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Structure of an intraplate fold-and-thrust belt: The Iberian Chain. A synthesis

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ABSTRACT

The Iberian Chain is a complex intraplate fold-and-thrust belt resulting from the convergence between the Eurasian, Iberian and African plates during the late Eocene to the Miocene. The main trend of its contractional structures is NW-SE, but E-W, NE-SW and N-S-trending structures are also present. The boundaries of the chain with its surrounding foreland basins are always thrusts. The North-Iberian Thrust separates the chain from the Ebro Basin to the North, while the Serranía de Cuenca Thrust makes the SE boundary of the chain, separating it from the Tajo Basin and La Mancha foreland areas. Between these thrusts, the contractional structure is basement-involved, while South of the Serranía de Cuenca Thrust only Mesozoic and Cenozoic rocks are involved in the thrust-system, detached in the evaporitic Triassic materials. Two parts can be differentiated considering the major structure of the chain. The western and central areas hold two major anticlinoriums separated by the Almazán Synclinorium. East of the Teruel Depression, E-W-striking N-verging thrusts in the North, and NW-SE-striking S-verging thrusts in the center and South are the dominant structures.

The crust thickened during the Cenozoic contraction generating a mean crustal thickening of about 5km. The horizontal shortening obtained from cross-sections is 32km, and from a density-gravity section of 57.5km. These two values may be considered end values.

The relief of the Iberian Chain has a strong areal coincidence with the contractional structures and the thickened crust, indicating that they are genetically related.

KEYWORDS **Thrust belt. Intraplate. Crustal thickness. Iberia.**

INTRODUCTION

From a geologic point of view, the Iberian Chain (Figs. 1; 2) includes all the contractional structures generated during the inversion of the Iberian Mesozoic basins. These structures developed within the Iberian Plate during the Cenozoic, resulting from the convergence of the Eurasian and African plates, which also generated the Pyrenean-Cantabrian Chain and the Betic-Balearic Chain at the North and South borders of the Iberian Plate. The Iberian Chain is surrounded by the Ebro, Duero and Tajo Cenozoic basins to the North, North-West and West, the Ebro and Duero basins separating it from the Pyrenean-Cantabrian Chain and being their common foreland basins (Hahne, 1930a, b; Brinkmann, 1960; Álvaro *et al*., 1979; Guimerà, 1983, 1984, 2004; Guimerà and Álvaro, 1990; Álvaro, 1995; Salas *et al*., 2001; Guimerà *et al*., 2004; De Vicente *et al*., 2009).

Along the Iberian Mediterranean coast, NE-SW coastal rifts (Fig. 2), related to the opening of the western

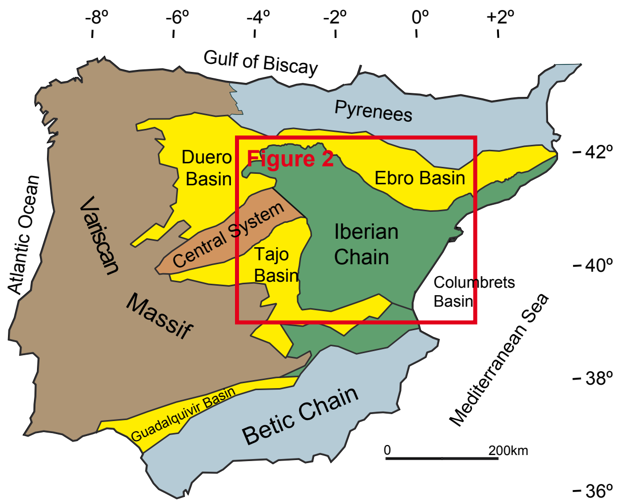

FIGURE 1. Major geological units of the Iberian Peninsula. Location of the study area (Fig. 2) is shown.

Mediterranean, overprinted the Iberian Chain during the late Oligocene and the Neogene (Fontboté, 1954a, b; Julivert *et al.*, 1974; Fontboté *et al.*, 1989, 1990; Roca and Guimerà, 1992). Hence the contractional structures may continue offshore, as do the Mesozoic Iberian basins (Columbrets Basin; Roca, 1996).

The Iberian Chain has a length of 400km along a NW-SE direction (between the Duero Basin and the Mediterranean) and its width is between 125km and 280km in a NE-SW direction (between the Tajo and Ebro basins). It is a fold-and-thrust belt whose structures are mostly NW-SE-oriented, although structures having mainly E-W and NE-SW and N-S orientations (Guimerà and Álvaro, 1990), are also present. Many of these trends are inherited from the Mesozoic rifting events or even the Variscan orogeny (Arche and López-Gómez, 1996; Salas *et al.*, 2001; De Vicente *et al.*, 2009) (Figs. 2; 3). The present relief of the Iberian Chain was generated during the Cenozoic, essentially as product of the contractional structures development.

The aim of this paper is to present a synthesis of the major contractional structure of the Iberian Chain, paying special attention to the geometry of the major-scale thrust system and the generated crustal thickening and relief.

MESOZOIC EVOLUTION

The Paleozoic and Mesozoic geological history of the present-day Iberian Chain generated rock successions and structures which strongly influenced the development of the Cenozoic fold-and-thrust belt. By the end of the Paleozoic, the Variscan orogeny resulted from the convergence and finally the collision of Laurussia and Gondwanaland, resulting in the formation of Pangea at the end of Carboniferous times. The generated orogen was strongly eroded during its development and subsequent early Permian times leading to the formation of the so-called Variscan basement (*e.g.* Matte, 2001). This basement crops out in extensive areas of the western Iberian Peninsula and is also beneath the whole Iberian Chain and the surrounding Cenozoic basins (Ábalos *et al.*, 2002). Here, it is mostly formed by Paleozoic rocks contractively deformed by faults and folds at different scales which experienced a metamorphism of low degree.

Since the late Permian, and during the whole Mesozoic and the early Paleogene, the evolution of the present-day Iberian Chain area was governed by the spread of the Tethys Ocean –located to the E and SE– and by the opening and northward propagation of the southern part of the North Atlantic Ocean since the Late Jurassic (*e.g.* Salas *et al.*, 2001). An extensional context was dominant during this period, the sedimentary basins being bounded by extensional faults and related deformations. As a result, thousands of meters (2000 to 7000m) of rocks were deposited, mainly of continental or shallow marine origin (limestone, shale, sandstone, conglomerate, gypsum and salt).

During the late Permian and the earliest Triassic, the conglomerates, sandstones and shales with the Buntsandstein facies were sedimented in basins bounded by high-angle extensional faults, mostly trending NW-SE but also NE-SW extensional faults (Arche and López-Gómez, 1996). A relatively uniform cover of constant thickness sedimented during the Middle Triassic-Middle Jurassic, which included salt layers in the Triassic Middle Muschelkalk and Keuper facies (future décollements). The Late Jurassic and the Early Cretaceous rocks were sedimented in basins bounded by listric extensional faults mostly striking E-W -but associated with NW-SE faults- (Cameros and Maestrat basins) (Guiraud and Séguret, 1985; Salas and Casas, 1993; Guimerà *et al.*, 1995; Salas *et al.*, 2001) (Fig. 3A).

During the Early and Middle Jurassic and the Late Cretaceous the rifting activity was low and thermal subsidence was dominant (Salas and Casas, 1993). By the end of the Cretaceous, palustrine sediments with marine incursions were deposited (Canérot, 1974; Gautier, 1980; Canérot *et al.*, 1982) showing they were near sea-level, 200 to 300m above the present one. Because of this evolution, the thickness and the lithological features of the Iberian Chain Mesozoic rocks strongly vary within it.

CENOZOIC CONTRACTION

The onset of contraction in the Iberian Chain was by the late Eocene (Díaz Molina and López Martínez, 1979;

González, 1989; González and Guimerà, 1993), while it ended during the early Miocene in the eastern part of the chain and during the late Miocene in its western part (González, 1989; Muñoz and Casas-Sainz, 1997). The resulted crustal thickening led to positive relief that was partially eroded, generating detrital deposits that where deposited in Cenozoic basins surrounding the chain. These deposits were also involved in the contractional deformation.

