conditioned by the dry or wet conditions of the materials. When considering lithospheres with a laterally heterogeneous composition, those with dry rheologies experience a more substantial deformation than the wet ones (Figures 5c and 5d; hete-comp series) because of the high strength contrast between the dry quartzite and the dry granite (Table 1). For this reason, in the central sector, a major mechanical weakening takes place, favoring the development of an abrupt topography on the surface. These results highlight the fact that lateral heterogeneities in lithological properties modify the strength of the lithosphere, concentrating the deformation in specific areas

[Fernández and Ranalli, 1997; Dixon et al., 2004; Yang et al., 2008].

6.2. Cenozoic Deformations in the Interior of the Iberian Lithosphere

[36] This study did not analyze complex rheologies such as thermomechanical coupling, erosion-sedimentation processes, brittle-elasto-ductile nonlinear rheologies, friction along faults, or the discontinuity of tectonic processes [Burov et al., 1998; Gerbault, 2000; Cloetingh et al., 2002; Célérier et al., 2005; Burg and Schmalholz, 2008]. However, lithospheres with marked lateral thermomechanical contrasts and dry crusts (TS-hete series, Figures 6b–6d) allow us to explain the main structural features of central Iberia: (a) the $\sim 20\%$ shortening in Guadarrama area [de Vicente et al., 1996], (b) the maximum heights of the range at ~ 2500 m, (c) the ~ 2500 m thickness of Tertiary sediments in the Duero Basin's depocenter and ~ 3200 m thickness in the Madrid Basin's depocenter [Racero, 1988; Querol, 1989; Gómez-Ortiz et al., 2005; de Vicente et al., 2007], and (d) the 3–6 km long crustal root under the mountain system [Suriñach and Vegas, 1988; ILIHA DSS Group, 1993; Muñoz-Martín et al., 2004; Gómez-Ortiz et al., 2005; de Vicente et al., 2007].

[37] The intraplate topography of the models represents the set of interior mountains and basins related to the uplifting of the Central System. The final heights of the range in the lithospheres with dry crusts are somewhat higher than those currently observed (Figure 6d), although much lower than those deduced from apatite fission tracks analysis [de Bruijne and Andriessen, 2000, 2002]. In the Sierra de Guadarrama, a maximum uplifting of ~ 6 km has been suggested, which would have begun during the Eocene with the Pyrenean deformations, although the major part could have taken place beginning in the Pliocene as a consequence of the Betic compressions and in combination with denudation processes [de Bruijne and Andriessen, 2000, 2002]. Because ignoring the effects of erosion when modeling generates higher topographies [Burg and Podladchikov, 2000], the reliefs obtained in the deformation models of the Iberian lithosphere could represent a maximum estimate. Although the limitations and conditioning inherent to both methods make any comparison difficult, the valuations of uplift based on fission tracks do not seem justified based on the vertical displacements predicted in this study. The uplift overestimation observed in fission tracks with regard to finite element modeling could be caused by the fact they make thermal cooling equivalent to uplifting without taking into account the characteristic hydrothermal processes of the Central System [González-Casado et al., 1996; Tornos et al., 2000; Martín-Crespo et al., 2004] that can reset this low-temperature technique. However, the low-temperature thermochronology results in mountain ranges of the NW Iberian Peninsula, with Cenozoic tectonic history and heights similar to the Central System, reveals uplifts that are consistent with the vertical displacements obtained in this work. Specifically, uplifts of 1,800–3,100 m have been recorded in the western lateral termination of the Alpine-Pyrenean Orogen (Cantabrian Mountains and Galaico-Leoneses Mountains) [Martín-González et al., 2011].

[38] The flexure of the basement of Tertiary basins can be related to the horizontal shortening process itself as it has

been observed in the Himalayas range [Burg and Podladchikov, 2000]. When applying the tectonic shortening, if the lithospheres show thermal and/or compositional heterogeneities, synforms adjacent to the antiformal rise develop at the surface (Figures 5a–5c and 6a). However, to obtain basement depths similar to the current ones, it is also necessary to include the sedimentary load (Figure 6d). Studies on the flexure of the Ebro (eastern continuation of the Duero Basin) and Guadalquivir basins (south of the Madrid Basin) also depend on these loads to explain the evolution of their basements [Gaspar-Escribano *et al.*, 2001; Garcia-Castellanos *et al.*, 2002].

