Tectonophysics xxx (2009) xxx–xxx

Tectonophysics

Large-scale distributed deformation controlled topography along the western Africa–Eurasia limit: Tectonic constrains

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article info

Article history:
Received 22 January 2008
Received in revised form 10 November 2008
Accepted 25 November 2008
Available online xxxx

Keywords:
Iberian Peninsula
Cenozoic morphotectonics
Strike–slip faulting
Lithospheric folding

abstract

In the interior of the Iberian Peninsula, the main geomorphic features, mountain ranges and basins, seems to be arranged in several directions whose origin can be related to the N–S plate convergence which occurred along the Cantabro–Pyrenean border during the Eocene–Lower Miocene time span. The Iberian Variscan basement accommodated part of this plate convergence in three E–W trending crustal folds as well as in the reactivation of two left-lateral NNE–SSW strike-slip belts. The rest of the convergence was assumed through the inversion of the Iberian Mesozoic Rift to form the Iberian Chain. This inversion gave rise to a process of oblique crustal shortening involving the development of two right lateral NW–SE shear zones. Crustal folds, strike-slip corridors and one inverted rift compose a tectonic mechanism of pure shear in which the shortening is solved vertically by the development of mountain ranges and related sedimentary basins. This model can be expanded to NW Africa, up to the Atlasic System, where N–S plate convergence seems also to be accommodated in several basement uplifts, Anti-Atlas and Meseta, and through the inversion of two Mesozoic rifts, High and Middle Atlas. In this tectonic situation, the microcontinent Iberia used to be firmly attached to Africa during most part of the Tertiary, in such a way that N–S compressive stresses could be transmitted from the collision of the Pyrenean boundary. This tectonic scenario implies that most part of the Tertiary Eurasia–Africa convergence was not accommodated along the Iberia–Africa interface, but in the Pyrenean plateboundary. A broad zone of distributed deformation resulted from the transmission of compressive stresses from the collision at the Pyrenean border. This distributed, intraplate deformation, can be easily related to the topographic pattern of the Africa–Eurasia interface at the longitude of the Iberian Peninsula.

Shortening in the Rif–Betics external zones – and their related topographic features – must be conversely related to more “local” driven mechanisms, the westward displacement of the “exotic” Alboran domain, other than N–S convergence. The remaining NNW–SSE to NW–SE, latest Miocene up to Present convergence is also being accommodated in this zone straddling Iberia and Morocco, at the same time as a new ill-defined plate boundary that is being developed between Europe and Africa.

1. Introduction

The gross topographic features in the interior of the Iberian Peninsula seem to be distributed following a regular pattern. Main basement uplifts on the Variscan area, the Iberian Massif, are arranged in a roughly E–W direction giving rise to elongated sierras parallel to the Iberian Peninsula northern border mountains, the Pyrenean and Cantabrian ranges. In addition to these basement uplifts, a swathe of elongated and narrow basement-built ranges runs in a NNE–SSW direction connecting the northern marginal mountains with the basement uplift located in the central part of the Peninsula. A low-deformed Alpine chain to the east completes this geomorphic frame in an almost conjugate NW–SE direction (Fig. 1).

A coherent plate-tectonics model has not so far been put forward for this intraplate deformation of the Iberian Peninsula. Conventionally, most of the uplifted areas of the plate interior have been vaguely related to the influence of the so-called Betic orogeny, while the building of the “intermediate Alpine chain”, the Iberian Chain, has been considered to be a consequence of the Pyrenean orogeny The mixed effect of both orogenies has led to complex evolutionary models for the different morphostructural units forming the present landscape of the Iberian Peninsula, rather complicated in plate-tectonics terms. The aim of this work is to propose such a coherent plate-tectonics model as well as to illustrate the mode of intraplate distributed deformation by means of a description of the main morphotectonic units of the interior of the Iberian Peninsula.

There are several incontestable facts which can constrain the topographic evolution in the interior of Iberia. Among these, and perhaps the most important, is the protracted, first N–S and then NNW–SSE, Africa–Eurasia convergence at the longitude of Iberia since the Late Cretaceous (Dewey et al., 1989; Mazzoli and Helman, 1994). This compressive regime led eventually to collision in the Iberia–Europe interface giving rise to the building of the Pyrenees in the

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doi: 10.1016/j.tecto.2008.11.026

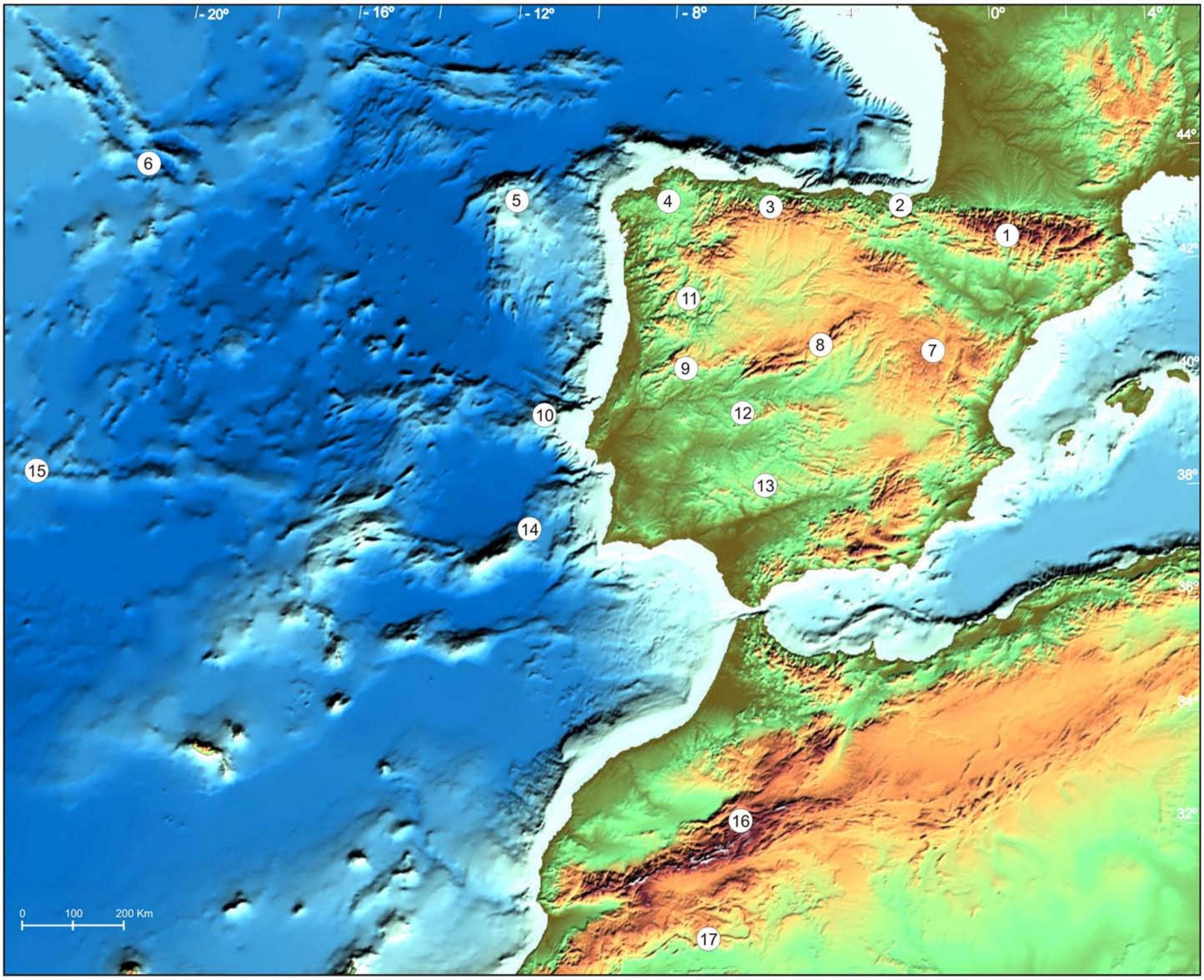

Fig. 1. Main topographic features on-offshore Iberia–NW Africa zone. 1) Pyrenees. 2) Basque Mountains. 3) Cantabrian Mountains. 4) Galician Massif. 5) Galicia Bank. 6) King's Through. 7) Iberian Chain. 8) Spanish Central System. 9) Portuguese Central System. 10) Extremadura Spur. 11) Vilariça Fault System. 12) Toledo Mountains. 13) Sierra Morena. 14) Gorringe Bank. 15) Gloria Fault. 16) Atlas. 17) Anti-Atlas.

Paleogene. Evidence for the transmission of some deformation from the collision border to the interior of the Southern (lower) plate (Iberia) can be found in the form of the Oligocene–Lower Miocene infill of sedimentary basins (Calvo et al., 1993; Calvo, 2004; De Vicente et al., 2007a) and during Oligocene exhumation and rock uplifting that have been deduced from apatite fission-track analysis in segments of some intraplate ranges (De Bruijne, 2001; De Bruijne and Andriessen, 2002; Barbero et al., 2005). Another important fact is the paleomagnetic evidence that precludes a significant N–S relative motion between Africa and Iberia since Jurassic times. Paleopoles for two Jurassic (200 Ma) dykes, Messejana–Plasencia in the south western-central Iberian Peninsula and Foum-Zguid in the Moroccan Anti-Atlas, yield the same – indistinguishable in palaeomagnetic terms – coordinates (Palencia et al., 2003; Palencia, 2004).Hence only substantial longitudinal (N–S) motion can have occurred between these two dykes after their emplacement prior to the opening of the Central Atlantic between Africa and America. This evidence supports the idea that Iberia should have been firmly moored to Africa during the Pyrenean collision, thus facilitating sufficient Iberia–Africa mechanical coupling for intraplate deformation (Vegas et al., 2005).

This intraplate deformation has been in part described as lithospheric folds in the Iberian Peninsula (Cloething et al., 2002) and the Moroccan Anti-Atlas (Teixell et al., 2003). These large scale folds correspond to some kind of crustal buckling induced by transmitted stresses from the Pyrenean plate-boundary over a broad area. In addition to these folds, an intraplate, left-handed NNE–SSW strike-slip corridor, assumes part of the deformation transferring crustal shortening from the plateboundary to the interior of Iberia (Vegas et al., 2004; De Vicente et al., 2005).

The intraplate Iberian framework is completed by a process of oblique inversion in the Eastern Mesozoic Rift, which can be considered as a conjugate NW–SE corridor accommodating mainly transpressive shortening. All these processes of intraplate deformation (crustal buckling, left-lateral simple shear and right-lateral oblique rift inversion) provide an overall tectonic setting for the main geomorphic units of the Iberian plate interior (Fig. 1). In fact, a pure-shear model of shortening can be envisaged for the origin of the mountain ranges and basins of the Iberian interior (Vegas, 2006). It is worth mentioning that most of this process of crustal shortening, and related morphogenesis, must be related to the N–S, Oligocene–Upper Miocene Africa–Europe

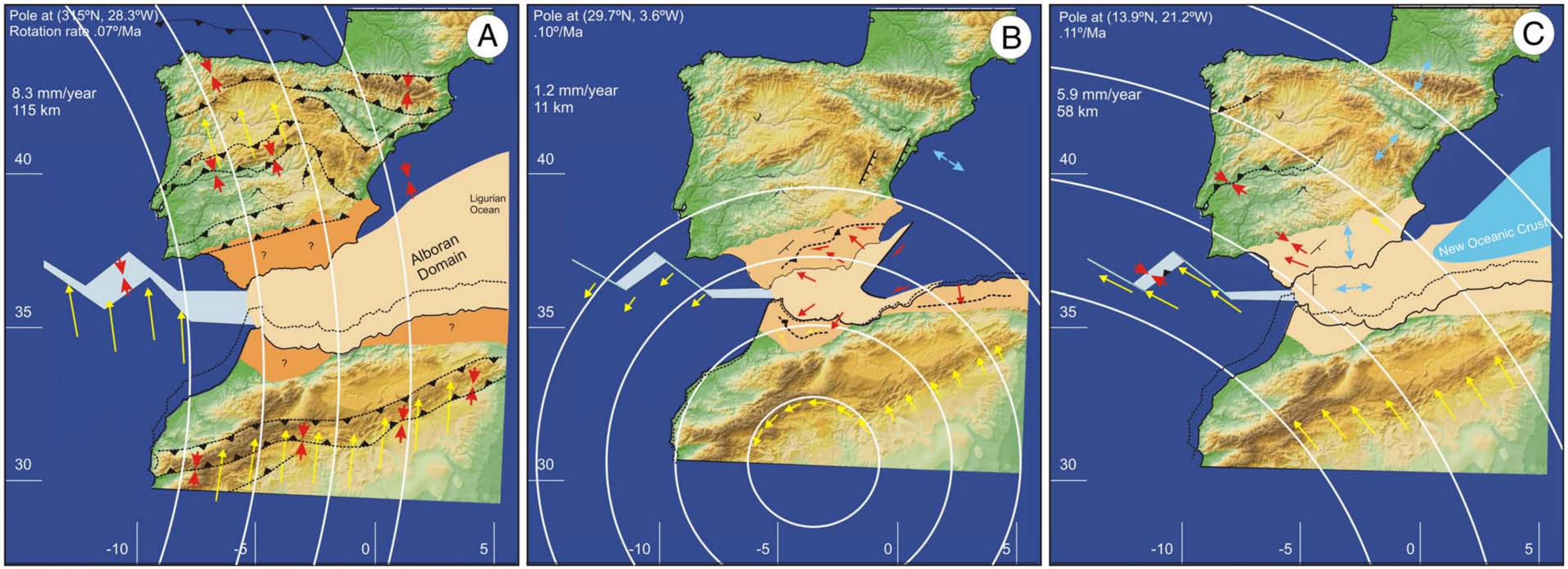

Fig. 2. Paleopoles of Iberia–Africa movement during the Cenozoic. Modified from Rosenbaum et al. (2002). A) Relative Africa–Iberia movement between An 13 An 6: Iberia moored to Africa. B) Relative movement between Africa–Iberia An 6 An 5: Iberia–Africa decoupling. C) Relative movement between Africa–Iberia An 5 Present day. Iberia is a part of Eurasia. Red arrows: Shortening. Blue arrows: Extension. Yellow arrows: mean displacements.

convergence. The deformation corresponding to the NNW–SSE to NW–SE Tortonian-up to Present plate convergence may represent a weak modification of the Iberian landscape superimposed on the previous main geomorphic pattern, specially in the easternmost part of Iberia and perhaps with the exception of the Atlantic border where Neogene to Present tectonic activity seems to be more important as witnessed by seismic activity (De Vicente et al., 2008).