Influence of Mesozoic structures on the Cenozoic contractional deformation

The orientation and size of the Cenozoic contractional structures were strongly controlled by the inherited Mesozoic extensional structure as well as, in the areas where the Triassic acts as a décollement, by the variations in the Mesozoic succession thickness. Moreover, they were influenced by the older Variscan structures which determined several features of the Mesozoic deformation.

Examples of the influence of Mesozoic structures onto the Cenozoic deformation are:

i) The inversion and internal deformation of the NW-SE Buntsandstein grabens which generated the folds and thrusts in the central Iberian Chain (Aragonese, Castilian and Valencian branches) (Guimerà *et al*., 1995; Arche and López-Gómez, 1996).

ii) The inversion of the E-W and NW-SE extensional faults bounding the Late Jurassic to Early Cretaceous Cameros and Maestrat basins which formed the thrust-belts or individual major thrusts that limit nowadays these

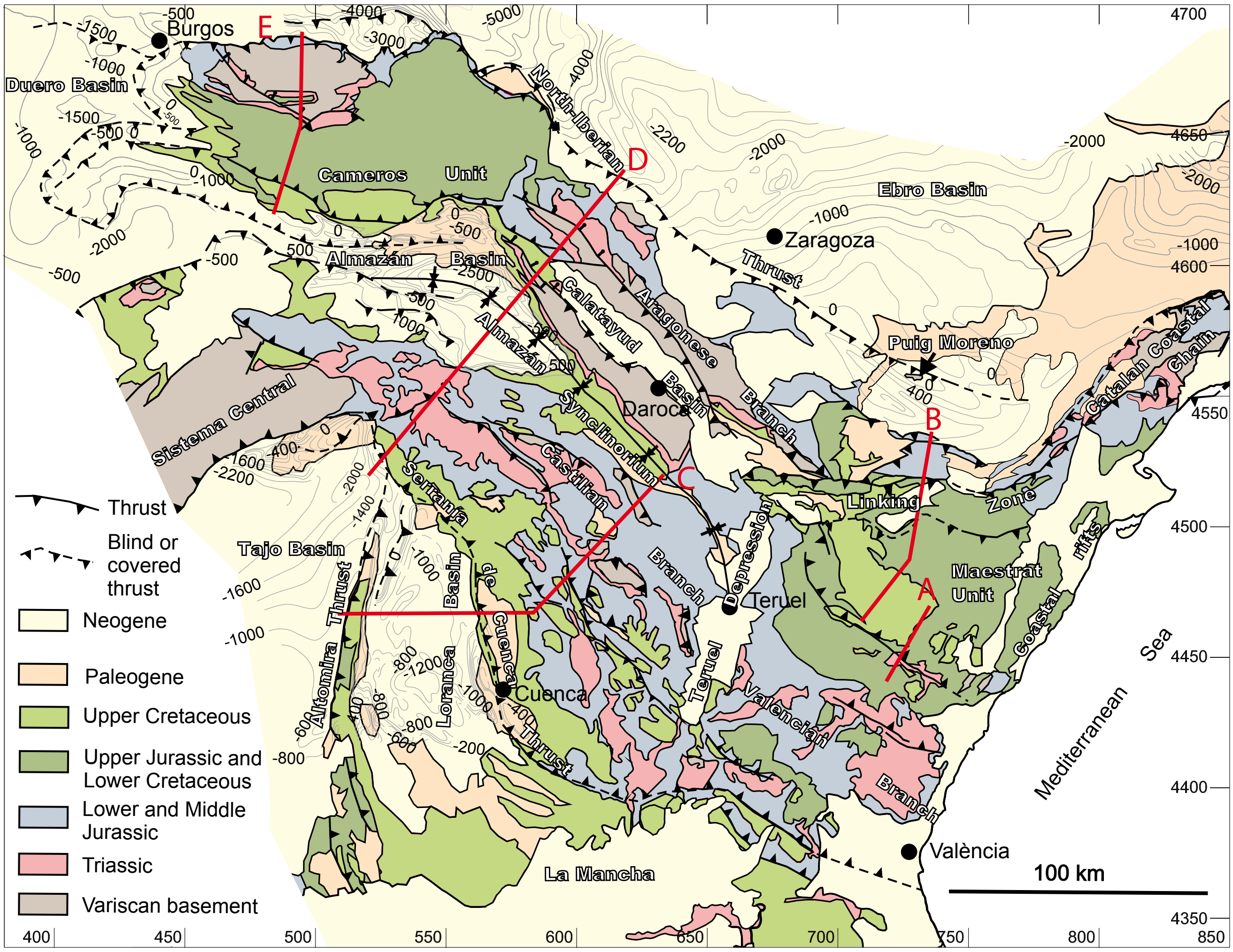

FIGURE 2. Simplified geological map of the Iberian Chain and surrounding Cenozoic basins showing the major tectonic structures and the geological units mentioned in the paper. Contour lines of the carbonatic Upper Cretaceous top (Duero and Almazán), the Cenozoic base (Ebro) and of the base of the Upper Albian Utrillas Formation (Tajo) are shown in the syn-orogenic basins. Contour lines of the Ebro, Duero and Tajo basins are from Instituto Teconológico Geominero de España (1990). Contour lines of the Almazán Basin are from Casas-Sainz *et al.* (2000). Location of geological cross-sections (A to E) of Figure 3 and the UTM coordinates in km (30T, ED50) are also shown.

