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Review Article

Tethys–Atlantic interaction along the Iberia–Africa plate boundary: The Betic–Rif orogenic system

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ABSTRACT

Initial SE-dipping slow subduction of the Ligurian–Tethys lithosphere beneath Africa from Late Cretaceous to middle Oligocene twisting to a later faster E-dipping subduction of the subcrustal lithosphere is proposed as an efficient geodynamic mechanism to structure the arcuate Betic–Rif orogenic system. This new subduction-related geodynamic scenario is supported by a kinematic model constrained by well-dated plate reconstructions, tectonic, sedimentary and metamorphic data sets. The slow initial SE-dipping subduction of the Ligurian–Tethys realm beneath the Malaguide upper plate unit is sufficient to subduct Alpujarride and Nevado–Filabride rocks to few tens of kilometers of depth in middle Eocene times. The shift from SE- to E-dipping subduction during latest Oligocene–early Miocene was possibly caused by both the inherited geometry of the highly segmented Ligurian–Tethys domain and by the fast roll-back of the subducted lithospheric slab. The early Miocene rather synchronous multiple crustal and subcrustal processes comprising the collision along the Betic front, the exhumation of the HP/LT metamorphic complexes, the opening of the Alboran basin, its flooring by HP Alpujarride rocks and subsequent HT imprint, can be explained by the fast NW- and W-directed roll-back of the Ligurian–Tethys subcrustal lithospheric slab. The W retreat of the Ligurian–Tethys lithosphere in middle–late Miocene times could partly explain the initiation of its lateral tear and consequent subcrustal processes. From latest Miocene onward the Betic–Rif system evolved under both the northerly push of Africa resulting in tightening at crustal and subcrustal levels and by the distinct current dynamics of the steep lithospheric slab. The SW-directed scape of the Rif fold belt is one of the most striking evidences linked to the recent evolution of the squeezed Betic–Rif system between Africa and Iberia.

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1. Introduction

Assessing the detailed movements of tectonic micro-plates involves serious difficulties because of their ability to move independently from the large limiting plates and due to the large-scale vertical axis rotations they can experience. The Iberian plate located at the western boundary of the Tethys Ocean in Mesozoic and early-mid Tertiary times, is one of these puzzling micro-plates (e.g., [Vissers and Meijer, 2011](#)). The combined study of large scale plate motions together with geological observations constrains the evolution of the Iberian plate despite arguable areas and time periods that need to be refined (e.g., [Roest and Srivastava, 1991](#); [Rosenbaum et al., 2002a](#)).

In late Paleozoic times, after the Hercynian orogeny, the Iberian plate was part of Europe and Africa. Soon after the Triassic rifting episode (c. 250 million years ago [Ma]), propagation of the Central Atlantic Ocean toward the north produced the eastward movement of Africa relative to Iberia–Europe, opening the transtensional Atlas and Betic–Rif basins in early and middle Jurassic times, respectively. The transtensional corridor along the future Betic–Rif segment developed during ~20 million years (m.y.) connecting the Atlantic to the Ligurian–Tethys oceanic domains (Fig. 1A). This corridor was characterized by segmented domains possibly composed of thinned continental crust with strong evidences of upper mantle exhumation and oceanization (e.g., [Jiménez-Munt et al., 2010](#); [Puga et al., 2011](#); [Schettino and Turco, 2010](#)). In early Cretaceous times, northward propagation of the Atlantic Ocean spreading along the western margin of Iberia produced the abandonment of the Ligurian–Tethys corridor and the opening of the Bay of Biscay along the already rifted Pyrenean basin with a concurrent counter clockwise rotation of Iberia (e.g., [Gong et al., 2009](#); [Vissers and Meijer, 2011](#)) (Fig. 1B). Upper mantle rocks were exhumed along the westernmost side of the Pyrenees (e.g., [Jammes et al., 2010](#); [Lagabrielle et al., 2010](#)). Northward propagation of the North Atlantic led to the abandonment of the Bay of Biscay–Pyrenean opening near the early–Late Cretaceous boundary. The Iberian plate acted as an independent micro-plate until Late Cretaceous (around Santonian times) when Africa started to move north and northwestward against Eurasia. The northern motion of Africa squeezed the Iberian plate, producing the Pyrenees orogenic chain along its northern margin and the Betic–Rif orogen along its southern boundary (Fig. 1C). Most of the Pyrenean shortening was completed by middle Oligocene times and from this time onward the convergence of Africa was mostly accommodated across the Betic–Rif orogenic system and within the Iberian plate (Fig. 1D). The interior of the Iberian plate was deformed, inverting all previously rifted regions and producing intraplate mountain ranges (e.g., [Casas-Sainz and de Vicente, 2009](#); [Casas-Sainz and Faccenna, 2001](#); [Gibbons and Moreno, 2002](#); [Vera, 2004](#); [Vergés and Fernández, 2006](#)) (Fig. 2).