2. Tectonic setting

The presence of maxima in the spectra of topography and of Bouguer Anomaly in profiles which cross Iberia and NW Africa, as well as the age of the mountain building, confirm the idea that this distributed deformation was produced as a consequence of the mechanical coupling of Africa and Iberia during a substantial part of the Cenozoic. These large scale structures have been interpreted as lithosphere folds (Cloetingh et al., 2002), with wavelengths between 400 and 600 km, and affected the lithosphere with continental and oceanic crust (De Vicente et al., 2007c). Thus, the structural heights off-shore W Iberia (Galicia Bank, Extremadura Spur and Gorringe Bank) can be considered as prolongations of the Pyrenees, the Central System and Sierra Morena (?).

This coupling can also be deduced from the paleo-reconstructions of the movement of Africa with respect to Iberia during the Tertiary (Rosenbaum et al., 2002). Between the anomalies An 13 and An 6 (Early Oligocene–Lower Miocene), Africa and Iberia came 115 km closer, which implies a N–S approximate shortening of 3–4% with a rate lower than one centimetre per year (Fig. 2a). This rate of shortening is perfectly compatible with the 10% to 20% shortening which can be measured in Central Spain for the Neogene (De Vicente et al., 1996).

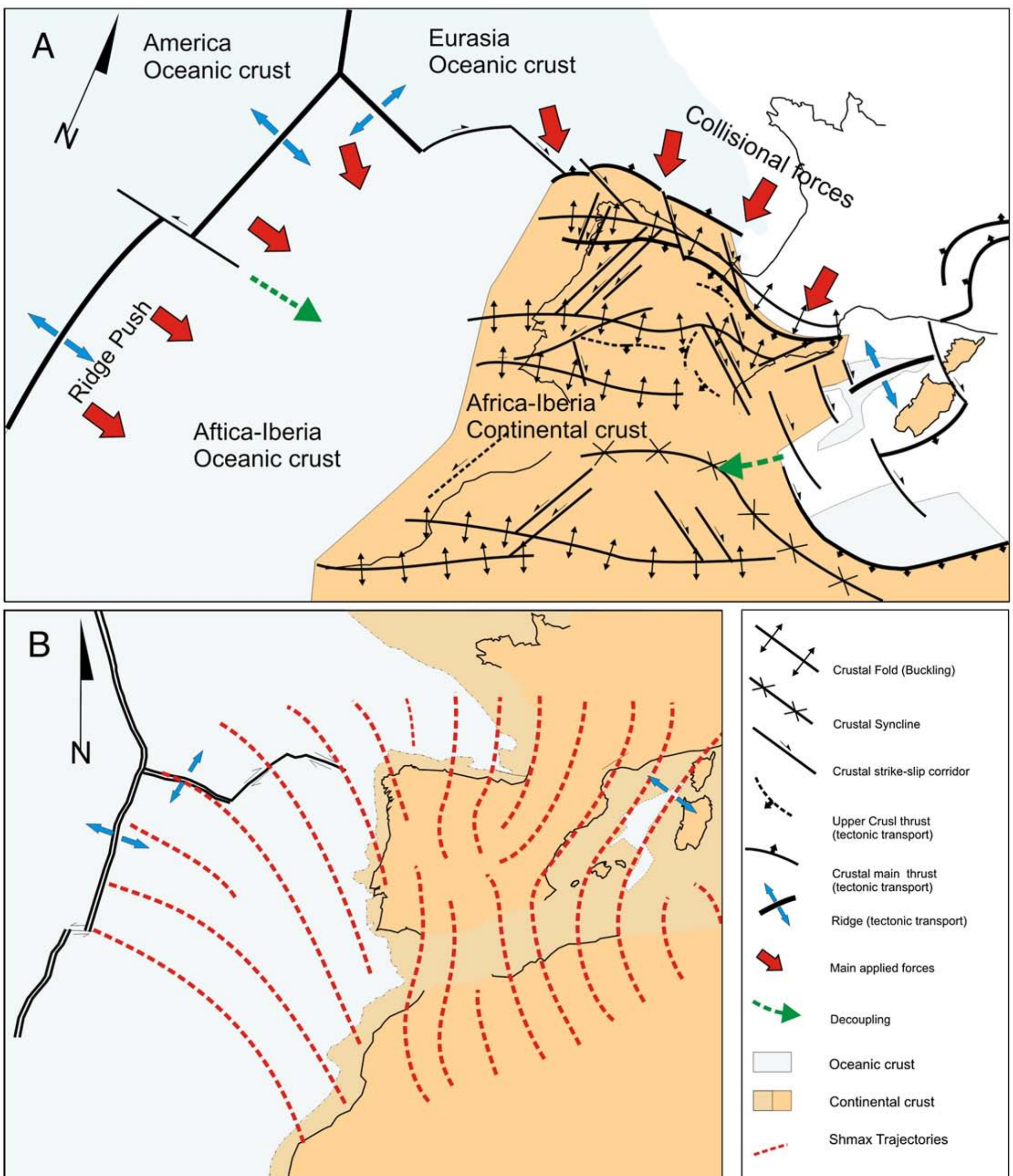

Fig. 3. A) Paleoplates sketch for Lithospheric folding development. B) Paleostresses reconstruction, we take S_{hmin} at the Ridge as ridge push to join with S_{hmax}. (30–20 Ma) Oligocene–Lower Miocene.

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Recent paleomagnetic studies (Palencia et al., 2003; Palencia, 2004) indicate that no important relative motion, in a paleomagnetic sense, has occurred at the boundary between Africa and the Iberian Peninsula since the Eocene. During this time, most of the convergence between both plates was resolved through deformations which affected the upper crust and the lithosphere, giving rise to higher order folds and the basic topographical pattern, both on and off shore (Figs. 1 and 2). Thereafter, boundary stresses responsible for the Cenozoic deformation of Iberia (and northwestern Africa) from Eocene up to Middle–Upper Miocene, must be referred to the Iberia–Europe interface, the Cantabrian–Pyrenean border. In this context, paleostresses with a NNW directed S_{hmax} (σ_y) can be envisaged at the western sector of the Pyrenean edge, until King's Trough, where the S_{hmax} was practically E–W, in accordance with the push of the Mid-Atlantic Ridge. It must be stressed that this situation is similar to the one which exists today between the Terceira Ridge and the Gloria Fault. Contrary to this, towards the E, the S_{hmax} direction would turn towards the NNE at the eastern end of the Pyrenees and to the NE in the Valencia Trough. The resulting geometry is similar to an indenter with its corners at the King's and Valencia Troughs, where extension occurred. This theoretical reconstruction of paleostresses draws a neutral point (σ_y=σ_x) in the interior of the Peninsula which must be interpreted as a certain tendency to triaxial compression paleostresses (De Vicente et al., 2005, 2007c) (Fig. 3a, b).

By analysing the macro-structural characteristics of basins, chains and resulting fault corridors in the Iberian interior, the most surprising factor is the presence of important changes in the direction of tectonic transport in structures simultaneously active during the Oligocene–Lower Miocene. The main vergences are centripetal towards the Duero and, especially, the Madrid Cenozoic basins (Fig. 4).

Data on Cenozoic paleostresses are relatively abundant in the Iberian Peninsula, especially in the Iberian Chain. The statistical analyses of the occurrence of tensors, fundamentally differentiated by the S_{hmax} orientation (Liesa and Simón-Gómez, 2007), indicate the presence of 4 maxima (N–S, NE–SW, NW–SW and E–W) which are related to Cenozoic *tectonic phases* with no large bending in their paleo-trajectories. Considering the data on paleostresses in other chains in the Iberian microcontinent (Andeweg et al., 1999; Escuder-Viruete et al., 2001; Antón, 2003), the ubiquitous presence of the Cenozoic paleostresses with N–S S_{hmax} is evident. It is also interesting that there is always a tensor close to axial compression (σ_1>σ_2=σ_3 σ_1=σ_y), perpendicular to the chain-basin thrust nearest to the measurement station. Taking these matters into account and considering the apparent constrictive conditions of the deformation, it is logical to think that a neutral point (constrictive deformation) was produced in the interior of the peninsula (triaxial to radial compression) in the distribution of the paleostresses during the Oligocene–Lower Miocene (De Vicente et al., 2005). In these circumstances almost every discontinuity becomes potentially active, so the inherited first order faults (discontinuities affecting the whole upper crust) can be reactivated and nucleate bending deformation and local stresses, not necessary related with the regional compressive stress field (De Vicente et. al, 2007b,c) (Figs. 3 and 4). However, it should be pointed out that, even though axial compression paleostresses are abundant, strike-slip type faulting predominates and triaxial compression paleostresses are scarce (Liesa and Simón-Gómez, 2007), Therefore, it must be assumed (and explained) a certain deformation partitioning from

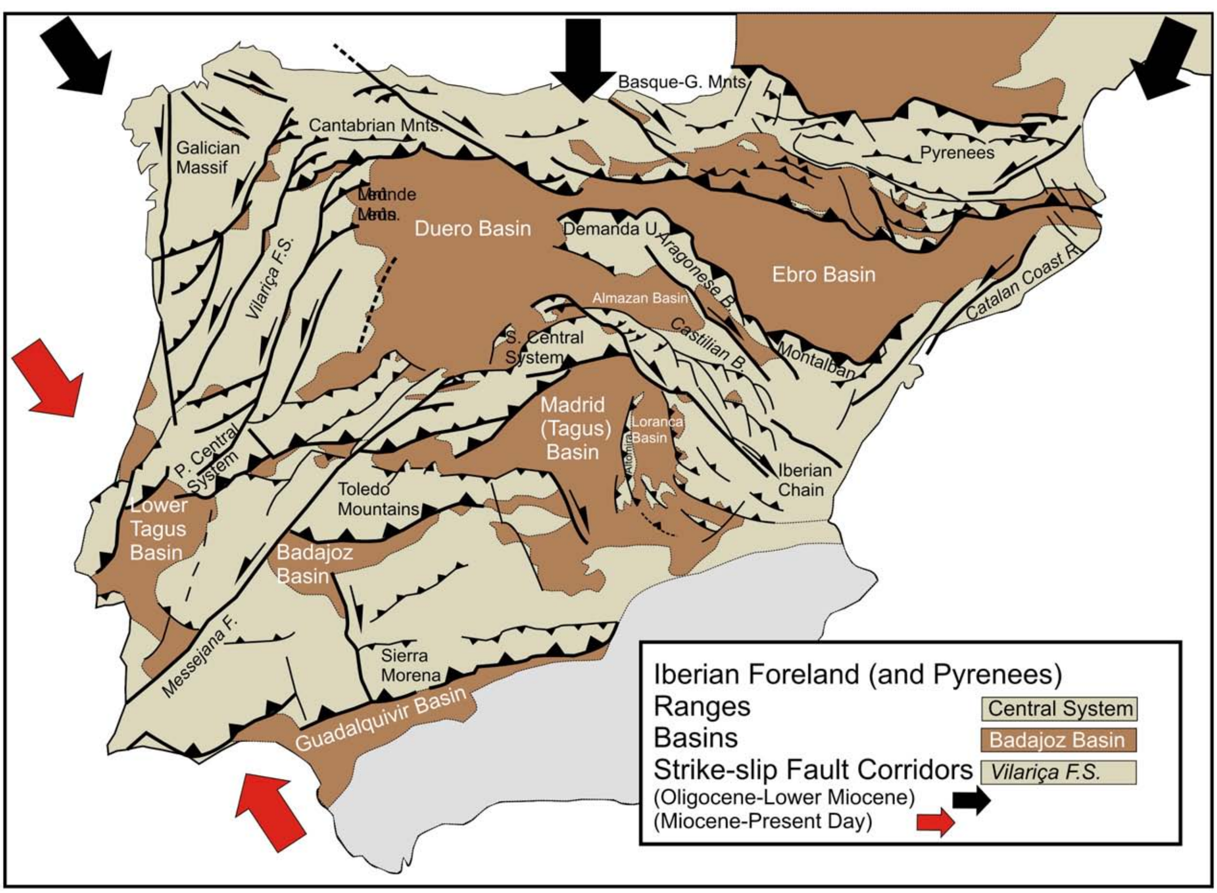

Fig. 4. The main tectonic units (basins, ranges, crustal buckling, and strike–slip fault corridors) of the foreland and Pyrenees onshore Iberia. Black arrows: shortening directions during the Oligocene–Lower Miocene. Red arrows: shortening direction during the Miocene–Present day (not Pyrenean related).

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general constrictive conditions to local conditions of strike-slip type, especially in the Iberian Chain.