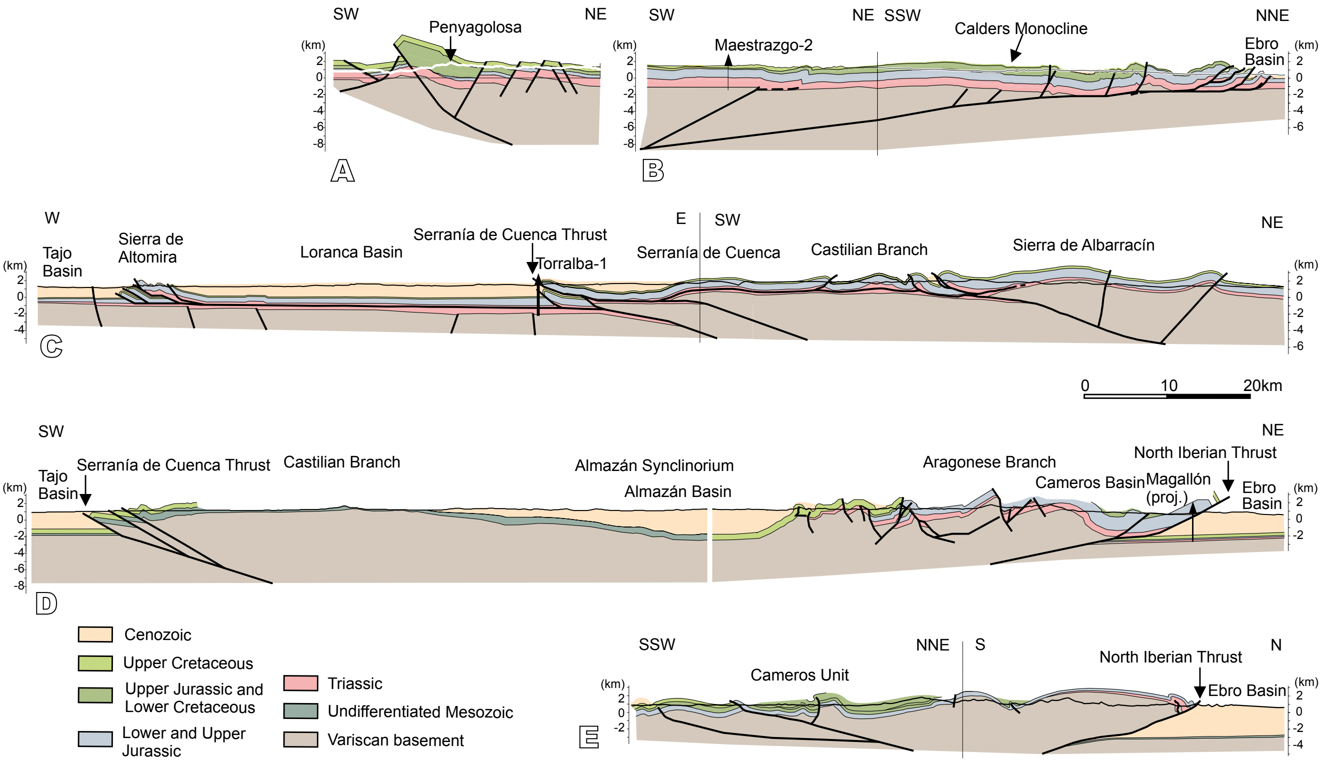

FIGURE 3. Geological sections across the Iberian Chain. B) Modified after Nebot and Guimerà (2016a, b). C) Modified after Guimerà and Álvaro (1990) and Muñoz Martín and de Vicente (1998). D) Modified after Guimerà *et al.* (2004). E) Modified after Guimerà *et al.* (1995). In B) the Variscan basement includes the Permian and Lower Triassic (acoustic basement). For location, see Figure 2.

basins (Casas-Sainz, 1993; Guimerà *et al.*, 1995, 2004; Nebot and Guimerà, 2016a, b).

Considering the vertical variations in the pre-Cenozoic stratigraphic succession, three major lithological units having a different mechanical behavior can be distinguished. The lower, rigid, made up by the Variscan basement and the Lower Triassic (Buntsandstein facies and part or the complete sequence of Muschelkalk carbonates). The intermediate unit behaves as a major décollement that decoupled the contractional deformation, being integrated by the weak lutites and evaporites of the Middle Muschelkalk and Keuper facies, with the carbonatic Upper Muschekalk in between. The evaporitic Middle Muschelkalk unit exists only in the eastern part of the chain (Catalan Coastal Chain and central and eastern Ebro Basin, Maestrat Basin and northern Valencian Branch). Consequently, in the central and western areas, the whole Muschelkalk form a single carbonate level, and there is only a major décollement located in the Keuper lutites and evaporites which allowed the detachment of the overlying sedimentary sequence (upper unit). This one is integrated by the siliciclastic and carbonate rocks of the Jurassic-Cretaceous, and syn-orogenic Cenozoic involved in the contractional structures. These three levels had a strong influence onto the geometry and tectonic style of the Cenozoic contractional structures, especially the tectonically incompetent intermediate level. As a result, thrust-systems having the sole thrust in the intermediate level developed in the upper one.

Major structure of the Iberian Chain

The major part of the Iberian Chain is thick-skinned, as it involves together the Variscan basement and the Mesozoic and Cenozoic cover. Only the more external parts are thin-skinned and the Meso-Cenozoic cover alone, detached from the basement along the Triassic evaporitic levels, is involved in the deformation (Figs. 3; 4).

Most contractional structures in the Iberian Chain strike NW-SE. On a map, two parts can be differentiated in the Iberian Chain (Fig. 2). The major structure of the western and central areas displays two major anticlinoriums separated by the Almazán Synclinorium. The northern anticlinorium, which contains the Cameros Unit and the Aragonese Branch, disappear to the SE, while the southern one contains the Castilian Branch. The Almazán Synclinorium spreads from the Almazán Basin to the NW, to the Teruel Depression in the SE, where it vanishes by a periclinal end. These major structures are not found East of the Teruel Depression. In this area, the E-W-striking Linking Zone –between the Iberian Chain and the Catalan

Coastal Chain, resulting from the inversion of the Lower Cretaceous Maestrat Basin (Salas *et al*., 2001; Nebot and Guimerà, 2016b)– cut the Aragonese Branch structures. South of the Linking Zone, the contractional structures of the Maestrat Unit and the Valencian Branch are NW-SE-trending.

The boundaries of the Iberian Chain and its surrounding foreland basins are always thrusts, although post-tectonic rocks cover these thrusts in some areas. The North-Iberian Thrust (also known as Cameros Thrust), separates the Iberian Chain from the Ebro Basin. It involves the Variscan basement and reaches a horizontal displacement of about 25km and a vertical throw of 4km in the frontal part of the Cameros Unit, where its displacement is maximum and is related to the inversion of the Upper Jurassic-Lower Cretaceous Cameros Basin (Guimerà and Álvaro, 1990; Casas-Sainz, 1993; Guimerà *et al*., 1995; Omodeo Salè *et al*., 2014). Its displacement decreases to the SE along the northeastern boundary of the Aragonese branch and disappear east of Puig Moreno, as can be deduced from the contour lines of the base of the Cenozoic in the Ebro Basin (Fig. 2). To the West, the North-Iberian Thrust is covered by post-tectonic late Neogene rocks, although its continuation can be deduced from the contour lines of the base of the Cenozoic, separating the eastern deformed area from the slightly E-dipping rocks in the whole Duero Basin (Fig. 2).

The Serranía de Cuenca Thrust forms the SE boundary of the basement involved structures in the Iberian Chain (Figs. 2; 3C, D). In the NW it links to the NE-SW thrust bounding the Sistema Central, and to the SE it extends to the Mediterranean coast, displaying an arcuate shape in its central part. In the footwall of the Serrania de Cuenca Thrust, along the central and southeastern parts of the chain, the Mesozoic and Paleogene cover appears deformed by a system of SW- to W-directed thrusts detached in the Triassic Keuper evaporitic facies, containing small piggy-back basins in their hanging-walls. From them, the most