The Pyrenees define a classical rectilinear and asymmetric doubly verging orogen with an antiformal stack of low-grade metamorphic Paleozoic to middle Triassic rocks in the inner part of the belt separating two thin-skinned cover thrust systems on each side of it, which are detached above upper Triassic evaporites (Fig. 3). This orogenic system was formed by the partial subduction of the Iberian crust beneath the European crust producing limited shortening of about

~150 km (e.g., [Choukroune and E.C.O.R.S. Team, 1989](#); [Muñoz, 1992](#); [Roure et al., 1989](#)). As a consequence of this Iberian continental underthrusting, the Ebro and Aquitaine foreland basins were formed by the downward flexure of the Iberian and European plates, respectively. The combination of limited shortening and the endorheic (closed) post-tectonic evolution of the Pyrenees up to early late Miocene caused the preservation of most of the syntectonic deposits providing an accurate time control for the Pyrenean fold belt evolution (e.g., [Puigdefàbregas and Souquet, 1986](#); [Vergés et al., 2002](#)).

On the contrary, the Betic–Rif orogen shows an arcuate geometry, which is the result of a more intricate geological history, comprising an outer zone formed by the Guadalquivir and Gharb (Rharb) foreland basins, the Gulf of Cadiz accretionary complex, the External Betic and Rif thrust systems and the Internal Betic and Rif HP metamorphic complexes (Fig. 4). The hinterland of this arcuate compressive belt is cut by the extensional system related to the back-arc Alboran basin development.

Despite the large amount of literature published on this region since the first explicative model by [Andrieux et al. \(1971\)](#), the complexity of the Betic–Rif orogenic system remains. At a large scale there is a general consensus that the Betic–Rif orogen was formed as a consequence of the southwest slab roll-back of lithosphere subducting beneath the large and over-thickened Alcapeca tectonic domain ([Bouillin et al., 1986](#)). This domain, originally located south of the present Balearic domain, was dismembered in several blocks (Alboran–Kabyliés–Peloritán–Calabria), which drifted radially, producing the closure of the Ligurian–Tethys oceanic domains (e.g., [Michard et al., 2002](#); [Ricou, 1994](#)). In this context, the Alboran block would have traveled to the SW and then W with respect to Iberia to finally collide with the SE margin of Iberia and with the NW margin of Africa to form the Betic–Gibraltar–Rif fold-and-thrust belt (e.g., [García-Dueñas et al., 1992](#); [Lonergan and White, 1997](#); [Royden, 1993](#)) (Fig. 2). Within this large scale picture, however, the geodynamic mechanisms that shaped the Betic–Rif system and governed its evolution are not well understood (e.g., [Calvert et al., 2000](#)) and interpretations might be based on mutually exclusive geodynamic models for the study area as compiled and discussed by [Doblas et al. \(2007\)](#).