This structural pattern is lost between the C6 and C5 anomalies (Lower Miocene–Upper Miocene), where the paleomagnetism indicate that the movement between Africa and Iberia was right lateral and parallel to the plate boundary, although at a very low velocity (1.2 mm per year) (Rosenbaum et al., 2002) (Fig. 2b). From a tectonic point of view, this sharp change in plate kinematics can be interpreted as the ceasing of the constrictive conditions of the deformation and the end of the active lithosphere folding on a board scale. The coincidence in time with the displacement of the Alboran Block towards the W implies that this emplacement mechanically decoupled Iberia from Africa.

The current kinematics between Iberia and Africa involves a convergence in the NW–SE direction, which is shown in the state and the orientation of the active stresses (De Vicente et al., 2008) (Fig. 2c). This new stress field is superimposed on the structures generated up to the Lower Miocene and does not have its origin in the northern edge, where mainly normal type focal mechanisms are registered (with moment tensor focal mechanisms). A new Africa–Europe plateboundary was created to the south of the Iberian Peninsula and its formation is still ongoing.

3. Intraplate deformation prompted by the Iberia (Africa)–Europe plate boundary

The northern edge of the Iberian Peninsula shows a rather continuous mountain belt from the Mediterranean Sea to the Atlantic margin (Fig. 1), creating a physical barrier that separates the Iberian interior from French territory and the narrow Cantabrian margin. The eastern, intra-continental segment of this belt corresponds to the *Pyrenees*, a well known collision orogeny along which a north-verging zone (*North Pyrenean*) and a south-verging one (comprising the *South Pyrenean* and *Axial* classic zones) can be distinguished (Fig. 5a) (Muñoz, 1985). The north-verging zone, referred to here as North Pyrenean, over-thrusts the Aquitania Tertiary basin along the so-called *North Pyrenean frontal thrust* and shows progressive north-verging thrusts from a strip of vertical structures (the boundary between north- and south-verging zones) to the frontal thrust. The south-verging zone, here named the *South Pyrenean*, is separated from the Ebro foreland basin by a frontal thrust, the *South Pyrenean frontal thrust*, which shows an irregular trace and is sometimes hidden under recent sediments. This southern zone consists of several thrust sheets occupying a broader zone than the northern one. The lower thrust sheets contain nappes involving basement and cover in a complex anti-formal stack structure. The upper sheets correspond to detached cover nappes which contain piggy-back basins and spread over the foreland basin with lateral terminations. This tectonic arrangement corresponds in depth to the underthrusted Iberian crust below the European one in a classic collision disposition with the cryptic suture placed along the boundary between the North and South Pyrenean zones (Fig. 5a, b). Balanced cross-sections estimate a crust shortening ranging from 125 km (Vergés et al., 1995) to 147 km (Muñoz, 1992) in the eastern and central Pyrenees. In the westernmost segment, these values decrease by up to 80 km (Teixell, 1998).

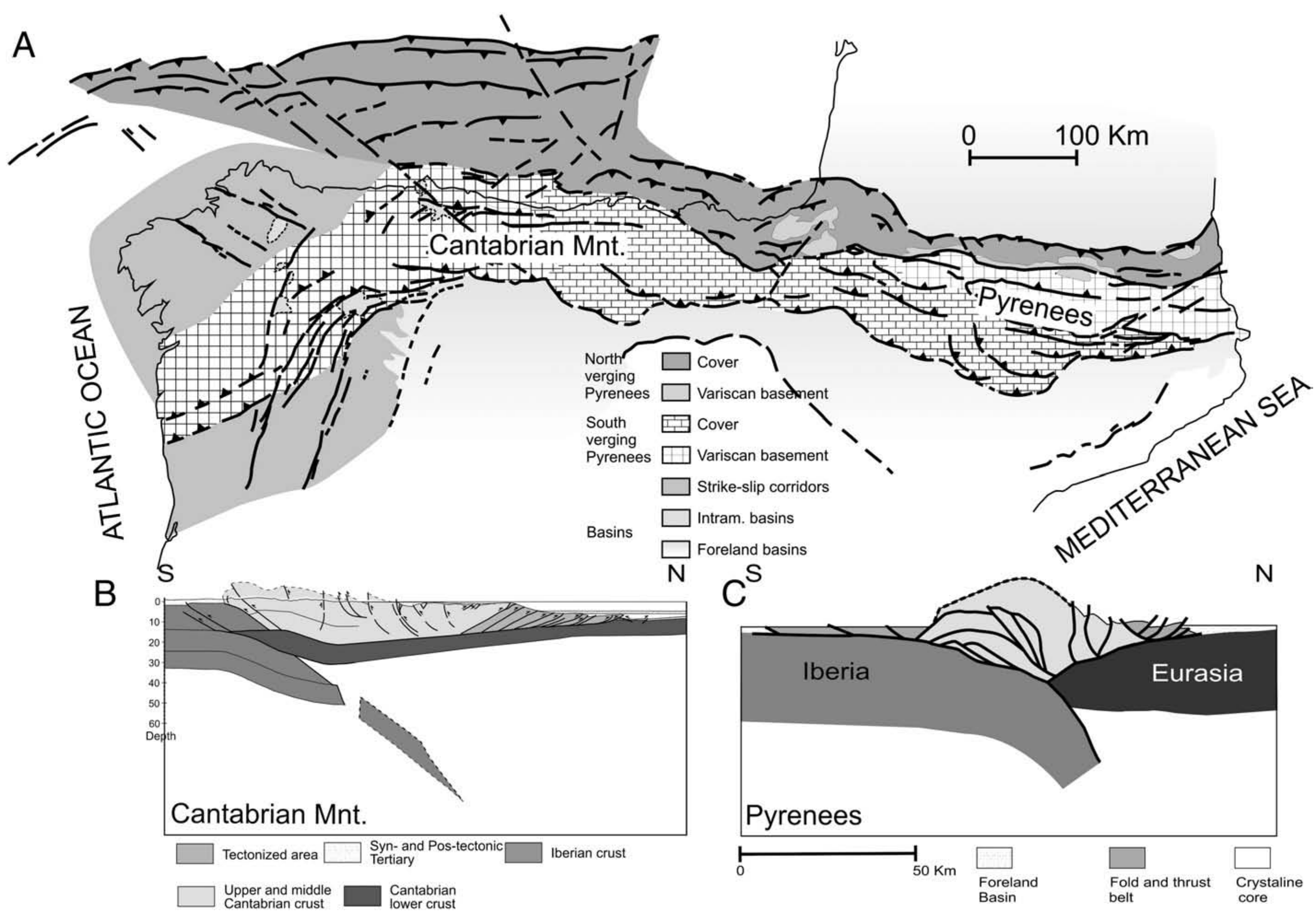

Fig. 5. A) Tectonic map of the Pyrenees showing the main vergences (North and South) B) Cross section, Pyrenees (Muñoz, 1992). C) Cross section, Cantabrian Mointains (Gallastegui, 2000).

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The Pyrenean structures continue westwards in the topographic relief of the coastal Basque region and in the more interior ranges which border the narrow corridor connecting the two large Ebro and Duero basins (Figs. 4 and 5). This region is called in literature as the Basque–Cantabrian Basin, although, surprisingly, it corresponds to the Basque coastal ranges that connect the Pyrenees and the Cantabrian Mountains. In this differentiated, intermediate segment – perhaps better described as Basque Mountains – the North Pyrenean front continues off-shore in the southern border of the Landes Plateau, a submerged prolongation of the undeformed Aquitania foreland basin, whereas the South Pyrenean border represents the prolongation of the limit with the foreland basin along the above-mentioned corridor. This region corresponds to the deformation of a deep Mesozoic basin where the North Pyrenean Zone is represented by north- to northeast-verging structures which continue off-shore in the margin. The South Pyrenean zone continues in the south-verging open folds of the Mesozoic cover which are detached from the basement and thrust over the Duero Basin. The deep structure of this segment has been interpreted as an underthrusting of the Iberian crust under the continental crust of the Landes Plateau. As in the neighbouring western Pyrenees, the cryptic suture may correspond to a zone of opposing vergences. A crust shortening of 86 km has been calculated for this sector (Pedreira, 2004).

The western part of the northern border of Iberia corresponds to the *Cantabrian Mountains*, a basement uplift which runs parallel to the shore line forming a prominent topographic relief. This basement uplift overthrusts the Duero Basin along a south-verging frontal thrust zone that can be considered as the prolongation, via the Basque region, of the South Pyrenean frontal thrust (Fig. 5c). Between this southern border and the shore line, other south-verging structures have been described (Alonso et al., 1996) and thereafter a prolongation of the South Pyrenean Zone can be envisaged. This type of deformation extends to the continental margin as well as in the north-verging structures of the continental slope (Gallastegui, 2000). The northern border of this off-shore deformation zone can be assimilated to the northwards shifted North-Pyrenean frontal thrust. The deep structure of this sector appears to be quite similar to the other segments of the northern Iberian border — including a calculated crustal shortening of 96 km (Gallastegui, 2000). Nevertheless, there is a main difference with respect to the other segments described before, since the deep structure does not correspond to the inversion and eventual decoupling of a former intra-continental plate boundary, as occurred in the deep trough of the Basque segment and in the transtensional broad realm of the Pyrenees. Instead, the Cantabrian Mountains represent the deformation of a Mesozoic rifted margin within which a crust-scale thrust uncoupled it from the undeformed Variscan crust. In fact, this decoupling created a new convergent plate-boundary that must be considered as the successor of the divergent one responsible for the spreading of the Bay of Biscay oceanic floor.

To the west, the southern Cantabrian thrust border (i.e. the South Pyrenean frontal thrust) and the associated structures continue without interruption along a series of arched thrusts that connect the E–W Cantabrian structures to two left-handed, NNE–SSW crust scale strike-slip corridors. The convex-to-the-south curved thrusts represent the compressive terminations of the left-handed corridors whose role is to efficiently transfer the compressive deformation, and subsequent crustal shortening, to the interior of the Iberian Peninsula (Vegas et al., 2004).

In a similar context, two right-handed NW–SE strike-slip faults (Cabrera et al., 2004) transfer part of the Cantabrian compressive deformation, and concomitant crustal shortening, to the NE–SW segment of the Galicia margin which connects the Cantabrian and Atlantic margins in the north-western corner of the Iberian Peninsula. It is this mechanism of transference which is the main cause of the western termination of the Cantabrian uplift and related crustal underthrusting. The area left between the two conjugate zones of transference is named here as the *Galicia Massif* which is characterized by a more peneplained landscape.

The above-described crust shortening and related compressive features resulted in the onset of plate convergence along a former, complex Iberia–Eurasia boundary. This Mesozoic plate boundary comprised three different segments, corresponding, from west to east, to an active oceanic spreading ridge (the Bay of Biscay ridge), a narrow intra-continental zone of deformation, which must assume extension and translation (the Mesozoic rupture of the Basque region) and a broad intra-continental zone of extended crust within which the Africa (with Iberia)–Eurasia transform-type relative motion was accommodated (the Pyrenean Mesozoic realm). Current plate reconstructions also describe some ill-defined transverse zones of transference. In this boundary, the general transtensional regime changed to compressive at the end of the Cretaceous, when the Africa–Eurasia relative motion changed to N–S convergence.

In the wake of this convergence, sea-floor spreading ceased in the Bay of Biscay in later Campanian times whilst tectonic inversion initiated in the intra-continental extended realms of the Basque–Gothic and Pyrenean regions. The way in which this inversion took place can be constrained within the frame of the plate reconstructions (Olivet, 1996; Thinon, 1999). In a first stage, decoupling of the Iberian and European lithospheres must have occurred in the eastern part of the Pyrenean realm, giving rise to a sort of incipient A-type subduction of the Iberian lithosphere under the European one. Later, this decoupling or rupture (and related underthrusting) propagated westwards in the Pyrenean realm and reached the Basque and Cantabrian segments, achieving the inversion of all the intracontinental part of the Mesozoic Iberia–Eurasia limit.

In the oceanic segment of this Mesozoic limit, the convergence was accommodated in the continental margin by means of the propagation of the intra-continental underthrusting. It seems odd that there was a localization of decoupling in the interior of the continental margin since the ocean-continent boundary is usually considered as the weaker zone in incipient convergent boundaries. Nevertheless, in this plate boundary, some aspects, such as the young oceanic lithosphere and the propagation of a north-verging crust-scale thrust must be taken into account. Once the crustal underthrusting was formed, its westward propagation ought to have been easier and with the same polarity within the weakened continental margin, as suggested by Gallastegui (2000). The western termination of the crustal underthrusting corresponds to strike-slip structures that transfer the vertical deformation to both the interior of Iberia and the oceanic realm. In its turn, the continuation of the convergent plate boundary corresponds to a compressive segment, the Biscay rise and an extensional one, the King's Trough, connected to the Mid Atlantic Ridge.

During and after blocking of convergence in the plate boundary, deformation was transmitted to the interior of the *lower plate* (i.e., the understhrusted plate), the Iberian foreland, causing most of the present geomorphic features. Moreover, the transmission of deformation to the foreland was effective not only in the Iberian Peninsula, as coherent and coeval geomorphic features were also developed in the African foreland. In this sense it can be concluded that convergence in the *Pyrenean* plate boundary triggered the development of the main topographic features in the Iberian Peninsula and North Africa, as far as the Anti-Atlas.