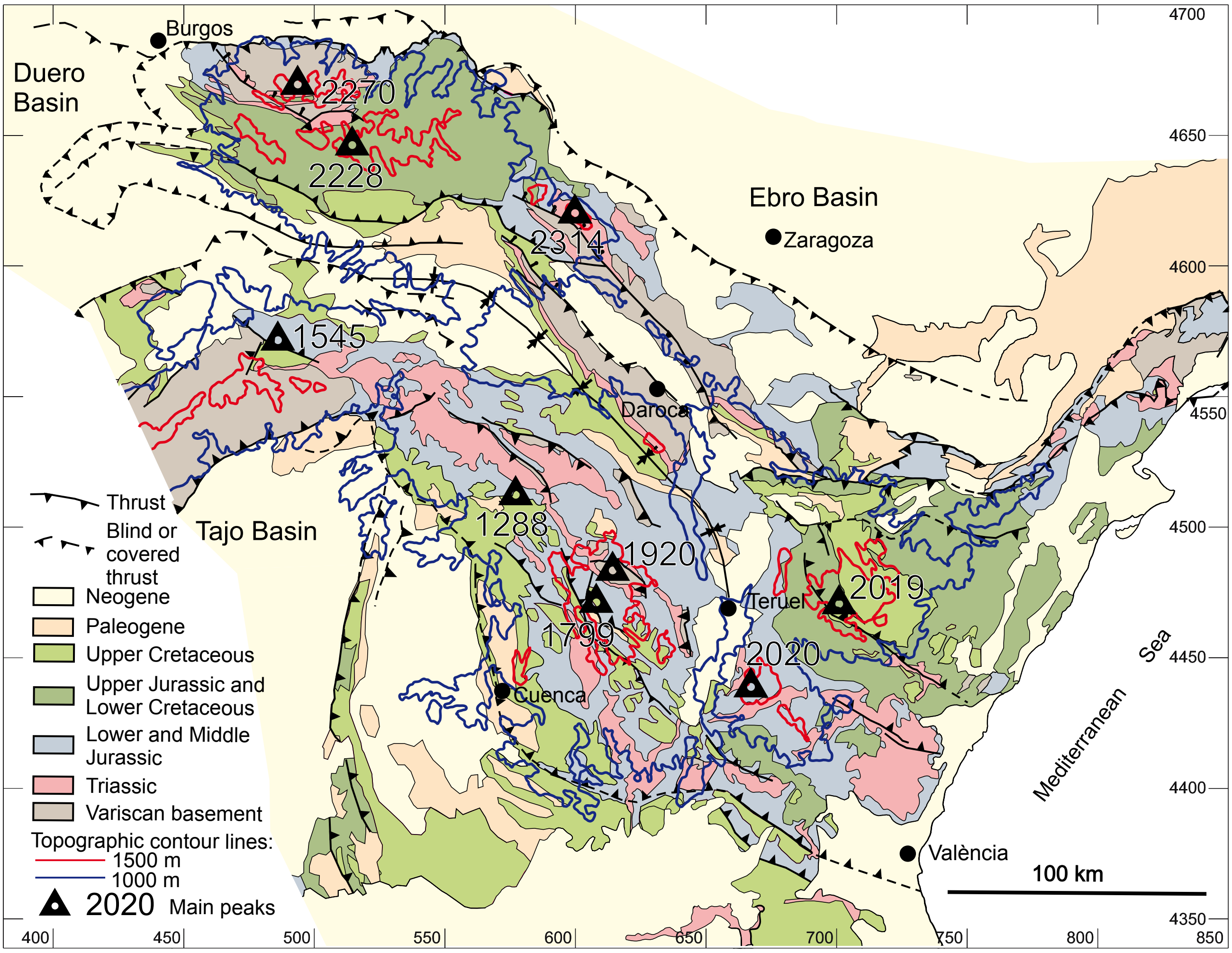

FIGURE 4. Main topographic features of the Iberian Chain superimposed to the simplified geological map.

notable is the N-S-trending Sierra de Altomira Thrust, which constitutes the frontal structure of the Iberian Chain in its central areas, and transported westwards the piggy-back Loranca Basin (Muñoz Martín and de Vicente, 1998; Valcárcel Rodríguez, 2015) (Figs. 2; 3).The displacement of the Serranía de Cuenca Thrust varies laterally, as can be deduced after the shortening which generates in its frontal area plus the structures located in its footwall. In section B of Figure 3 it is estimated in 13.5km (6km in the outcropping area of the thrust plus 7.5km around the Sierra de Altomira Thrust plus the Loranca Basin internal deformation, Muñoz Martín and de Vicente, 1998). To the North this displacement decreases (Muñoz Martín and de Vicente, 1998; Valcárcel Rodríguez, 2015) and in section C (Fig.3) it is estimated to be at least 4km. To the SE, a few thrusts detached at the Keuper facies can be found to the south of the Serranía de Cuenca Thrust.

The Almazán Synclinorium separates the NE and SW anticlinoriums of the Aragonese and Castilian branches (Figs. 2; 3D). It spreads from the large syncline-shaped Cenozoic Almazán Basin up to the Teruel Depression where its SE periclinal end is found. It is cored by Paleogene and Upper Cretaceous rocks. Figure 3D illustrates as the latter were formed by the displacement of the Iberian Thrust sheets over the North Iberian and the Serranía de Cuenca thrusts. The Almazán Synclinorium appears as a non-uplifted area between the Aragonese and Castilian branches, as the base of the Cenozoic (top of the pre-orogenic rocks) is in a similar depth in the Almazán Basin and in the Tajo and Ebro foreland basins (Fig. 3D). The Almazán Basin is so a piggy-back basin within the Iberian Chain, as stated by Guimerà *et al.* (1995) and Casas-Sainz *et al.* (2000).

The Iberian Chain is narrowest across the Cameros Unit, coinciding with a major displacement of the North-Iberian Thrust. As this displacement decreases to the SE, the chain widens, while its internal deformation increases.

The thickness of the crust involved by the thrust system of the Iberian Chain is controversial. Some authors assume that only the upper crust is involved in this thrust system, as was first proposed by Banks and Warburton (1991) and later on by Guimerà and Álvaro (1990), Guimerà *et al.* (1995, 2004), Salas *et al.* (2001). Others authors include all the crust in the thrust system (Salas and Casas, 1993; Anadón and Roca, 1996; Casas-Sainz *et al.*, 2000). In the Linking Zone (Fig. 2), for the Calders Monocline (Fig. 3B) being generated, Nebot and Guimerà (2016b) proposed the existence of a basement thrust with a low-angle ramp of about 9ºS, that reaches a depth of 7.5km below sea level. In the Cameros Unit, the North-Iberian Thrust geometry is well constrained by seismic data (Guimerà and Álvaro, 1990; Casas-Sainz, 1993; Guimerà *et al.*, 1995). It shows a dip of about 18º to the South and ramp geometry in the hanging-wall. Moreover, the kilometric size of the anticlines with Variscan basement in the cores of the Aragonese, Castilian and Valencian branches needs a mid-crustal detachment. This mid-crustal detachment is supported by the magnetotelluric data (Seillé *et al.*, 2015) that shows the presence of conductive bodies at the depth prolongation of the main outcroping thrusts that disapear at a depth of 8–10km.

The regional shortening direction in the Iberian Chain is about NNE-SSW (Guimerà and Álvaro, 1990; Casas-Sainz and Simón-Gómez, 1992; Guimerà, 1994; Guimerà *et al.*, 1995; Nebot and Guimerà, 2016b), parallel to the transport direction of the North-Iberian Thrust and the Linking Zone Thrust belt. This direction is perpendicular to the main orientation of the Pyrenees and parallel to the Cenozoic convergence between Iberia and Eurasia (Roest and Srivastava, 1991; Mouthereau *et al.*, 2014). WNW-ESE-striking thrusts North and South of the Cameros Unit and those of the Linking Zone are dominant dip-slip thrusts, while some NW-SE thrusts experienced a right-lateral component, as the Daroca thrust, South of the Calatayud Basin (Guimerà, 1984; Colomer and Santanach, 1988), or some faults in the north-western part of the Castilian Branch (Rodríguez Pascua, 1993).

Crustal thickening

The gravity anomalies beneath the Iberian Chain indicate that the crust is thickened. The Bouguer anomaly maps (Salas and Casas, 1993; Mezcua and Benarroch, 1996; Guimerà *et al.*, 2016) display negative values for the gravity anomalies in all the chain, reaching a minimum of -110MGal in the Castilian Branch, West of Teruel. Moho depth maps based on topography and Bouguer anomalies were elaborated by Rivero *et al.* (2008) and Guimerà *et al.* (2016) applying a 15km moving average search to entail crustal significance, and crust-mantle density contrast of 300 to 500kg·m^{-3}. According to these models, a minimum Moho depth of -32 to -36km was calculated beneath the Iberian Chain and a maximum Moho depth ranging from -37 to -44km for the central part of the chain (Fig. 4A, B). Close to the Mediterranean coast the Moho depth shows a coastal parallel gradual but rapid shallowing to the East, which is not related to the contractional history of the chain, but to the younger thinning processes linked to the opening of the western Mediterranean (Roca and Guimerà, 1992; Rivero *et al.*, 2008).

Using the Bouguer anomaly, Salas and Casas (1993) and Guimerà *et al.* (2016) performed crustal density-gravity sections obtaining a maximum Moho depth of -43 and -40km, respectively, at the previously mentioned Bouguer minimum. In these sections, the Moho depth decreases towards the Ebro and Tajo basins, reaching values of -33 and -35km, respectively. The Moho depth calculated in the model by Guimerà *et al.* (2016) varies between -33km beneath the Ebro Basin to -40km at the center of the Iberian

Chain (Fig. 5C). These values are intermediate between these of the Topographic Moho (Fig. 5A), and the ones obtained in the Gravimetric Moho (Fig. 5B). Nevertheless, only a major Moho low was computed in the gravity model.