[39] A high strength in the upper mantle is necessary for the thickening associated with orogenic processes to be generated [Burov and Watts, 2006; Yang *et al.*, 2008]. However, the lateral strength contrast in the mantle favored by temperature variations, aside from isostatic compensation processes, makes possible the development of crustal roots even with wet mantles (Figure 6b). A question that should be considered when adopting a weak- (hydrous) or strong- (anhydrous) mantle rheology is its geodynamic implication. Accordingly, studies of Variscan calc-alkaline gabbroid intrusions indicate that the lithospheric mantle under the Central System must have been heterogeneous [Orejana *et al.*, 2009]. However, although during the Mesozoic no important magmatic events have been recorded in the area of study, the effects of the rifting of the Tethys Sea and the Atlantic Ocean on mantle rheology and on the continuity of its characteristics up to the present-day should be considered. For these reasons, it seems difficult to limit the state of mantle hydration to dry or wet conditions, and deformation models in which both situations have been considered may be equally valid.

6.3. Initial Configuration of the Iberian Lithosphere

[40] When decreasing the dip of the Central System borders toward the surface (Figure 7c) these limits acquire, after shortening, an even smaller inclination than that interpreted by geophysical methods [Racero, 1988; Querol, 1989; Gómez-Ortiz *et al.*, 2005; de Vicente *et al.*, 2007]. This variant was introduced to consider an initial geometry similar to the present one, but it seems appropriate to assume that the contacts were probably strongly inclined in the early Tertiary. When rectilinear limits are simulated, they undergo a slight decrease of the dip close to the surface when the deformation ends, in agreement with geophysical interpretations and geological observations. That increase of the inclination with depth in the contacts between the mountain uplift and the sedimentary basins has been related to a tectonic change from a strike-slip regime during the Eocene to a compressive regime during the lower Oligocene-Miocene [de Vicente *et al.*, 2007]. However, just only the shortening of the lithosphere by thermomechanical contrasts efficiently generate this type of structural configuration.

[41] In contrast, when reducing the initial thickness of the model (Figures 6d and 7d), we achieved a more accurate adjustment of the final thickness of the crust (~30 km in the basin areas and ~35 km in the Central System area) and of the lithosphere (100–108 km). The Cretaceous paleogeography of the Central System was characterized by the transition from a marine environment in the east to the completely

emerged sectors in the west [Martín-Serrano *et al.*, 1996; de Vicente *et al.*, 2007]. This scenario can be compared to today's Mediterranean coastal zone of the Iberian Peninsula, conditioned by the Tertiary extension of the Valencia Trough and the Alboran Basin. The average crustal and lithospheric thicknesses along the coast are ~25 and ~75 km respectively [Vergés and Fernández, 2006; Díaz and Gallart, 2009]. Therefore the lithospheric structure of the peninsular interior at the end of the Mesozoic could have presented smaller thicknesses to those found today.

6.4. Deformation Mechanisms of the Iberian Lithosphere

[42] Present-day crust and lithospheric structure of the Iberian Peninsula can be explained based on the shortening of a heterogeneous thinner lithosphere (Figure 6d) from the Cantabrian-Pyrenean and Betic limits during the Cenozoic through overlapping mechanisms of thickening and crustal and lithospheric folding. The folding of the topographic surface can be altered by lateral compositional variations in the upper crust, whereas the Moho flexure mainly depends on temperature contrasts. Because the wavelength of the folds in the upper crust is shorter than in the mantle, we could infer a crust-mantle decoupling during the convergent tectonic activity [Cloetingh *et al.*, 1999, 2002]. The results obtained support and unify the suggested models of folding and fracturing of the upper crust and ductile thickening of the lower crust [Vegas *et al.*, 1990; de Vicente *et al.*, 1996] and lithospheric folding [Cloetingh *et al.*, 1999; Casas Sainz and Faccenna, 2001; Cloetingh *et al.*, 2002; de Vicente *et al.*, 2007; de Vicente and Vegas, 2009]. It does not seem necessary to resort to intracrustal detachments from the Betics to the Central System [Ribeiro *et al.*, 1990; Gómez-Ortiz *et al.*, 2005], although to analyze its importance, it would be necessary to include more complex rheologies that could make strain localization possible.