4. The intraplate deformation

The collision process at the Cantabro–Pyrenean edge led to the transmission of active tectonic stresses towards the interior of the Iberian microcontinent during the Eocene–Lower Miocene. The intraplate accommodation of this shortening was in three E–W crustal folds, connected by fault corridors which also affected all the crust. The structures generated by crustal buckling are the Central

System, the Toledo–Guadalupe Mountains and the Sierra Morena (Fig. 4). They correspond to crustal-scale folds and their related thrusts in the upper crust were developed on a Variscan basement with a thermotectonic age of 350 Ma, and seem to be decoupled from lithospheric-scale folds (Cloetingh et al., 2002). Given their size, these folds affect the Iberian Massif, including its off-shore prolongation, and are oblique, and even perpendicular, to the Variscan grain, indicating that their general orientation was imposed by their geometric relation with the E–W Cantabro–Pyrenean edge. The crustal folds affect also the eastern part of the Iberian Peninsula, where there was a structural NW–SE grain and a thinned crust inherited from the Triassic rifting. The interference of the central fold (Central System) with the NW–SE structural grain in the Iberian Chain gave rise to an apparently complex structure, where the deformation was distributed between E–W thrust and right lateral NW–SE faults. The strike-slip movement appears to be concentrated in the Castilian Branch of the Iberian Chain, where the interference of structures is due more to deformation partitioning than to the superimposition of several *tectonic phases*.

On the Atlantic shore, another strike-slip fault corridor has been mentioned and this connects the western termination of the Pyrenees with the Central System (Vegas et al., 2004): the Vilariça and Régua Fault System, with a NNE direction and left lateral movement (traditionally considered as Late-Variscan in part). In this case, the absence of sedimentary cover and the presence of small Cenozoic basins produce types of structures which are easier to interpret. Their contrast with the fault corridors in the Iberian Chain will enable us to better understand how the Cenozoic inversion was accommodated in the Mesozoic rift formed in the eastern part of the Iberian Peninsula.

Therefore, such a topographic pattern of the Iberian Peninsula, mountain ranges and elevated topographic areas related with strike-slip fault corridors are also evident. We provide a more detailed tectonic description of these two corridors with substantial horizontal slip below.

4.1. The Central System

The Spanish Central System constitutes the most prominent topographic elevation in the interior of the Iberian Peninsula (Fig. 6a).

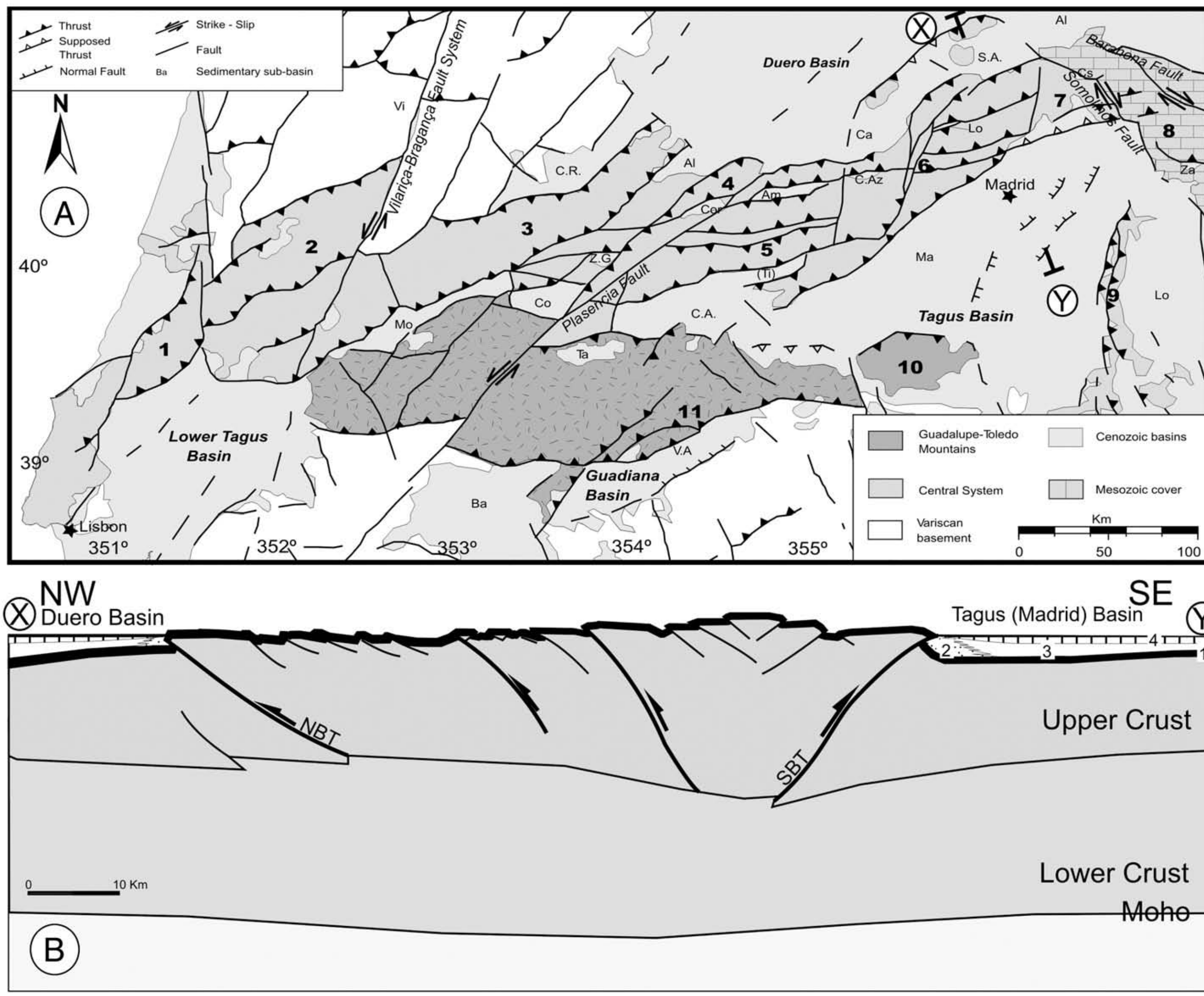

Fig. 6. A) Tectonic sub-units of the Spanish–Portuguese Central System. 1—Montejunto Range. 2—Serra da Estrela. 3—Gata-Peña de Francia. 4—Sierra de Avila. 5—Gredos. 6—Somosierra and Guadarrama. 7—Transition zone (Tamajon thrusts). 8—Iberian Chain. 9—Altomira Range. 10—Toledo Mountains. 11—Guadalupe Ramp. Minor Basins: Vi—Vilariça. Mo—Moraleja. Co—Coria. CR—Ciudad Rodrigo. Ba—Badajoz. VA— Vegas Altas. CA—Campo Arañuelo. Am—Ambles. Caz—Campo de Azalvaro. Lo—Lozoya. SA—Sepulveda–Ayllon. AL—Almazan. Za—Zaorejas. Lo—Loranca. Ma—Madrid. Ti—Tietar. B) Cross section showing the upper crustal structure of the Somosierra–Guadarrama zone of the Central System (De Vicente et al., 2007a,b,c), 1) Late Cretaceous sedimentary cover, 2) Sin-tectonic conglomerates, 3) Paleogene, 4) Neogene. SBT, NBT, Southern and Northern Border Thrusts.

It corresponds to the main divide which separates two large Tertiary basins, the Duero Basin to the north and the Tagus (Madrid) Basin to the south. It extends in a roughly ENE–WSW direction for more than 300 km with peaks reaching more than 2500 m. Together with the Sierra de Gata, Malcata, and the Estrela–Lousã mountain ranges (Portuguese Central System), it forms a stepped ("en echelon") range of NE–SW basement elevations, producing a topographic high that runs along more than 700 km, from near the Atlantic Ocean to the western border of the Iberian Chain. The Estremadura Spur in the Atlantic offshore can also be considered as a continuation of this range. The Central System can be described as a Cenozoic E–W to NE–SE directed basement uplift comprising several stepped segments. This crustal fold which affects the whole crust (thick-skinned tectonics, see De Vicente et al., 2007 for a recent and detailed description) with a total shortening of 22% (De Vicente et al., 1996) and a maximum accumulated structural relief of more than 5 km. It shows an antiformal geometry in the upper crust with thickening in the lower crust (Suriñach and Vegas, 1988), (Fig. 6b).

Deformation along this basement uplift is clearly asymmetric, as is evidenced by the existence of an unique and large (crustal-scale) thrust at its southern border, while at the northern border there is a normal sequence of NW verging thrusts (progressively younger thrusts towards the Duero foreland basin), whose activity ended during the Lower Miocene. This deformation was accomplished under axial compression corresponding to a NW–SE to NNW–SSE shortening, The age of this episode of deformation is Oligocene–lower

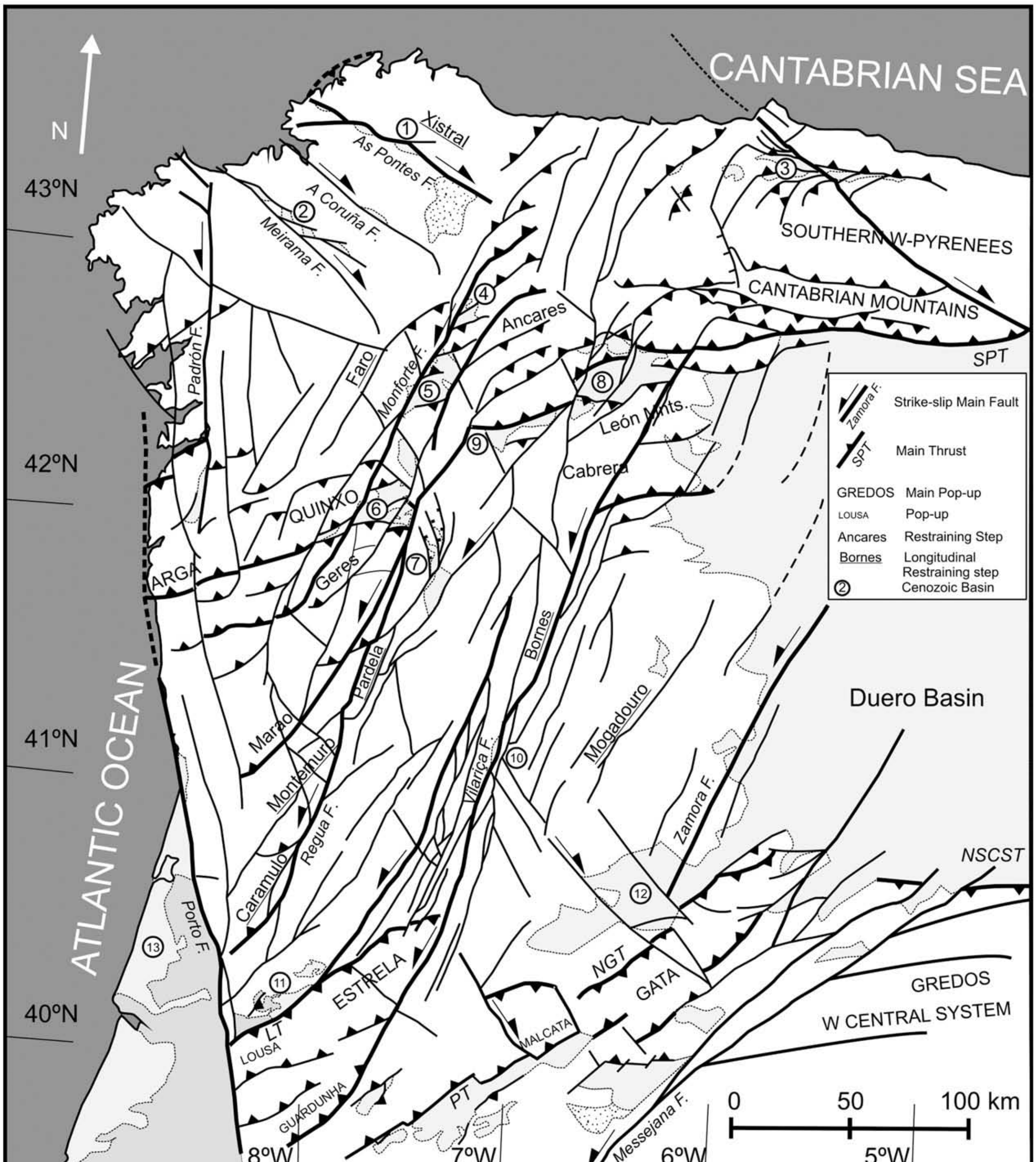

Fig. 7. Tectonic map of the distributed deformation of the Vilariça left lateral strike–slip fault system. Cenozoic Basins: 1) As Pontes. 2) Meirama. 3) Oviedo. 4) Sarria. 5) Monforte. 6) Xinzo. 7) Verín. 8) El Bierzo. 9) O Barco. 10) Vilariça. 11) Lousa. 12) Ciudad Rodrigo. 13) Lusitanian.

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Miocene and Oligocene–Pliocene (mainly upper Miocene) in the eastern and western segments respectively (Cabral and Ribeiro, 1990; De Vicente et al., 1996).

The sedimentary cover is absent to the W, whereas it increases in thickness towards the E, but no basement decoupling is founded. Lateral and oblique ramps of the main thrusts are oriented N–S to NNE–SSW (left lateral) and NW–SE (right lateral).

The easternmost zone has a tectonic evolution close to that of the Iberian Chain, where extensional stresses are recorded during the upper Miocene to present day time span. Nevertheless, active thrusting is registered in the central (De Vicente et al., 2007a) and the most westerly zones (Cabral and Ribeiro, 1988; Cabral, 1995; Ribeiro et al., 1996; Ferreira-Soares et al., 2005) where an M 6 reverse focal mechanism has been documented (Stich et al., 2005). Important uplifting in the central zone of the range has been deduced form apatite fission track analysis (De Bruijne and Andriessen, 2002).