These models can be grouped in two categories depending on whether they are related or not to subduction processes (e.g., [Calvert et al., 2000](#); [Doblas et al., 2007](#); [Michard et al., 2002](#); [Torne et al., 2000](#)). Models unrelated to subduction suggest the formation of a 50–60-km thick orogenic belt located between Africa and Iberia that subsequently collapsed radially forming the Betic–Rif thrust system as well as the exposures of HP metamorphic rocks along the Internal Units (e.g., [Kornprobst and Vielzeuf, 1984](#); [Platt and Vissers, 1989](#)). Models involving subduction are not unique in proposing a variety of subduction polarities: N-dipping, E-dipping and double subduction slabs (dipping to SE in Paleogene times and to the NW in Neogene times) (e.g., [Doblas et al., 2007](#); [Michard et al., 2002](#); and references herein). Most of the proposed subduction models include association with slab roll-back, delamination, slab break-off or slab roll-back and lithospheric tearing (e.g., [Calvert et al., 2000](#); [Faccenna et al., 2004](#); [García-Castellanos and Villaseñor, 2011](#); [Gueguen et al., 1998](#); [Gutscher et al., 2002](#); [Spakman and Wortel, 2004](#)).

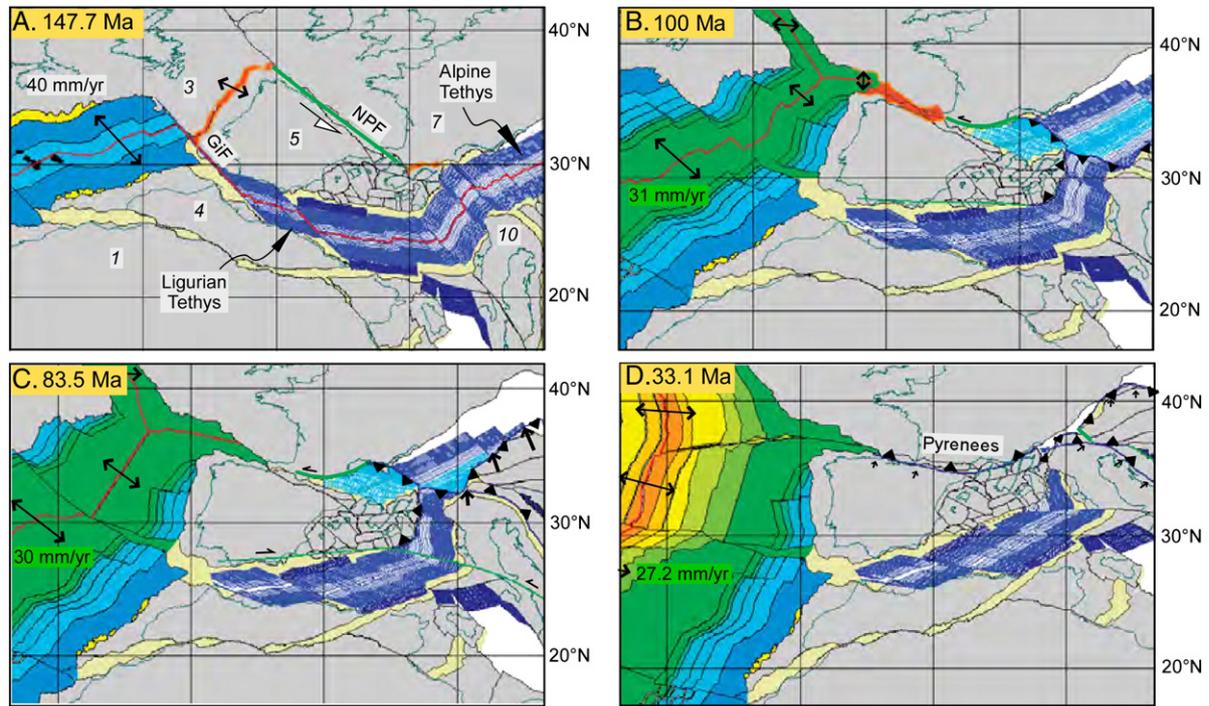


Fig. 1. Plate tectonic maps showing the evolution of Africa, Iberia and Europe from the late Jurassic to the middle Oligocene times (from Schettino and Turco, 2010). 1. NW Africa, 3. North America, 4. Morocco, 5. Iberia, 7. Eurasia, 10. Adria, GIF Gibraltar Fault, and NPF North Pyrenean Fault.