[43] Cloetingh *et al.* [2002] have demonstrated that the intraplate lithospheric folding in Iberia contributes to current high and anomalous topography of the peninsula. The process is conditioned both by compressions coming from the north and the south and by surface erosion. However, their results show a harmonic folding of the entire crust, and a characteristic feature of the crust in the peninsula is its variation in thickness, with remarkable thickenings under the Alpine deformation belts. Analogue models also support buckling of the whole Iberian lithosphere, modified by brittle deformation of the upper crust and ductile flow in the lower crust, as a consequence of convergence [Fernández-Lozano *et al.*, 2011]. Other authors suggest rheological variations acquired during different tectonothermal events as an alternative mechanism [Vergés and Fernández, 2006]. The deformation patterns obtained in the present work, which simulate both surface relief and Moho topography during the Cenozoic convergence (Figure 6), are conditioned by (a) the initial thickness of the crust and lithosphere, (b) the deformation by pure shear of the various lithologies, (c) the temperature and composition of lateral variations, (d) the isostatic buoyancy processes, and (e) the folding of the different lithospheric layers. Thus, in order to explain the structure of the Iberian lithosphere in addition to the folds in the crust and in the lithosphere, the lithospheric

heterogeneities caused by the different tectonothermal processes since the Late Palaeozoic are necessary for the irregular thickening of the crust.

[44] Not only the Iberian but also the North African topography have been related to the development of the lithospheric folding generated during the Cenozoic [Cloetingh *et al.*, 2002; Teixell *et al.*, 2003; Casas-Sainz and de Vicente, 2009; Muñoz-Martín *et al.*, 2010]. In the Iberian case, the Pyrenean collisional limit extends from the Mediterranean Sea to the Atlantic Ocean, even to the continental-oceanic crustal margin in the Galicia Bank [de Vicente *et al.*, 2009a]. In the same way, the area of intraplate deformation in the center of the Iberian micro-continent extends from the Mediterranean (Iberian Chain) westward to the Spanish-Portuguese Central System and may continue until the Extremadura spur in the oceanic crust. The same argument can be used until the south of the Atlas range. These lithospheric deformations (folds and/or thickenings) occurred in ancient Mesozoic rifts (Pyrenees and Iberian Chain) in the east, with a complete or partial tectonic inversion. In the case of the Central System, it seems that the location of these lithospheric deformations was due to the distinct rheology inherited from the Variscan orogeny (in contrast to the West Asturian-Leonese and Central Iberian zones) and by the concentration of granitic rocks within the Central Iberian Zone (with the associated thermal effect).

7. Conclusions

[45] Taking into account the limitations imposed by the incomplete knowledge of the structure of the intraplate Iberian lithosphere and by the adopted simplifications, the integration of the available information based on the finite element technique has proven essential for modeling several of the most relevant geodynamic processes of the central Iberian Peninsula during the Cenozoic. The deformation models have enabled us to a) analyze the effects of tectonics and sedimentary load and b) simulate the current structure of the crust and lithosphere of the Central System and the Duero and Madrid basins.

[46] The thermomechanical heterogeneities generate variations in the strength of the crust and of the lithospheric mantle, which favor its irregular thickening and folding by tectono-sedimentary loads. The thermal contrasts mainly affect the crust-mantle limit, whereas the compositional contrasts in the upper crust determine the surface topography.

[47] The deformation of a thinner Late Mesozoic lithosphere with a heterogeneous Variscan rheology because of the shortening from the borders of Pyrenean and Betic collisions during the Cenozoic (and to the Tertiary sedimentary load) is in agreement with the general features of the current intraplate structure.

[48] The deformation model of the Iberian intraplate lithosphere corresponds in surface to a large antiform with minor undulations, which resembles the current topography in ranges and intramountain basins of the Central System, and two synclinal troughs, which delineate the basements of the Duero and Madrid basins. To isostatically compensate this feature, a synform takes shape in the crust-mantle limit, i.e., in the crustal root of the Central System, which generates the thickening of the lower crust. Finally, the base of the lithosphere is deformed through a large concave flexure.

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