Compression with a NW–SE S_{hmax} seems to have been the characteristic stress regime for the western Central System since the Miocene.

4.2. The Vilariça fault system

In the NW of the Iberian Peninsula, the deformation front associated to the S Pyrenean Thrust is distributed among more thrusts with fewer individual vertical offsets. They represent an imbricate system with a tectonic transport direction mainly towards the S, with some important backthrusts, and superimpose the S edge of the Cantabrian Mountains over the El Bierzo Basin (Heredia et al., 2004).

From this point, the mountain alignments with N10–30E trends predominate, and are structured towards the Vilariça Fault System (Cabral, 1989; Cabral and Ribeiro, 1990) up to the S of the Estrela–Lousã Massif.

Three main NNE–SSW faults can be recognised in the tectonic mapping: From E to W, 1) the Vilariça Fault (Bragança–Vilariça–Manteigas) with a length of more than 200 km and a maximum left lateral horizontal offset of 8 km (Cabral, 1989), 2) the Régua Fault (Verín–Régua–Penacova) and 3) the Monforte Fault, the shortest of the three. The Zamora Fault, somewhat less defined, limits the Duero Basin to the W (Fig. 7).

In the NW corner of Galicia, right lateral faults trending N110–140E predominate (As Pontes, Meirama). This layout is lost to the E of the Monforte fault. As a whole, this corresponds to a transpressive deformation zone which locally contains small Cenozoic basins in the restraining bends of the faults (Santanach, 1994).

At least in part, the Vilariça fault system must have activity subsequent to the Pyrenean thrusts (Cabral, 1989; Cabral and Ribeiro, 1988; Cabral, 1995; Ribeiro et al., 1996; Heredia et al., 2004), as the instrumental seismic activity indicates (De Vicente et al., 2008).

This alpine deformation has been described as the interference of structures due to the simultaneous accommodation of the crust in the *Cantabrian–Pyrenean Front* and in the *Vilariça Fault System* (Heredia et al., 2004).

However, as has been mentioned above, it is necessary to bear in mind that the S Pyrenean Thrust (Cantabrian) in an E–W direction is the surface expression of a limited subduction towards the N of the lower and intermediate crust of the northern Iberian margin (Gallastegui, 2000). The Vilariça Fault System indicates that, towards the W, the shortening cannot be assumed by the relatively simple mechanism of a detachment under the Western Pyrenees (Cantabrian Mountains). On the contrary, the shortening is accommodated in the interior of the crust, allowing the transfer of the deformation more towards the S, up to the Central System. Therefore, this is not interference between two types of structures which have been able to move at different times, but rather a transition in the style of the deformation between two closely related types of accommodation of the deformation.

The Cenozoic El Bierzo Basin plays the key role in the interaction between the S. Pyrenean Thrust and the Vilariça Fault System is the Cenozoic basin of El Bierzo. The faults of Régua and Vilariça (?) converge here forming a terminal restraining stepover which, towards the E is structured in a series of narrow pop-ups and pop-downs, characteristics of the south-western border of the Cantabrian Pyrenees (Fig. 8a).

Between the Monforte and the Régua faults, the Cenozoic basins of Sarria, Monforte and Xinzo, have a very similar structure: pop-down flower structures between restraining bends of the Vilariça–Régua–Monforte left lateral fault system. Farther to the S, along the layout of the Vilariça Fault, where the deformation associated to strike-slip increases, the geometry of the basins is elongated, with no bends, and prominent scarps in the direction of the faults. The best example is the Vilariça Basin in a NNE–SSW direction, 20 km long and 2–3 km wide (Cabral, 1989) (Fig. 8b, c). Taking this into account, the Vilariça Fault System transfers the Pyrenean deformation towards the S through a relatively wide transpressive zone, measuring approximately 100 km in width, and with a length of 300 km which connects the two larger structures generated by buckling of the whole crust (Pyrenees and Central System) through a series of compressive splays (Fig. 9a). Nevertheless, important thrusting activity is also registered up to the Upper Miocene (Cabral and Ribeiro, 1988).

From our morphotectonic analysis, the Vilariça Fault ends in the NE with a series of horse tail splays up to the S. Pyrenean Thrust. The most important of these lifts the Montes de León which extends towards the S, in a longitudinal restraining step (Serra Bornes) until it connects with the NE edge of the Estrela–Lousã Massif. Therefore, it is in this intermediate stretch of these faults that the horizontal movement is larger.

This structural system ends abruptly towards the Atlantic in the Porto Fault (Porto–Coimbra–Tomar) (Cabral and Ribeiro, 1990). As a whole, the Vilariça, Régua, Monforte Fault Systems can be interpreted as a corridor of left lateral strike–slip faults which end in embedded splays. In the opposite sides to the compressive splays right lateral strike–slip faults are frequent (Galicia Massif, E of the Estrela–Lousã Massif) (Fig. 9a.).

The importance of this structural system can be observed in the Moho depth map of the zone (Tesauro et al., 2007). To the W of the Vilariça Fault, there is a clear thinning of the crustal thickness which appears at 37 km in the Montes de León and 30 km in the area of Verín. This clear structural step in the Cantabrian Range has been interpreted to be related to the Variscan paleo–Moho preserved during 300 Ma. (Gallastegui, 2000). In the light of our interpretation, the abrupt changes in the Moho depth of the NW of the Iberian Peninsula are due to the Cenozoic deformation associated to the transmission of the Pyrenean deformation towards the Central System (Fig. 9b, c).

4.3. The Castilian branch of the Iberian chain

The Iberian Chain is an intra-plate chain with a thick sedimentary cover in the interior of the Iberian micro-continent. The main thrusts are located to the N (Demanda Unit) and in the middle (Montalbán–Utrillas Thrusts), both with tectonic transport towards the N. Both structures are the final result of the inversion of faults originated during the intra-plate extension during the Early Cretaceous (Salas et al., 2001). However, the origin as a basin of this inverted zone during the Cenozoic is related to the Permo-Triassic rifting processes (Alvaro et al., 1979; Sopeña, 2004).

The Castilian Branch of the Iberian Chain is a NW–SE structural alignment which constitutes the westernmost sector of this intra-plate deformation zone where further deformation is accumulated due to strike–slip (Fig. 10a, c). The N–S structural grain of the variscan basement does not play a relevant role in Cenozoic cored thrusts.

From the analysis of the isopachs of the Buntsandstein formation, the geometry of the rifting related normal faults can be reconstructed.

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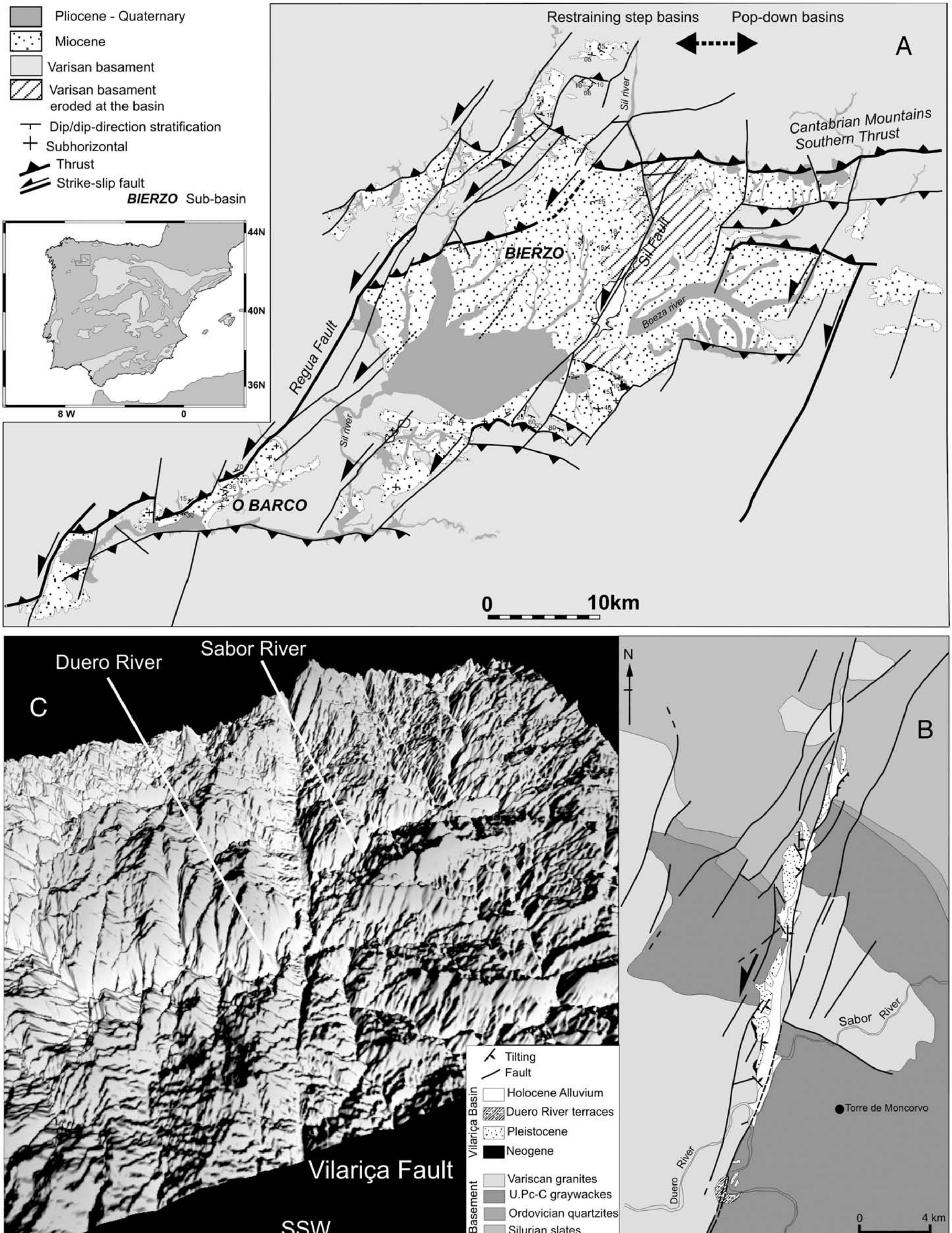

Fig. 8. A) Tectonic map of El Bierzo cenozoic Basin at the intersection of the S. Cantabrian Thrust and the Vilariça Fault System. B) Vilariça Basin tectonic map (Cabral, 1989). C) Digital elevation model of the Vilariça Fault crossing the Duero River stream.

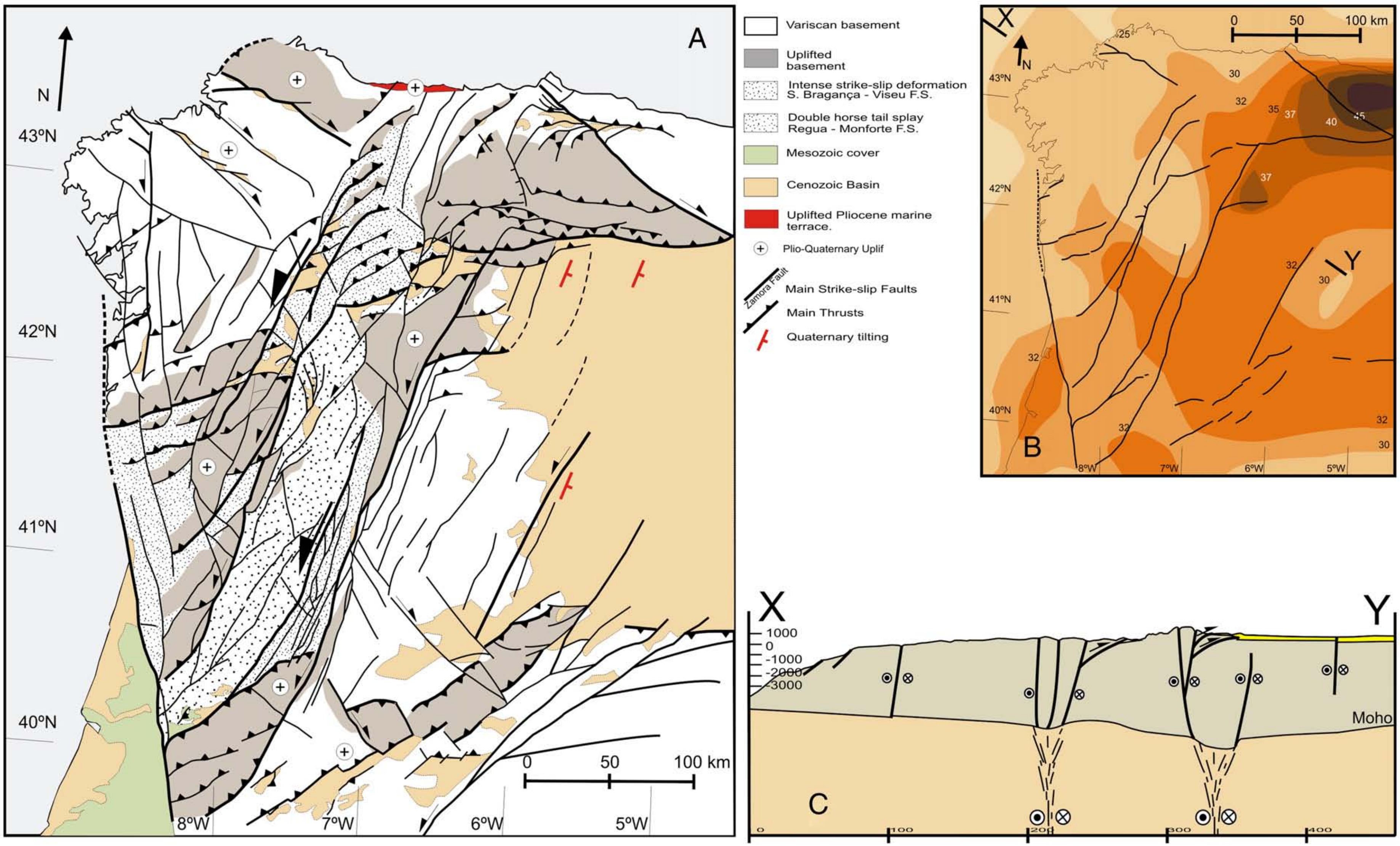

Fig. 9. A) Morphotectonic interpretation of NW Iberia indicating the main relief zones along the Vilariça Fault System. B) The Moho depth at the Vilariça Fault System zone (Tesauro et al., 2007). C) Interpretative cross section of the upper crustal structure related to the Cenozoic deformation in the Vilariça Fault System.