Considering, as stated previously, that the uppermost Cretaceous rocks where just above sea-level (about 200m above the present one) when they were sedimented, while now there are above 1000m in most of the Iberian Chain, and locally above 1500m, the crustal thickening had to be generated during the Cenozoic contraction. The mean topography of the Iberian Chain along the section of the gravity model has been computed (Fig. 3 in Guimerà *et al.*, 2016) applying a 15km moving average search to entail crustal significance. After this calculation, most of the Iberian Chain mean topography is above 1000m, reaching a maximum of 1500m, while the average Moho depth is about -36km (Fig. 6A). A maximum crustal thickness of about 41km and mean crustal thickness of about 37km were then obtained.

It must be taken into account the km-thick Mesozoic succession found in the Iberian Chain, especially in the Late Jurassic-Early Cretaceous basins. Previously to the contractional deformation, the crust (including the Mesozoic) had to be thinner than that in the undeformed Ebro and La Mancha area. Moreover, in the adjoining Late Jurassic-Early Cretaceous Columbrets basin –locally about 10km thick– (Fig. 1), which was only slightly deformed during the Cenozoic, the crust is currently only 12–20km-thick and the extension is mostly concentrated in the middle and lower crust (Roca, 1996; Etheve *et al.*, 2018).

Considering this, a local airy isostatic calculation was performed to estimate the crustal thickness beneath the Iberian Chain at the end of the Mesozoic (Fig. 6B). This was done by comparing the present crustal depth beneath La Mancha area, and its topographic elevation, with the one within the Iberian Basin at the end of the Cretaceous, which was about sea-level at the time, that is, about 200m above the present one. The density values of Figure 5C were used for this calculation. A crustal thickness of 30.3km was obtained for the Iberian Basin at the end of the Cretaceous and a mean crustal thickness of 32.3km for the whole thinned area (Fig. 6B). Comparing the maximum and mean crustal thicknesses before and after the Cenozoic contraction, a maximum crustal thickening of about 9km and a mean crustal thickening of about 5km were obtained.

Amount of horizontal shortening

The horizontal shortening of the Iberian Chain was estimated after cross-sections made from field data (Fig. 3), and after a surface balance of the density-gravity model of Figure 5C (Fig. 6A, B).

Across the Aragonese and Castilian branches transect (Fig. 3D) the horizontal shortening is estimated in 32.5km (16.6km after the internal deformation of the Aragonese

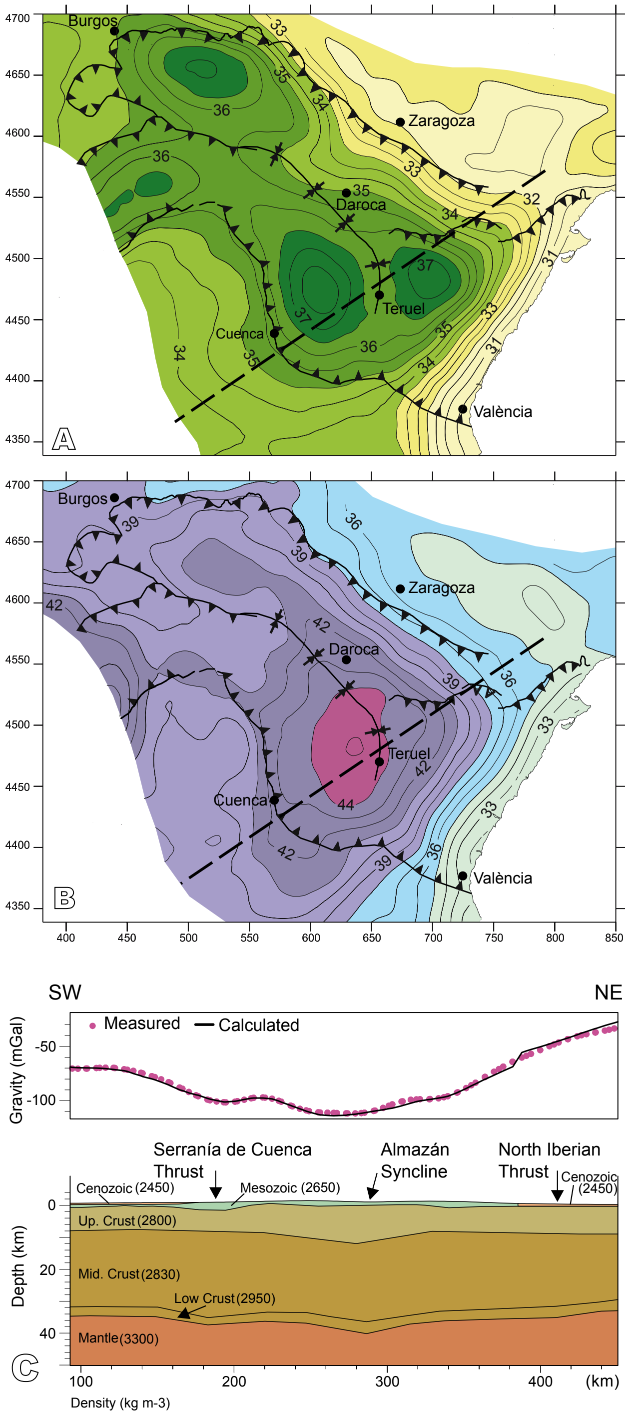

FIGURE 5. Crustal structure of the Iberian chain. A) Airy Moho depth map (in km, contour interval is 0.5km) obtained from the topography and B) Moho depth map (in km, contour interval is 1km) derived from the Bouguer anomaly map. UTM coordinates (30T, ED50) and the major thrusts are shown. Both maps are modified after Guimerà *et al.* (2016). C) Two-dimensional density model across the Iberian Chain, modified after Guimerà *et al.* (2016). For location see A and B.