The main aim of this paper is to propose a new kinematic model integrated in the large-scale geodynamic framework of the Western Mediterranean region, and taking into account the large geological and geophysical datasets existing for the study region. To this end, we include an overview compilation of previous works related to the tectonic, sedimentary and volcanic evolutions together with

geophysics of the Betic–Gibraltar–Rif system. The presented kinematic model is based on the following considerations: i) the reconstruction starts in Late Cretaceous times at the onset of northern Africa convergence, ii) displacements and initial configuration are based on plate reconstructions of the Atlantic–Ligurian–Tethys region, iii) the model assumes that subduction polarity changes

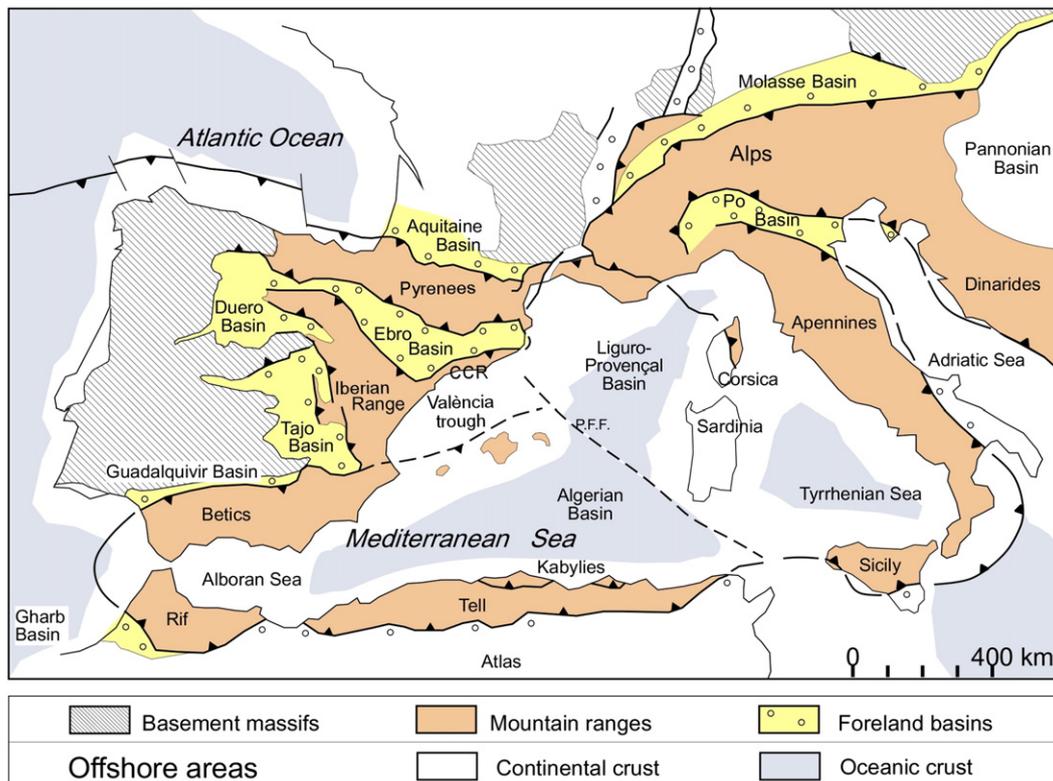


Fig. 2. Tectonic map of the western Mediterranean showing the main orogenic belts and foreland basins (Vergés and Sàbat, 1999). P.F.F., Pierre Fallot Fault.