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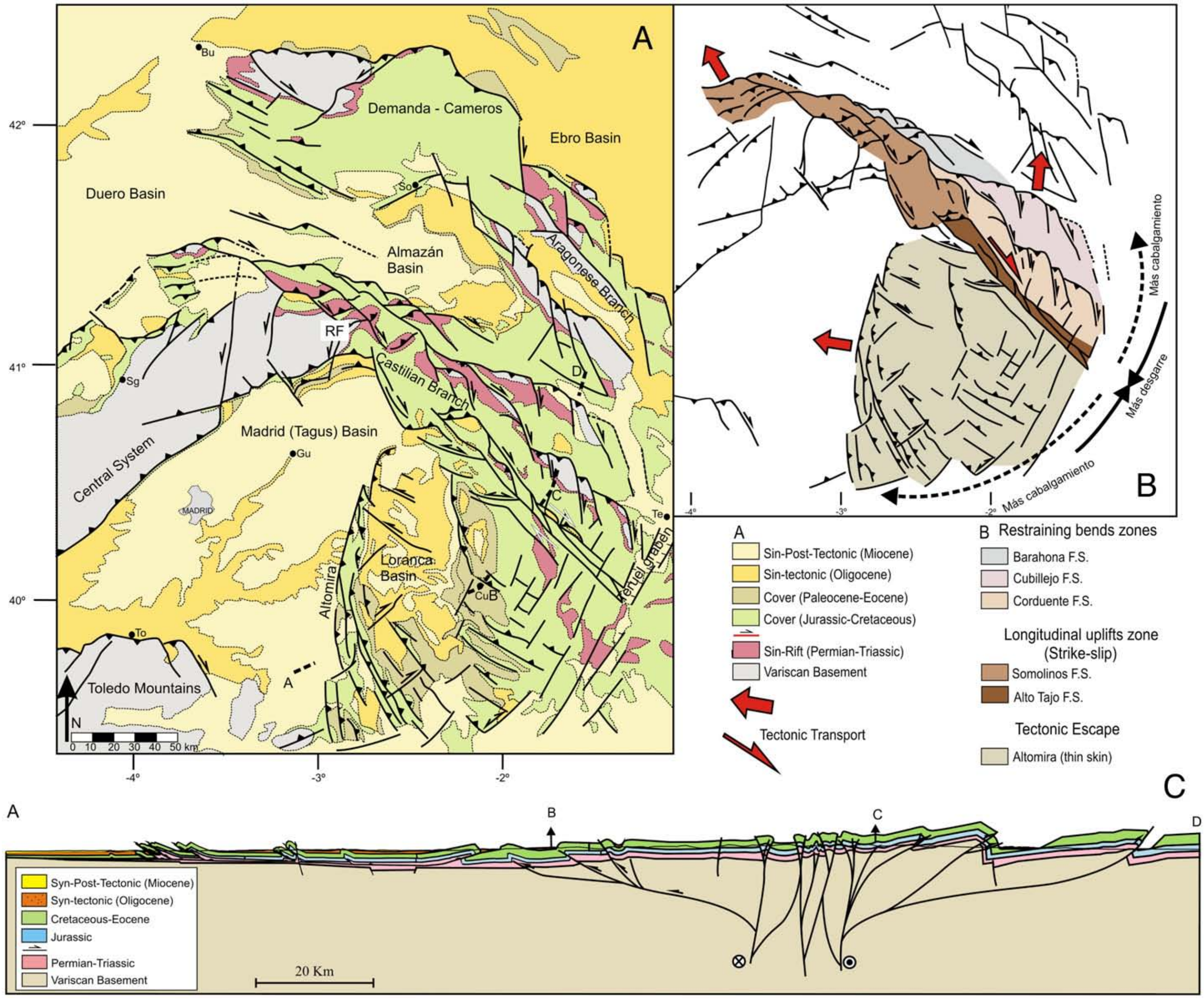

Fig. 10. A) Tectonic map of the Cenozoic deformation in the Castilian Branch of the Iberian Chain RF, Riba de Santiuste fold (shown in Fig. 13c). B) Tectonic interpretation and main sub-units with mean vergences and tectonic transport. C) Cross section showing the right lateral strike–slip related deformation, and positive flower structure, of the Castilian Branch of the Iberian Chain.

The Somolinos Fault would be the most important western structure of the Triassic Rift (edge of the rift) (Arche and López-Gómez, 1996; Sopeña, 2004). However, the general geometrical pattern corresponds more to a zone in which the extension is distributed in many faults, and appears scarcely concentrated in individual long grabens (Fig. 11a).

The extensional Cretaceous stage, which is so important in other sectors of the Iberian Chain, does not appear here except for a slight NW–SE synclinal at the base of the Utrillas Formation (Late Cretaceous) (Fig. 11b).

During the Cenozoic there was a (partial) inversion of these faults. As regards the contractive structure (Fig. 10b), the Castilian Branch of the Iberian Chain has a variety of tectonic transport directions. This structuring, together with the location of the most important thrusts, makes it possible to define a series of sub-units separated by faults, or long systems of faults: 1) The Barahona Fault System which provides details of transpressive culminations with different linkage between strike-slip faults, oblique faults and thrusts (Bond, 1996). 2) The Somolinos Fault System that terminates towards the NW in a series of imbricated thrusts between divergent splays. In its central part there is a number of short en echelon folds vergent towards the SE with NE–SW axes, and which affect the Variscan basement. Southwards this fault system ends at a restraining bend that lodges the Cenozoic Basin of Zaorejas. This sub-unit is prolonged towards the SE in a narrow corridor which shows the highest deformation levels in the Castilian Branch: 3) The Alto Tajo Fault System (Rodríguez-Pascua and De Vicente, 1998). It is a flower like pop- up structure with very straight NW–SE folds over a long course (Fig. 12a), with sub-vertical axial planes and box geometry. In some synclinals there are Cenozoic sediments which trace long NW–SE basins. We have interpreted this group of structures as forced folds (bending folds) along the layout of the NW–SE right lateral strike–slip faults, rooted in the basement and affecting the cover to different degrees (Fig. 12b).

Towards the N a series of restraining bends, with basement outcrops, accommodate most of the horizontal movement of the 4) Corduente strike–slip to thrust zone, through E–W thrusts vergent to the N. As a whole, this is a transpressive area with less horizontal movement than the previous one. The Piqueras Cenozoic basin is located in the footwall block of a restraining bend, where simultaneous progressive unconformities

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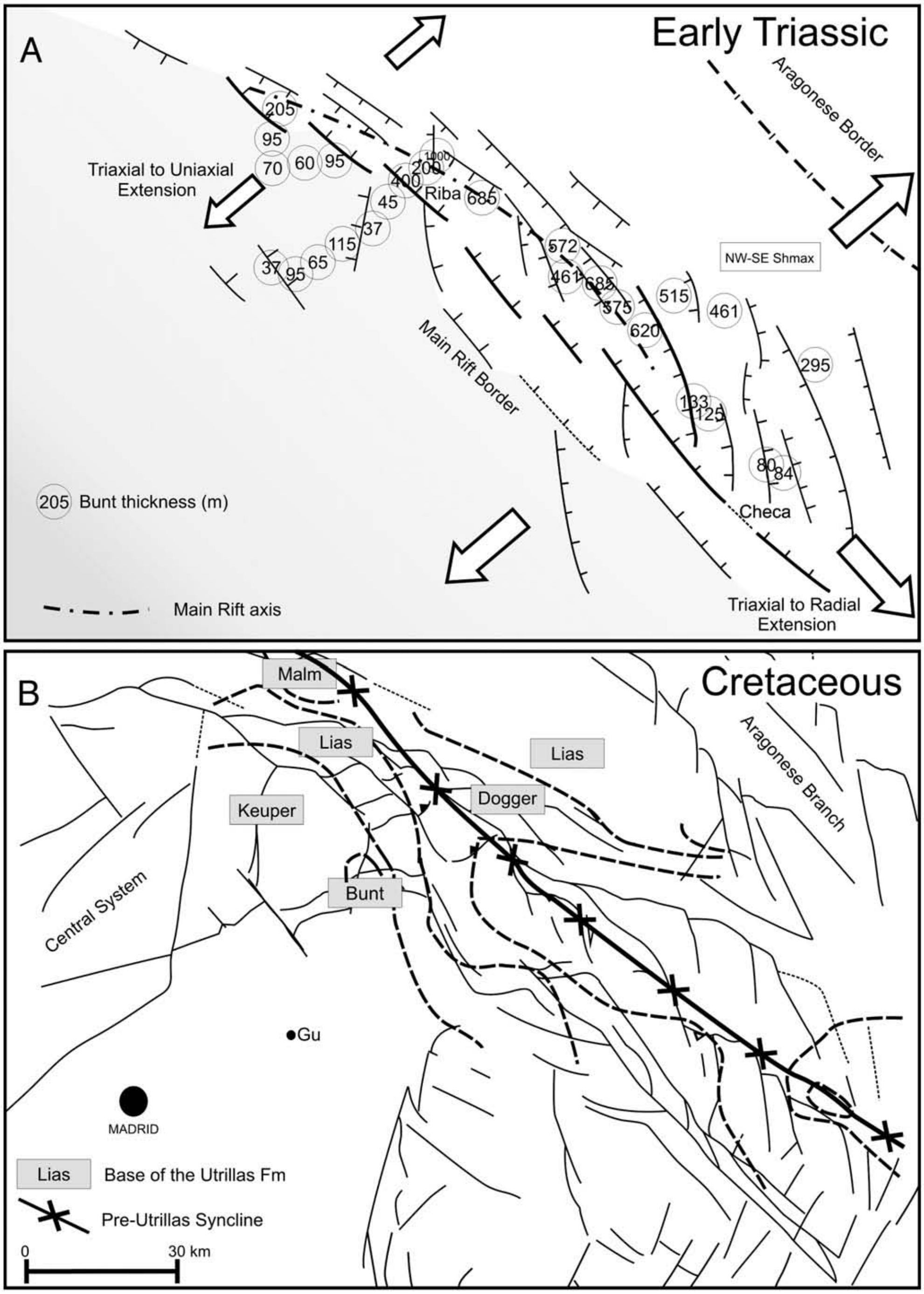

Fig. 11. The Mesozoic extensional processes previous to the Cenozoic (partial) inversion of the Castilian Branch. A) Triassic rifting (Buntsandstein). B) Early? Cretaceous intraplate extension.

are visible, on the thrust edge (E–W) and in the strike–slip related border (NNW–SSE).

More to the north, and limited by the Cubillejo Fault, a zone appears which almost mimics the one described above, but with a greater thrust component towards the N (5) Cubillejo Thrust to strike–slip zone.

6) The Altomira Unit is very different from those mentioned above. It is a belt of thick skin folds and thrusts, in the E, and thin skin in the W, whose tectonic transport (SW to W) and orientation of structures progressively change from E to W. This is a tectonic escape towards the W under constrictive conditions of the deformation (Muñoz-Martín et al., 1998) in a generalised stress field with N–S compression. At the SE end of this unit there are NE–SW folds, which are clearly linked to thrusting, and are sub-parallel to those of the Alto Tajo, therefore, a deformation partitioning did also occur in structures with the same orientation.

Therefore, the interference of structures is common throughout the Iberian Chain, with different chronological relationships between the different fold directions (Simón-Gómez, 1986; Andeweg et al., 1999). In our opinion, this is the result of one single process of strain partitioning under constrictive conditions of the deformation. In this situation, the earlier, first order structures would control and force the local, simultaneous occurrence of one or another type of paleostresses. The geometry of the rift, previous to the inversion, may have led to fold interference (and associated stresses) during the same tectonic event, as has been recognised in the High Atlas in a very similar tectonic situation (Beauchamp, 2004).