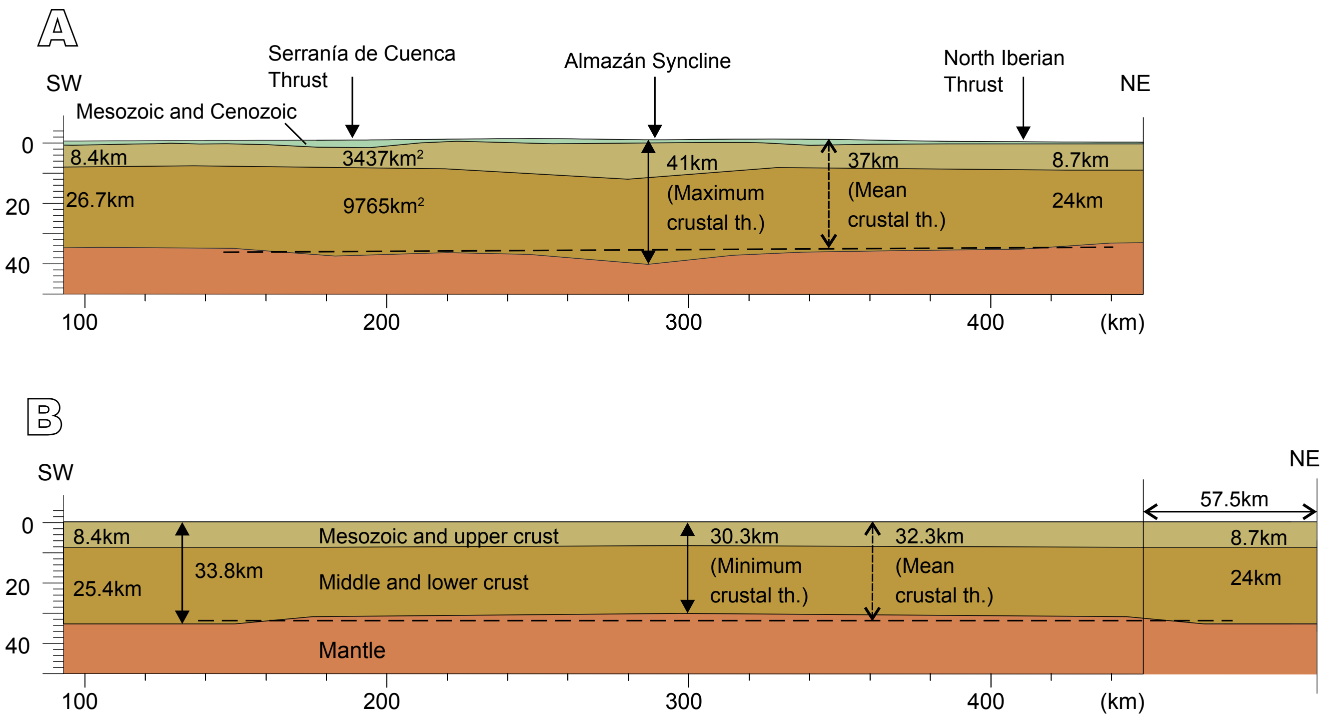

FIGURE 6. Hypothesis on the crustal evolution of the Iberian Chain. A) Density model of Figure 5C at the same vertical and horizontal scale. The width and surface of the Mesozoic plus the upper crust, and the middle-lower crust is shown. B) Crust thickness restored before the Cenozoic contraction. In A and B the dashed line at the base of the crust indicates the mean crustal thickness in the thickened or thinned crust. A Cenozoic crustal horizontal shortening of 57.5km is obtained to preserve the surface of A. The mean topography should be added to the Moho depth to obtain the crustal thickness.

Branch, plus 10km of North Iberian Thrust displacement, Guimerà *et al.*, 2004, plus 6km across the Castilian Branch). A transect across the central part of the chain can be obtained by combining sections A and B (Fig. 3), which gives a horizontal shortening of 31.5km (20km in section B across the Castilian Branch, the Loranca Basin and the Altomira Thrust, and 11.5km after section B). The shortening values along the two transects analyzed are similar, and may be considered a minimum value, as the sections were drawn conservatively and no structures smaller than a hectometric scale were taken into account.

The density-gravity section presented by Guimerà *et al.* (2016) was used to elaborate a surface balance as an approximation to the horizontal shortening across the Iberian Chain (Fig. 6). As discussed previously, the upper crust appears as more thickened beneath the Iberian Chain than the middle and low crust. Beneath the Ebro and Tajo basins it was assumed no tectonic crustal thickening. Making a balance of areas, 57.5km of horizontal shortening is obtained for the whole crust. The middle and lower crust recover an almost constant thickness, as it experienced a Mesozoic thinning bigger than that of the upper crust, while the latter appears as more thickened beneath the Iberian Chain. These two values (32 and 57.5km) may be taken as end values for the Cenozoic horizontal shortening in the Iberian Chain.

Relief of the Iberian Chain

A good match is obtained when comparing the major geological structure of the Iberian Chain with its topography, pointing to this relief as a consequence of the Cenozoic contraction. More than half the surface of the Iberian Chain is above 1000m, while several areas having a surface higher than several hundreds of kilometers are above 1500m. Locally, the relief exceeds 2000m (Fig. 4). The main arguments for the Cenozoic origin of the relief are:

i) By the end of the Late Cretaceous, before the Cenozoic contraction, all the area previously covered by marine sediments became continental palustrine with sporadic marine intercalations (Canérot, 1974; Gautier, 1980; Canérot *et al.*, 1982). This indicates that, at that time, the Iberian Chain domain was about at the sea-level which was 200 to 300m above the present one.

ii) In the North and central parts of the chain, the areas with higher relief are coincident with the two big anticlinoriums previously described. These anticlinoriums are clearly depicted by the Upper Cretaceous rocks, which being the younger rocks before the Cenozoic contraction, are good markers of the overall deformation. The Upper Cretaceous rocks are found progressively at higher elevations through the inner parts of the chain, being over 1800m at the core of the Castilian Branch and West and South of Teruel (Figs. 3; 4).

The erosion of the contractive Iberian Chain building is generally low. This can be deduced after the conservation of Upper Cretaceous rocks in wide areas of the Iberian Chain, especially in the more elevated of the Castilian Branch and the Maestrat Unit. Mesozoic rocks are preserved in most of the chain, even in the more elevated areas above 1500m. Variscan Basement outcrops are found at the core of kilometric-scale anticlines, indicating that they resulted from the erosion of local structures, of dimensions small than the big anticlinoriums (Fig. 4). Two areas have experienced a bigger erosion: the Cameros Unit and its adjacent Aragonese Branch, which constitute the northern big anticlinorium (Fig. 4). The erosion of these areas was synchronous to the Cenozoic contraction, as can be deduced after the syn-orogenic lower Miocene rocks of the Calatayud Basin which are involved by basement thrusts (Julivert, 1954; Colomer and Santanach, 1988). The erosion of the Cameros Unit and the Aragonese Branch resulted from their uplift because of the northeast-directed displacement of the hanging-wall of the North-Iberian Thrust over a foot-wall ramp of this fault (Guimerà *et al*., 1995; Casas-Sainz *et al*., 2000; Guimerà *et al*., 2004) (Figs. 2; 3D, E; 4).

Coevally to the contractional deformation, planation surfaces developed over the Iberian Chain (Solé Sabarís, 1978; Guimerà and González, 1998; Casas-Sainz and Cortés Gracia, 2002; Casas-Sainz and de Vicente, 2009) that were overlain by alluvial sediments at the external parts of the chain, ranging in age from Lower Miocene in the East (González, 1989) to Upper Miocene in the northwest (Muñoz and Casas-Sainz, 1997). These deposits belong to the younger deposits of the foreland basins which at that time were endorheic and so sedimented with a rising base-level, which also helped to reduce the erosion of the chain. The capture of these foreland basins to the Atlantic or the Mediterranean by the end of the Miocene (Riba *et al*., 1983; Serrat, 1992) started the fluvial downcut and erosion of the foreland basins and the Iberian Chain. This downcut is more intense in the northern side of the chain where the Ebro Basin has been strongly eroded. Nevertheless, this downcutting has not yet affected the whole Iberian Chain, and wide areas, especially the more elevated ones, still preserve the late Miocene smooth relief (Guimerà and González, 1998, Guimerà *et al*., 2010).

CONCLUSIONS

The Iberian Chain is a complex intraplate fold-and-thrust belt generated during the late Eocene to Miocene, resulting from the convergence between the Eurasian, Iberian and African plates. The main trend of its Cenozoic contractional structures is NW-SE, but E-W-, NE-SW- and N-S-trending structures are also present, as many of those structures are inherited or influenced by the Mesozoic extensional structures or the Variscan ones.

The boundaries of the chain with its surrounding foreland basins are always thrusts. The North-Iberian Thrust separates the chain from the Ebro Basin to the North, while the Serranía de Cuenca Thrust makes the SE boundary of the chain, separating it from the Tajo Basin and La Mancha foreland area. The area between these main thrusts is basement involved. West and South of the Serranía de Cuenca Thrust, the contractional Iberian Chain building is only made up by Mesozoic and Cenozoic rocks detached in the evaporitic Upper Triassic Keuper facies.

Considering the major structure of the chain, two parts can be differentiated. The western and central areas have two major anticlinoriums (the northern one containing the Cameros Unit and the Aragonese Branch, and the southern one constituted by the Castilian Branch) separated by the Almazán Synclinorium, while these three units are not present West of the Teruel Depression. This eastern area includes the E-W-striking N-verging Linking Zone in the North, and NW-SE-striking S-verging thrusts in the South.

A mean and a maximum crustal thickening of about 5km and 9km, respectively, resulted from the Cenozoic contraction. As the middle and lower crust is likely to have been more thinned during the Mesozoic, the upper crust appears as more thickened than the middle and lower crust, which recovered an almost constant thickness.

The horizontal shortening obtained from the cross-sections is 32km, and from the density-gravity section of 57.5km for the upper crust. These two values may be considered the two end values.

The relief of the Iberian Chain has a strong areal coincidence with the contractional structures and the thickened crust. Hence, they are genetically related, and are a consequence of the Cenozoic contraction. Moreover, planation surfaces developed onto the orogenic building coeval to the contractional structures activity, overlain by the alluvial systems directed to the endorheic foreland basins. Eventually, by the end of the Miocene these basins were captured, and they and the Iberian Chain started the present period of erosion.