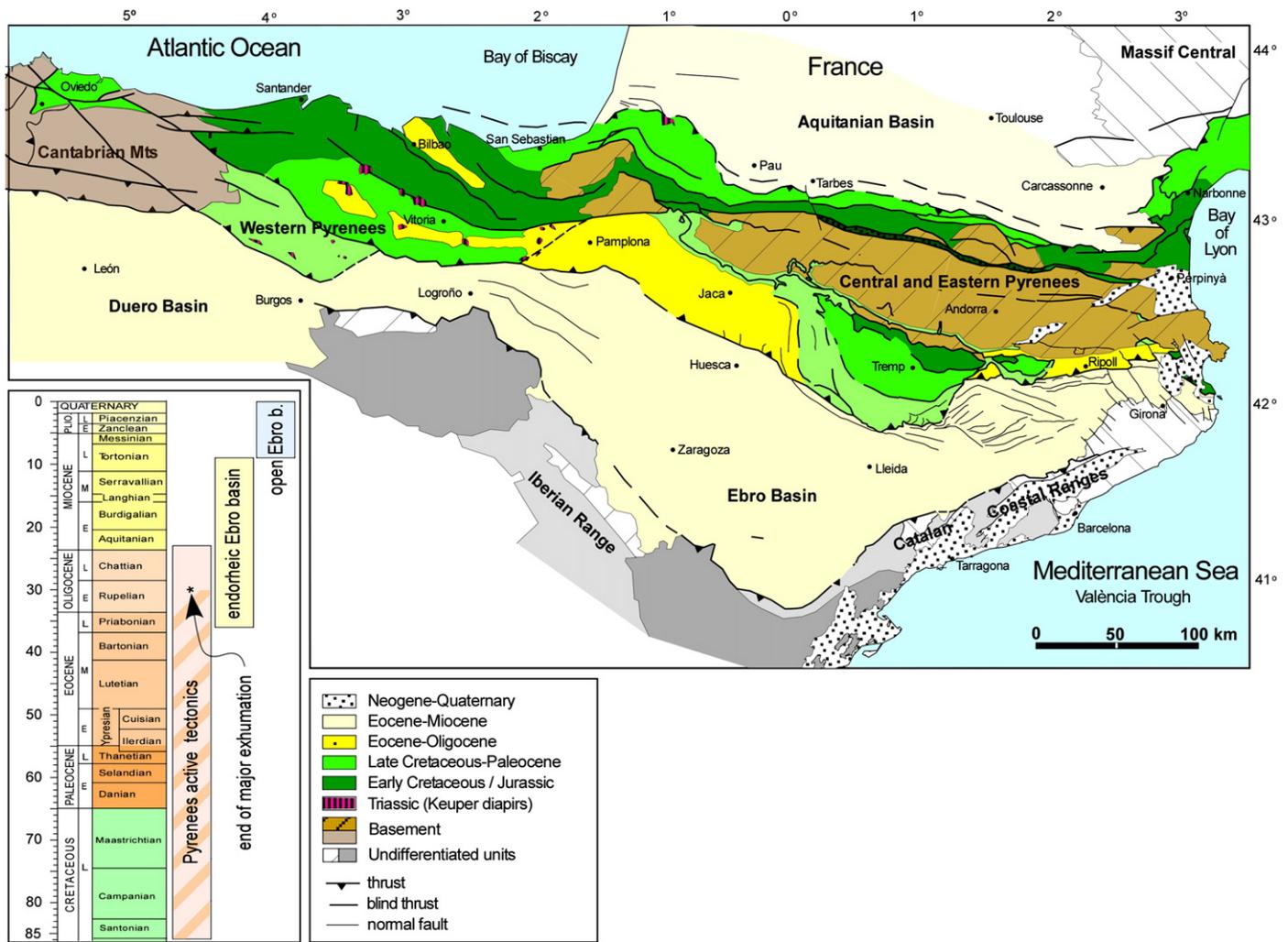


Fig. 3. Map of the Pyrenees showing the main tectonic units and time diagram highlighting the principal periods of mountain chain building and adjacent sedimentary basin evolution.

laterally from NW dipping in the Algerian segment to SE dipping in the Betic–Rif segment, and iv) the last stages must fit with the crust and upper mantle structure imaged by recent geophysical data and modeling.