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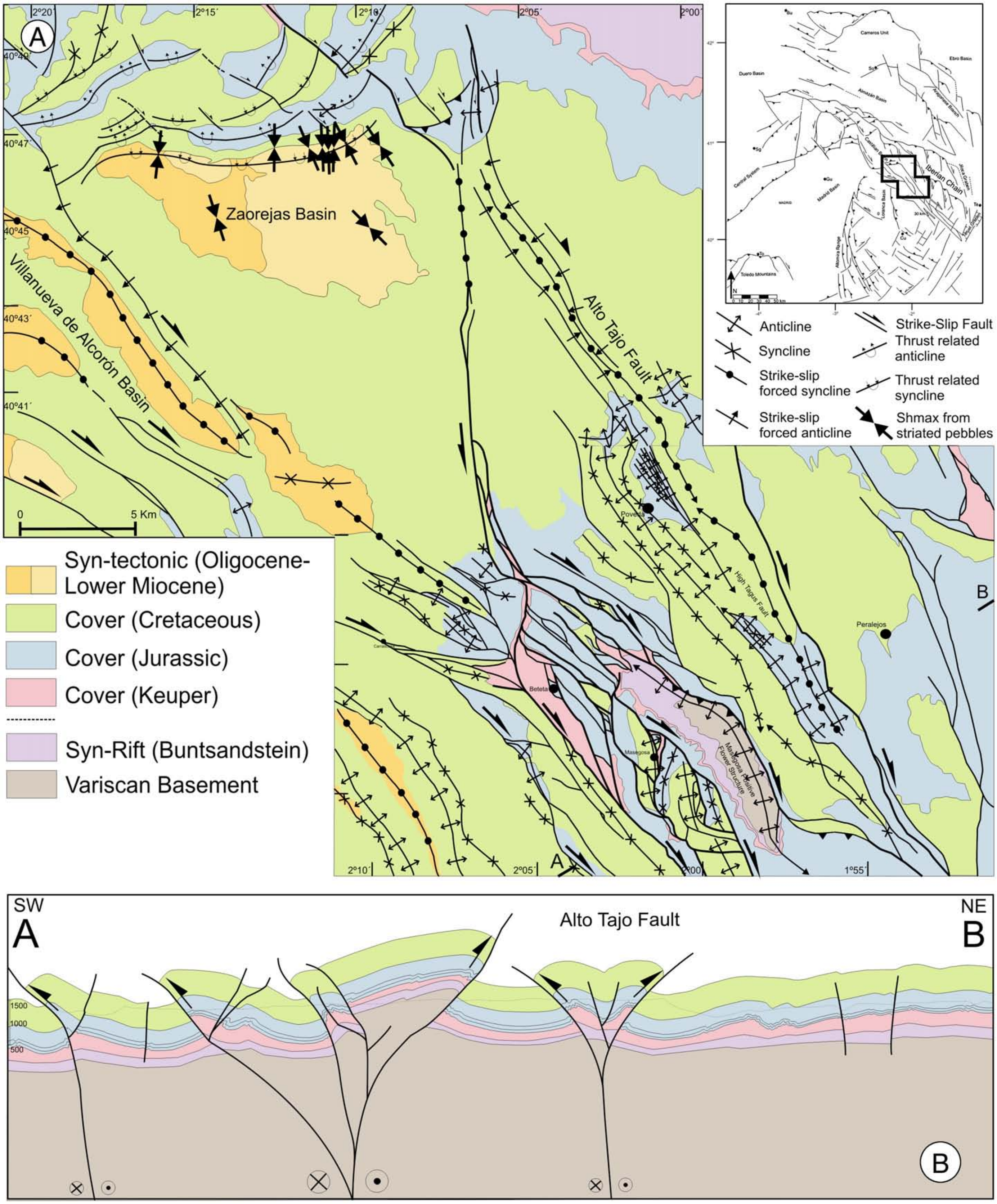

Fig. 12. A) Tectonic map of the Zaorejas Basin and the Alto Tajo Fault System. B) Cross section of the general positive flower structure.

This group of macro-structural tectonic characteristics can be explained through a generalised N–S shortening (Guimerà et al., 2004). It is, above all, the prior E–W structures which best accommodate the inversion in the Iberian Chain. However, in the Castilian Branch, this was not the predominant structural grain. Therefore, this involves an oblique inversion, where the previous main structures (normal NW–SE faults) work as strike–slip faults. In many cases, the thrust structures nucleate in the accommodation zones and relay ramps of the old Triassic Rift (De Vicente et al., 2007b) (Fig. 13a, b, c). With an

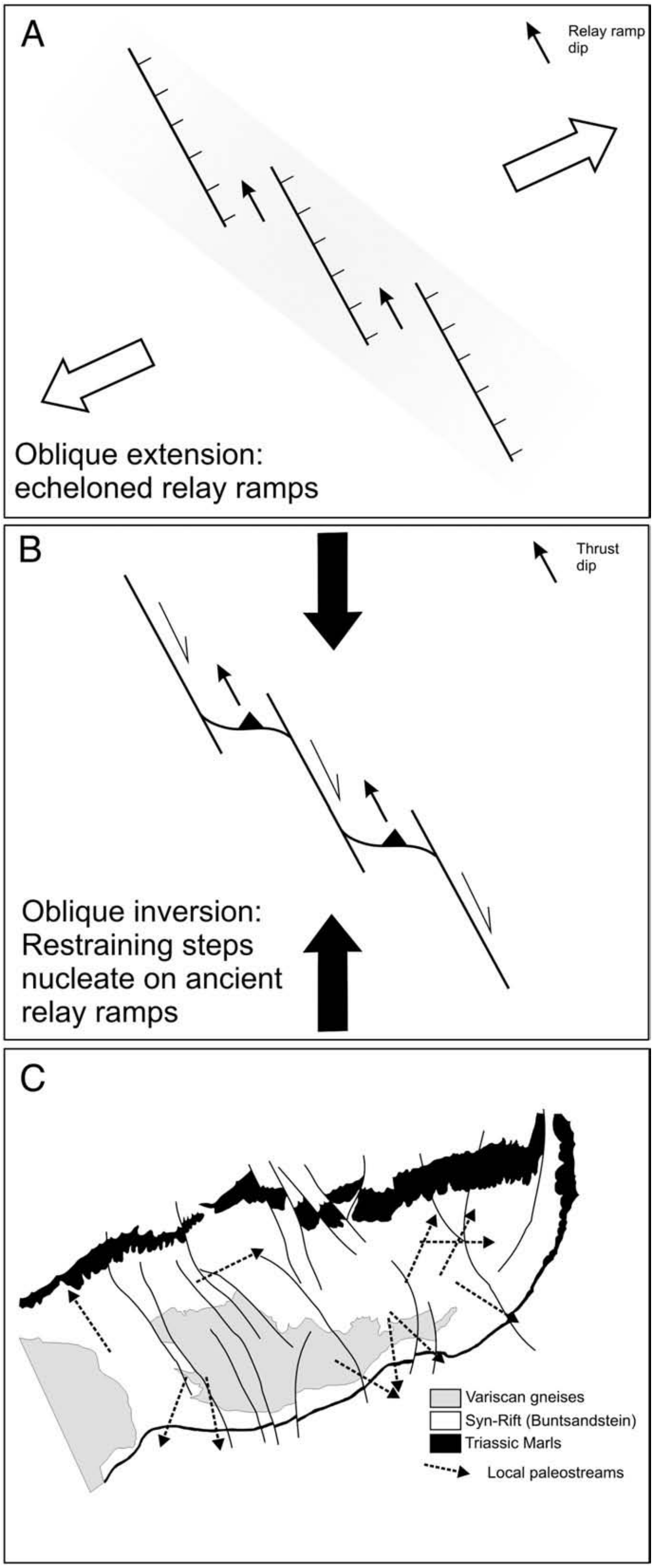

Fig. 13. A), B) Sketches showing how the nucleation of Cenozoic folds is located on ancient relay ramps and accommodation zones of the Triassic Rifting. C) Geological map of the Riba de Santiuste anticline with paleo-currents related to the Triassic infilling, indicating that the fold nucleation occurred on a sedimentation paleo-high (Sanchez-Moya, 1992) (Shown on Fig. 10a).

important horizontal movement component the inversion is just *partial*, as can be observed in the layout of the Somolinos Fault where, at the contact with the Central System.

Deformation by strike–slip increases from E to W towards the Alto Tajo Fault System. In the Castilian Branch this is very evident between the Cubillejo Fault and the Corduente Fault, where restraining bends are always E–W oriented.

5. Model of distributed deformation and resulting topography

From an analysis of the sedimentary fill of the Cenozoic basins and from the low temperature thermo-chronological data, it can be deduced that there was a common evolution as regards the time and the way in which deformation is distributed under strike–slip corridors and the thrusting through crustal buckling. The total thickness since the Paleocene reaches more than 5000 m in the Ebro Basin, 2500 m in the Duero Basin and 3500 m in the Tagus Basin. The common features in the sedimentary filling of the Iberian basins were shown by Calvo et al. (1993) and Calvo, (2004). The first deformations seem to have already been registered in the Eocene, with very thick sedimentary sequences continuous in time with higher sedimentation rates at the northern edge of the Pyrenean foreland (Duero to the W, Ebro to the E). The Duero and Tagus Basins begin to be individualised, and this marks the birth of the Central System as a range. In contrast, the intra-mountain basins in the Iberian Chain and in the Vilariça Fault System have an extremely varied record regarding thickness (in general low) and facies where the lacustrine and clastic alluvial deposits predominate.

In contrast with the previous more homogeneous stage, during the Oligocene–Lower Miocene, the sedimentary filling of the basins was characterised by the growing up of several tecto-sedimentary units, separated by discontinuities and/or sedimentary ruptures within short periods of time (Calvo et al., 1993; De Vicente et al., 2007a). Although the number and the specific ages of these units varied from one basin to another, indicating more activity in different range borders, the relevant fact is that the greatest number of sedimentary ruptures was concentrated during this period of time, thus this set of units can be interpreted as sin-tectonic.

From the Late Oligocene, wide saline lake systems were developed, which indicates that the saline materials of the Mesozoic and of the Paleogene had already been tectonically uplifted and were eroded.

During the Lower–Middle Miocene, the continued tectonic activity at the edges of the ranges gave rise to high sedimentary load–subsidence relations with wide expansion of lake systems, first evaporitic and then carbonated. This context indicates the existence of a large relief which was being rapidly eroded.

The Miocene of many basins can be divided into three units limited by discontinuities and/or sedimentary ruptures: the lower Ramblian–Aragonian, Lower Aragonian–Upper Vallesian and Upper Vallesian–Turolian (although the absolute ages are being revised) (Calvo, 2004).

Towards 9 Ma (Upper Vallesian or Lower Tortonian) a very important change takes place in all the basins. Strong discontinuities appear that separate stratigraphic successions deposited in well differentiated paleogeographic contexts. At this time a generalised change of regime took place from endoreic to exoreic (Calvo, 2004).

Without discarding global eustatism, this common evolutionary pattern as regards the filling of the Cenozoic basins of Iberia must be evidence of the changes in the dynamics of the plates, and of the effective transmission of the tectonic stresses from the active edges towards the interior.

Thus, although the deformation seems to have begun in the Eocene, it is during the Oligocene–Lower Miocene (Active N edge) that the effective folding of the upper crust was clearly evidenced in the sedimentary filling. A generalised uplifting also took place in the entire Peninsula and this explains the average height anomaly of Iberia. This uplifting is congruent with a very spread out deformation affecting a very wide area (Africa and Iberia mechanically coupled).

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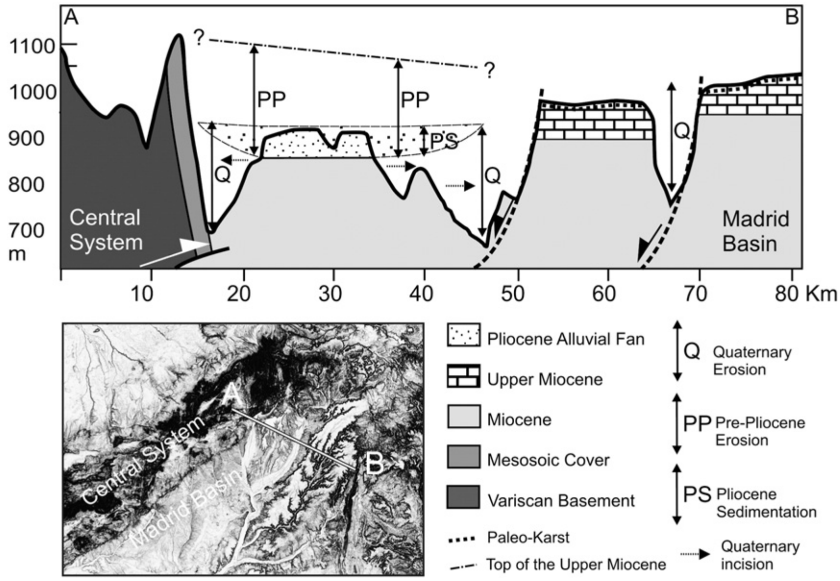

Fig. 14. Upper Miocene erosion, prior to Pliocene alluvial fan sedimentation, in the Madrid Basin. Pliocene alluvial fans are being sediment on an Upper Miocene paleo-surface.

The mechanical decoupling, which supposed the emplacement in the S (Betics) of the Alboran Block, was initially evidenced during the Middle Miocene by rapid erosion of the relief created in the previous stage. At the end of this period, there was some tendency to acquiesce with the development of erosion surfaces and, finally, the change towards the exoreism.

From this perspective, this drastic change in the evolution of the Cenozoic basins would be the delayed effect of the mechanical decoupling of Iberia and Africa. The Plio-Quaternary pulse, which implies the superimposition of the stress field which is still active today, reworks the previously created relief and is related to the development of wide Pliocene alluvial fans, immediately previous to the implementation of the Quaternary fluvial network. These fans are located on a paleorelief which had already partially eroded the previous Miocene filling. This is evidence of a greater capacity of transport (and sedimentation), within the tendency of the erosion, initiated at the end of the Upper Miocene (Fig. 14). Whether this Pliocene crisis is tectonic or climatic is still under debate (Martín-Serrano et al., 1996. De Vicente et al., 2007a). However, the data on Apatite fission tracks taken in the Central System (De Bruijne, 2001; De Bruijne and Andriessen, 2002.) do register Pliocene uplifting (or rather, exhumation).