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REFERENCES

Ábalos, B., Carreras, J., Druguet, E., Escuder Viruere, J., Gómez Pugnaire, M.T., Lorenzo Álvarez, S., Quesada, C., Rodríguez Fernández, L.R., Gil-Ibarguchi, J.I., 2002. Variscan and Pre-Variscan Tectonics. In: Gibbons, W., Moreno, T. (eds). The Geology of Spain. The Geological Society of London, 155-183.

Álvaro, M., 1995. La Cadena Ibérica. In: Gutiérrez Elorza, M., Meléndez Hevia, A. (eds). XXIX Curso de Geología práctica de Teruel. Universidad de Verano de Teruel, 5-14 julio 1995, Servicio de Publicaciones, Zaragoza, Universidad de Zaragoza, 1-27.

Álvaro, M., Capote, R., Vegas, R., 1979. Un modelo de evolución geotectónica para la Cadena Celtibérica. Acta Geologica Hispanica, 14, 172-177.

Anadón, P., Roca, E., 1996. Geological setting of the Tertiary basins of Northeast Spain. In: Friend, P.F., Dabrio, C.J. (eds). Tertiary Basins of Spain. The Stratigraphic Record of Crustal Kinematics. World and Regional Geology Series, Volume 6. Cambridge, Cambridge University Press, 43-48.

Arche, A., López-Gómez, J., 1996. Origin of the Permian-Triassic Iberian Basin, central-eastern Spain. Tectonophysics, 266(1-4), 443-464.

Banks, C.J., Warburton, J., 1991. Mid-crustal detachment in the Betic system of the southeast Spain.Tectonophysics, 191(3-4), 275-289.

Brinkmann, R., 1960. Aperçu sur les Chaînes Ibériques du Nord de l'Espagne. In: Durand-Delga, M. (ed). Livre à la Mémoire du Professeur Paul Fallot, Mémoires Société Géologique de France, hors série, 1, 291-299.

Canérot, J., 1974. Recherches géologiques aux confins des Chaînes Ibérique et Catalane (Espagne). PhD Thesis. Toulouse, Université Paul-Sabatier, Trabajos de Tesis, Madrid, E.N.A.D.I.M.S.A., serie 5, 4, 517pp.

Canérot, J., Cugny, P., Pardo, G., Salas, R., Villena, J., 1982. Ibérica Central-Maestrazgo. In: García, A. (ed.). El Cretácico de España. Madrid, Universidad Complutense de Madrid, 273-344.

Casas-Sainz, A.-M., 1993. Oblique tectonic inversion and basement thrusting in the Cameros Massif (Northern Spain). Geodinamica Acta, 6(3), 202-216.

Casas-Sainz, A.-M., Simón-Gómez, J.L., 1992. Stress field and thrust kinematics: a model for the tectonic inversion of the Cameros Massif (Spain). Journal of Structural Geology, 14(5), 521-530.

Casas-Sainz, A.-M., Cortés Gracia, A.L., 2002. Cenozoic landscape development within the Central Iberian Chain, Spain. Geomorphology, 44(1-2), 19-46.

Casas-Sainz, A.-M., de Vicente, G., 2009. On the tectonic origin of Iberian topography. Tectonophysics, 474(1-2), 214-235.

Casas-Sainz, A.-M., Cortés Gracia, A.L., Maestro-González, A., 2000. Intraplate deformation and basin formation during the Tertiary within the northern Iberian plate: Origin and evolution of the Almazán Basin. Tectonics, 19(2), 258-289.

Colomer, M., Santanach, P., 1988. Estructura y evolución del borde sur-occidental de la Fosa de Calatayud-Daroca. Geogaceta, 4, 29-31.

De Vicente, G., Vegas, R., Muñoz-Martín, A., Van Wees, J.D., Casas-Sáinz, A., Sopeña, A., Sánchez-Moya, Y., Arche, A., López-Gómez, J., Olaiz, A., Fernández-Lozano, J., 2009. Oblique strain partitioning and transpression on an inverted rift: The Castilian Branch of the Iberian Chain. Tectonophysics, 470(3-4), 224-242.

Díaz Molina, M., López Martínez, N., 1979. El Terciario continental de la Depresión Intermedia (Cuenca). Bioestratigrafía y Paleogeografía. Estudios geológicos, 35, 149-167.

Etheve, N., Mohn, G., Frizon de Lamotte, D., Roca, E., Tugend, J., Gómez-Romeu, J., 2018. Extreme Mesozoic crustal thinning in the eastern Iberia margin: The example of the Columbrets Basin (Valencia Trough). Tectonics, 37(2) DOI: 10.1002/2017TC004613 [In-press version]

Fontboté, J.-M., 1954a. Sobre la evolución tectónica de la depresión del Vallés-Penedés. Sabadell, Arrahona (Publicaciones del Museo de Sabadell), 37pp.

Fontboté, J.-M., 1954b. Las relaciones tectónicas de la depresión del Vallés-Penedés con la Cordillera Prelitoral Catalana y con la depresión del Ebro. In: Tomo extraordinario "Libro Homenaje al Profesor Eduardo Hernández Pacheco. Madrid, Boletín de la Real Sociedad Española de Historia Natural, 52, 281-310.

Fontboté, J.-M., Guimerà, J., Roca, E., Sabat, F., Santanach, P., 1989. Para una interpretación cinemática de la génesis de la cuenca catalano-balear: datos estructurales de sus márgenes emergidos. Libro homenaje a Rafael Soler. Asociación de Geólogos y Geofísicos Españoles del Petróleo, 37-51.

Fontboté, J.-M., Guimerà, J., Roca, E., Sàbat, F., Santanach, P., Fernández-Ortigosa, F., 1990. The Cenozoic Geodynamic Evolution of the Valencia Trough (Western Mediterranean). Revista de la Sociedad Geológica de España, 3(3-4), 249-259.

González, A., 1989. Análisis tectosedimentario del Terciario del borde SE de la Depresión del Ebro (sector bajoaragonés) y cubetas ibéricas marginales. PhD Thesis. Zaragoza, Universidad de Zaragoza, 507pp. [Unpublished]

González, A., Guimerà, J., 1993. Sedimentación sintectónica en una cuenca transportada sobre una lámina de cabalgamiento: la cubeta terciaria de Aliaga. Revista de la Sociedad Geológica de España, 6(1-2), 151-165.

Gautier, F., 1980. Instituto Geológico y Minero de España, Memoria del Mapa geológico de España, 1:50,000, hoja 543

(Villarluengo). Madrid, Servicio de Publicaciones, Ministerio de Industria y Energía, 45pp.

Guimerà, J., 1983. Évolution de la deformation alpine dans le NE de la Chaîne Ibérique et dans la Chaîne Côtière Catalane. Comptes-rendus des séances de l'Académie des sciences. Paris, Série 2, 297(5), 425-430.

Guimerà, J., 1984. Palaeogene evolution of deformation in the northeastern Iberian Peninsula. Geological Magazine, 121(5), 413-420.

Guimerà, J., 1994. Cenozoic evolution of eastern Iberia: Structural data and dynamic model. Acta geològica hispànica, 29(1), 57-66.

Guimerà, J., 2004. Cadenas con cobertera: Las Cadenas Ibérica y Costera Catalana. In: Vera, J.A. (ed.). Geologia de España. Madrid, Sociedad Geológica de España-Instituto Geológico y Minero de España, 602-615.

Guimerà, J., Álvaro, M., 1990. Structure et évolution de la compression alpine dans la Chaîne Ibérique et la Chaîne côtière catalane (Espagne). Bulletin de la Société Géologique de France, 6(2), 339-348.

Guimerà, J., González, A., 1998. El relieve de la Cadena Ibérica como un producto de la compresión alpina. Geogaceta, 24, 163-166.