2. The Pyrenean orogenic system

The northern and southern boundaries of the Iberian plate were deformed during the northern convergence of Africa against Europe, building up the Pyrenees and the Betic–Rif orogenic systems, respectively. Typically, deformation in these two systems has been assumed as diachronous with Late Cretaceous to mid Oligocene shortening in the Pyrenees and late Oligocene–early Miocene onward in the Betic–Rif orogen. Synchronously, the interior of the Iberian plate suffered inversion of the major Mesozoic rift basins, in a similar manner as regions away from the Iberian plate such as Central Europe (e.g., Kley and Voigt, 2008) (Fig. 2). To better understand how the convergence between Africa and Iberia was accommodated since Late Cretaceous to Recent times, we include a short summary of the evolution of the Pyrenees orogenic system. In this section we review the well-documented timing of the Pyrenean shortening with special focus on the proposed time for its cessation, which has a strong influence on the early stages of the Betic–Rif orogenic growth. The shortening between Iberia and Europe was accommodated by the limited north-dipping subduction of the Iberian

plate beneath Europe giving rise to the Pyrenean orogen (Choukroune and E.C.O.R.S. Team, 1989; Muñoz, 1992; Roure et al., 1989) (Fig. 3). This doubly verging orogen comprises from north to south the following domains: a) the Aquitaine retroforeland basin, related to the northern Pyrenean wedge (e.g., Biteau et al., 2006); b) the north-directed North Pyrenean thrust system; c) the Axial Zone of the chain, formed by Paleozoic rocks and a lower-middle Triassic cover tectonically stacked in three main south-directed thrust sheets; d) the south-directed and larger South Pyrenean thrust system; and e) the Ebro foreland basin associated to the southern Pyrenean wedge (e.g., Muñoz, 1992; Muñoz et al., 1986; Vergés et al., 1995). The well exposed and well preserved south Pyrenean thrust systems and adjacent foreland basin allow dating many of the individual growth structures, using the classical biostratigraphic and magnetostratigraphic techniques. The abundance of dated structures allowed deciphering the evolution of this relatively small orogenic system (Vergés et al., 2002). On the other hand, application of thermochronological dating in both basement rocks and clastic Eocene–Oligocene deposits along the front of the Axial Zone allowed estimating the timing of exhumation along the hinterland of the Pyrenees. The combination of these two databases constitutes the basis to unravel the entire Pyrenean deformation history.

The development of the Pyrenean continental collision thrust belt started during the Late Cretaceous by the tectonic inversion of previous Mesozoic extensional grabens around Santonian times or