There is more substantial evidence which indicates changes to the geodynamic configuration, and to the acting intraplate stresses, towards the 9 and 3.7 Ma. This involves the appearance of volcanism in the interior of the Iberian plate (Campos de Calatrava) probably

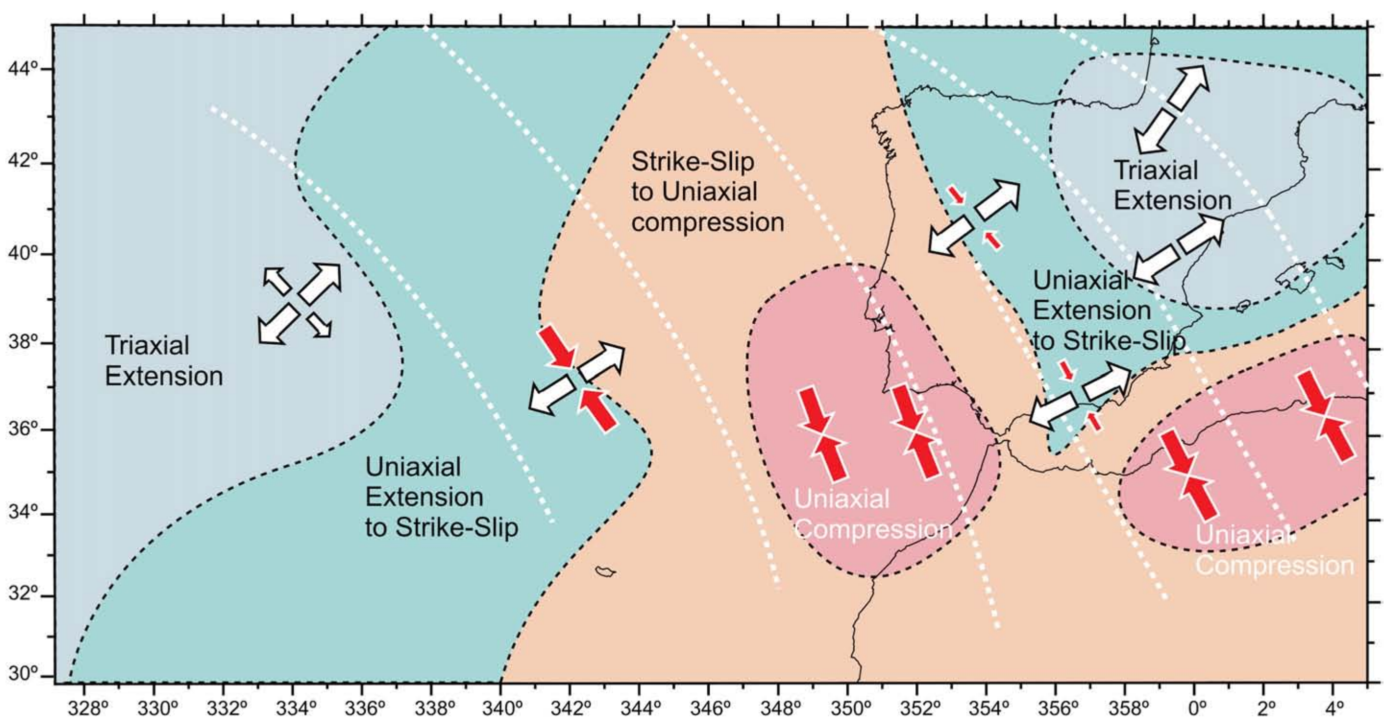

Fig. 15. The type of tectonic active stresses between the Mid-Atlantic Ridge and Algeria (De Vicente et al., 2008). Note the progressive increasing in the compressive character from the Gloria Fault up to the Gulf of Cadiz–SW Iberia; this tendency ceases at the Betics.

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favoured by the mechanical decoupling of Iberia and Africa. This is developed in two phases, the first is less important and is of an ultrapotassic nature (8.7–6.4 Ma), and the second is of an alkaline and ultra-alkaline nature (3.7–0.7 Ma) (Ancochea, 2004). These two volcanic crises are related to the more recent sedimentary ruptures registered in the Cenozoic basins, and would indicate geodynamic rather than climatic changes. The final outstanding fact is that the centres of volcanic emission are aligned NW–SE (Ancochea and Brändle, 1982), that is to say, parallel to the tectonic stress trajectories which initiated at that time and are similar to those of the present time.

6. Mechanical decoupling and active stresses

This new stress field, just described above, started in the Upper Miocene and it is similar to the present active stresses; therefore, it can be investigated from the point of view of instrumental seismicity. From active stress analysis (Ribeiro et al., 1996; Borges et al., 2001; Stich et al., 2005; De Vicente et al., 2005, 2007a, 2008), along the Eurasia–Africa western boundary, the type of stresses progressively change eastwards from triaxial normal faulting (extension) ($\sigma_z > \sigma_y > \sigma_x$, $\sigma_z = \sigma_1$) to axial thrusting (compression) ($\sigma_1 = \sigma_y$, $\sigma_z = \sigma_x$) along the Terceira Ridge, the Gloria Fault zone and the Gulf of Cadiz. This is

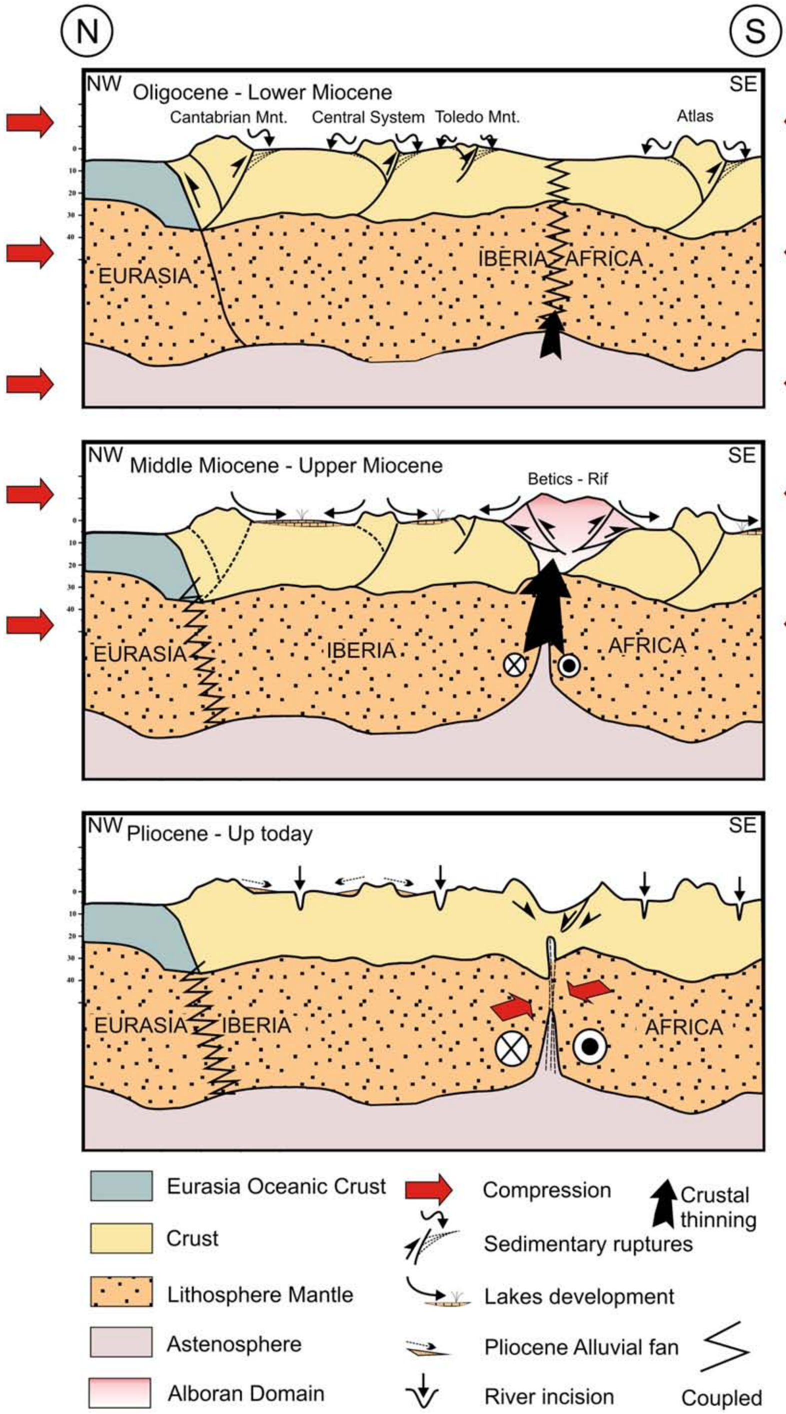

Fig. 16. Cross section sketches (not to scale) of the tectonic evolution of Eurasia–Iberia–Africa during the Cenozoic. Decoupling between Iberia and Africa is driven by the westwards displacement of the Alboran–Betics-Rif Block.

accompanied by a clock wise rotation of the S_{hmax} trend, from N137°E up to N162°E. Both tendencies are broken in the Betics–Alboran-Rif zone, where axial normal faulting (extension) ($\sigma_z = \sigma_y$, $\sigma_1 = \sigma_2$) predominates related to a N155°E S_{hmax}. In the Tell (N Algeria), axial compression reappears (N150°E S_{hmax}). In the Iberia foreland zones, the normal faulting stress (extension) increases from South to North and from West to East, so in the NE corner of the Iberian Peninsula, triaxial normal faulting (extension) appears, whereas the SW zone is close to axial compression (Fig. 15). The most westerly part of the Central System in Portugal undergoes active NE–SW thrusting. During the Pliocene, Central Ranges of similar trends, located eastwards, were also active (Guadalupe). It seems to be clear that, maintaining a NW–SE S_{hmax}, extension has migrated westwards since the Upper Miocene–Pliocene up to the current situation. This points to the idea that the most westerly parts of Iberia and Morocco, where the Alboran domain has not yet arrived, are still mechanically coupled. This process must have been simultaneous to the Betics extensional collapse which prevented the transmission of compressive stresses towards the foreland.

The slow convergence initiated at 9 Ma seems again to be very distributed but clearly not so generalised (especially eastwards Iberia, where normal faulting predominates, and not so evident in the Africa foreland) and causes the formation of the new Africa–Iberia frontier (Fig. 16).

7. Conclusions

A) During the Oligocene–Lower Miocene, and related to the blocking of the active plate border between Iberia–Africa and Eurasia (Pyrenees), the deformation was transmitted to the interior of the lower plate, the Iberian–African foreland, leading to most of the topographical features of the Iberian Peninsula and of the NW end of Africa. Thus, the Cantabrian–Pyrenean convergence would have been responsible for the regular structuring of basins and chains in a wide zone more than 3000 km long as far as the Anti-Atlas.

B) This wide zone of deformed foreland, very much distributed and regular, enables the generation of decoupled lithospheric–upper crustal folds.
The intraplate accommodation of this N–S shortening was taken in a series of folds by E–W to NE–SW crust buckling, connected by strike–slip fault corridors which also affected the crust.

C) The foreland where the stresses were applied had a large N–S to NW–SE Variscan grain and towards the E, was previously extended during the Mesozoic with NW–SE normal faults and, locally, E–W. Therefore, the resulting topographic pattern was more conditioned by the geometry of the limit of the active plates than by the previous structural grain. Thus, the folds in the Crust were nucleated in zones previously extended and on non-extended Variscan massif and even off-shore. This fact is clearly visible in the Cantabrian–Pyrenean belt and in the Central System-Iberian Chain. In this latter range, the inversion was incomplete as the main guidelines (NW–SE) were mainly reactivated as strike–slip faults.

D) The two main Cenozoic strike–slip fault corridors in Iberia are the Vilariça fault system to the W (left lateral) and the Iberian Chain to the E (right lateral). Both have numerous restraining bends, with small intra-mountain basins, terminating in contractional horse tail splays at their connections with the folds from the buckling of the crust (Central System). The whole set of macro-structures in Iberia can be explained through a generalised N–S shortening with constrictive conditions of the deformation. In the Iberian Chain, the previous E–W structures are those which best accommodate the inversion. In the Castilian Branch there was oblique inversion, where the main previous structures (NW–SE normal faults) acted as right lateral strike–slip faults. In many cases, the thrust structures were nucleated in accommodation zones and on relay ramps of the ancient Triassic rift. The folding and paleostress interference would be the result of local and forced accommodation of the N–S constriction conditions.

E) The progressive emplacement of the Betic–Alboran-Rif block towards the W mechanically decoupled Iberia from Africa, and the mechanical–deformational conditions which had favoured the development of lithospheric folds ceased. Only in the westernmost part (Portugal), and in the Gulf of Cadiz, the present conditions of deformation allow active thrusting. Therefore, the stress field presently active in the foreland is acting on a crust with a previously built relief. The simultaneous transition to exoreic of all the Iberian Cenozoic basins would be a delayed effect of the ceasing of the conditions for the formation of long wavelength folds.

Acknowledgements

The study was supported by Consolider Ingenio 2006 “Topoiberia” CSD2006-00041 and Spanish National Research Program CGL2006-13926-C02-01-02 “Topoiberia Foreland”. These projects are a part of the Spanish contribution to TOPO-EUROPE. The constructive review of Prof. Joao Cabral and an anonymous reviewer has helped greatly in improving the original version of the manuscript.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.tecto.2008.11.026.

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Please cite this article as: de Vicente, G., Vegas, R., Large-scale distributed deformation controlled topography along the western Africa–Eurasia limit: Tectonic constrains, Tectonophysics (2009), doi:10.1016/j.tecto.2008.11.026