Guimerà, J., Alonso, A., Mas, R., 1995. Inversion of an extensional-ramp basin by a newly formed thrust: the Cameros basin (N. Spain). In: Buchanan, J.G., Buchanan, P.G. (eds.). Basin Inversion. The Geological Society of London, 88 (Special Publications), 433-453.

Guimerà, J., Mas, R., Alonso, A., 2004. Intraplate deformation in the NW Iberian Chain: Mesozoic extension and Tertiary contractional inversion. Journal of the Geological Society of London, 161(2), 291-303.

Guimerà, J., González, A., Salas, R., Sancho, C., 2010. Origin and preservation of the late contractional relief of an intraplate thrust-belt: the NE Iberian Chain (Iberian Peninsula). Vienna, 2-7 May 2010, EGU General Assembly, pg. 5135.

Guimerà, J. Rivero, L., Salas, R., Casas, A., 2016. Moho depth inferred from gravity and topography in an intraplate area (Iberian Chain). Tectonophysics, 666, 134-143.

Guiraud, M., Séguret, M., 1985. A releasing solitary overstep model for the Late Jurassic-Early Cretaceous (Wealdian) Soria strike-slip Basin (Northern Spain). In: Biddle, K.T., Christie-Blick, N. (eds.). Strike Slip Deformation Basin Formation and Sedimentation. The Society of Economic Paleontologists and Mineralogists,Special Publication, 37, 159-175.

Hahne, C., 1930a. Das Keltiberische Gebirgsland östlich der Linie Cuenca-Teruel-Alfambra. Beiträge zur Geologie der westlichen Mediterraneangebiete, 5 (Translated into Spanish in 1943). Publicaciones alemanas sobre Geología de España, 2, 7-50.

Hahne, C., 1930b. Stratigraphische und tectonische Untersuchungen in der Provinzen Teruel, Castellón und Tarragona (Spanien). Zeitschriftder Deutschen Geologischen Gesellschaft, 82, 79-112. [Translation to Spanish, 1943. Publicaciones alemanas sobre Geología de España, 2, 51-98]

Instituto Tecnológico Geominero de España, 1990. Documentos sobre la Geología del subsuelo de España. Madrid, Servicio de Publicaciones del Ministerior de Industria y Energía, 13vols.

Julivert, M., 1954.Observaciones sobre la tectónica de la depresión de Calatayud. Sabadell, Arrahona (Publicaciones del Museo de Sabadell), 17pp.

Julivert, M., Fontboté, J.M., Ribeiro, A., Conde, L., 1974. Mapa tectónico de la Península lbérica yBaleares. 1:1,000,000. Madrid, Instituto Geológico y Minero de España, 113pp.

Matte, P., 2001. The Variscan collage and orogeny (480-290 Ma) and the tectonic definition of the Armorica microplate: a review. Terra Nova, 13(2), 122-128.

Mezcua, J., Gil, A., Benarroch, R., 1996. Estudio Gravimétrico de la Península Ibérica y Baleares. Madrid, Instituto Geográfico Nacional, Ministerio de Fomento, 1-7.

Mouthereau, F., Filleaudeau, P.-Y., Vacherat, A., Pik, R., Lacombe, O., Fellin, M.G., Castelltort, S., Christophoul, F., Masini, E., 2014. Placing limits to shortening evolution in the Pyrenees: Role of margin architecture and implications for the Iberia/Europe convergence. Tectonics, 33(12), 2283-2314.

Muñoz Martín, A., de Vicente, G., 1998. Cuantificación del acortamiento alpino y estructura profunda del extremo sur-occidental de la Cordillera Ibérica (Sierras de Altomira y Bascuñana). Revista de la Sociedad Geológica de España, 11(3-4), 233-252.

Muñoz, A., Casas-Sainz, A.-M., 1997. The Rioja Trough (N Spain): tectosedimentary evolution of a symmetric foreland basin. Basin Research, 9(1), 65-85.

Nebot, M., Guimerà, J., 2016a. Structure of an inverted basin from subsurface and field data: the Late Jurassic-Early Cretaceous Maestrat Basin (Iberian Chain). Geologica Acta, 14(2), 155-77.

Nebot, M., Guimerà, J., 2016b. Kinematic evolution of a fold-and-thrust belt developed during basin inversion: The Mesozoic Maestrat basin, E Iberian Chain. Geological Magazine, 155(3), 630-640

Omodeo Salè, S., Guimerà, J., Mas, R., Arribas, J., 2014. Tectono-stratigraphic evolution of an inverted extensional basin: the Cameros Basin (north of Spain). International Journal of Earth Sciences, 103(6), 1597-1620.

Roca, E., 1996. La cubeta mesozoica de las Columbrets: aportaciones al conocimiento de la estructura del surco de Valencia. Geogaceta, 20(7), 1711-1714.

Roca, E., Guimerà, J., 1992. The Neogene structure of the eastern Iberian Margin: structural constraints on the crustal evolution of the Valencia Trough (WesternMediterranean). In: Banda, E., Santanach, P. (eds.). Geology and Geophysics of the Valencia Trough, Western Mediterranean. Tectonophysics, 203(1-4), 203-218.

Rodríguez Pascua, M.A., 1993. Cinemática y dinámica de las deformaciones alpinas en la zona del Alto Tajo (Guadalajara). Degree Thesis. Madrid, Universidad Complutense de Madrid, 118pp.

Riba, O., Reguant, S., Villena, J., 1983. Ensayo de síntesis estratigráfica y evolutiva de la cuenca terciaria del Ebro. In: Comba, J. (ed.) Libro Jubilar J.M. Ríos, Geología de España 2. Madrid, Instituto Geológico y Minero de España, 131-159.

Rivero, L., Guimerà, J., Salas, R., Casas, A., 2008. Constraints on the Moho depth in the Iberian Chain. Geo-Temas, 10, 673-675.

Roest, W.R., Srivastava, S.P., 1991. Kinematics of the plate boundaries between Eurasia, Iberia, and Africa in the North Atlantic from the Late Cretaceous to the present. Geology, 19(6), 613-616.

Salas, R., Casas, A., 1993. Mesozoic extensional tectonics, stratigraphy and crustal evolution during the Alpine cycle of the eastern Iberian basin. Tectonophysics, 228(1-2), 33-55.

Salas, R., Guimerà, J., Mas, R., Martín-Closas, C., Meléndez, A., Alonso, A., 2001. Evolution of the Mesozoic Central Iberian Rift System and its Cainozoic inversion (Iberian Chain). In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (eds.). Peri-Tethys Memoir, 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins. Mémoires du Muséum national d'Histoire naturelle, 186, 145-186.

Seillé, H., Salas, R., Pous, J., Guimerà, J., Gallart, J., Torné, M., Romero-Ruiz, I., Diaz, J., Ruiz, M., Carbonell, R., Mas, R., 2015. Crustal structure of an intraplate thrust belt: The Iberian Chain revealed by wide-angle seismic, magnetotelluric soundings and gravity data. Tectonophysics, 663, 339-353.

Serrat, D., 1992. La xarxa fluvial dels Països Catalans. In: Guimerà, J., Serrat, D. (eds.). Història Natural dels Països Catalans. Geologia II, Barcelona, Fundació Enciclopèdia Catalana, 376-389.

Solé Sabarís, L., 1978. La Meseta. In: De Terán, M., Solé Sabarís, L., Bolos, M., Bosque Maurel, J., Cabo, A., Capel, H., López Gómez, A., Roselló, V., Vila Valentí, J., Villegas Molina, F. (ed.). Geografía Generalde España. Barcelona, Editorial Ariel, 42-62.

Valcárcel Rodríguez, M.M., 2015. Reconstrucción y restitución 3D del anticlinal de Puerta-Pareja y estudio de su evolución deformacional y cinemática en su contexto regional (cuenca de Loranca y Sierra de Altomira). PhD memoir, Universitat de Barcelona, 270pp. [Available at: https://www.tdx.cat/handle/10803/336373]

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