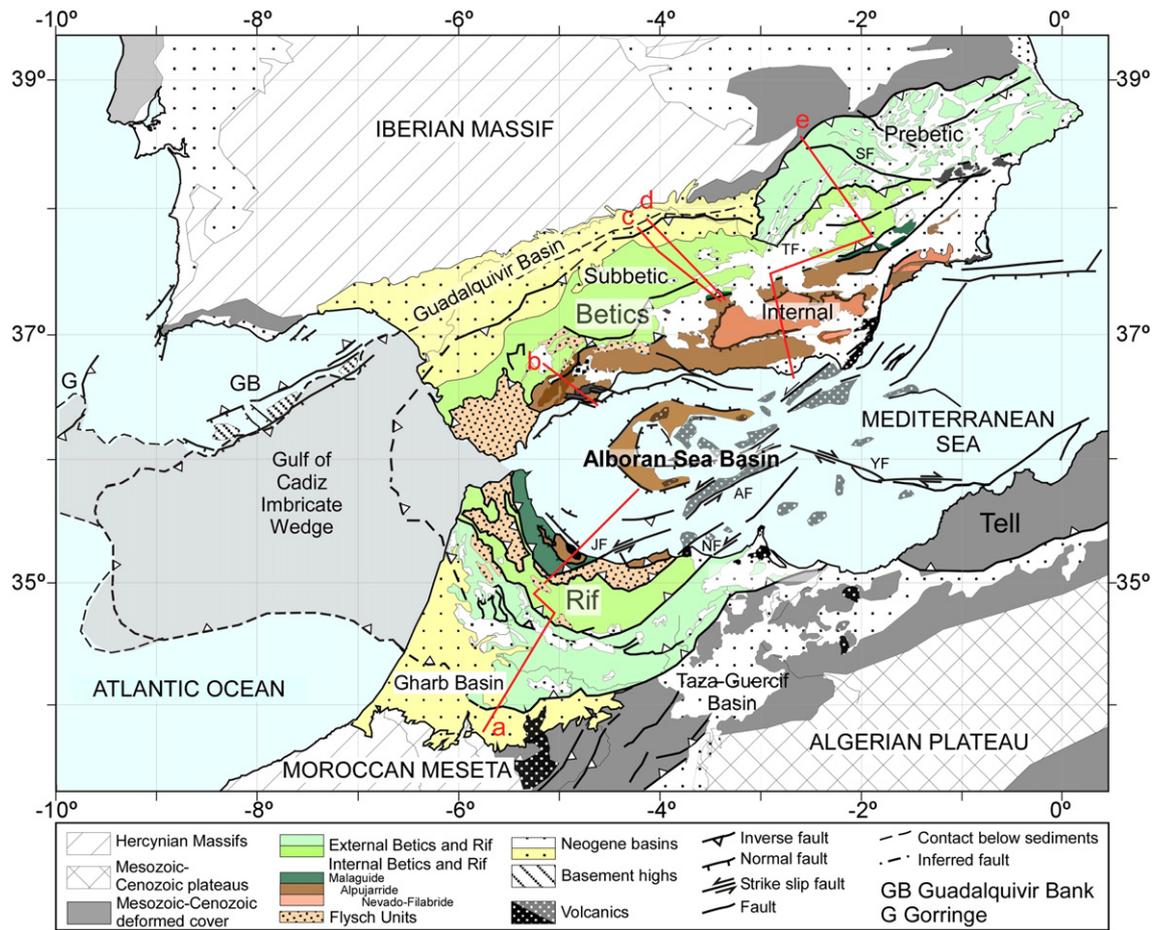


Fig. 4. Map of the Betic–Rif system showing the principal tectonic units and associated Neogene basins (from Iribarren et al. 2007). Red lines show the location of five cross-sections discussed in text: a. Michard et al. (2002); b. Mazzoli and Martín-Algarra (2011); c. Platt et al. (2003); d. García-Hernández et al. (1980); and e) Banks and Warburton (1991). SF. Socovos Fault (Betics); TF. Tiscar Fault (Betics); JF. Jehba Fault (Rif); NF. Nekor Fault (Rif); AF. Alboran Ridge Fault (Alboran); and YF. Yusuf Fault (Alboran).

somewhat earlier (e.g., Garrido-Megías and Ríos, 1972; Puigdefàbregas and Souquet, 1986; Souquet and Déramond, 1989; Specht et al., 1991) (Fig. 3). The Ebro basin, the south Pyrenean flexural foreland basin, was filled with marine sediments up to the early late-Eocene, corresponding to the near top of the Cardona evaporitic level (~36 Ma; Costa et al., 2010). The end of marine deposition is related to the uplift of the western Pyrenees, closing the connection of the south Pyrenean foreland basin with the Atlantic Ocean (e.g., Burbank et al., 1992; Puigdefàbregas et al., 1992; Riba et al., 1983; Vergés and Burbank, 1996; Vergés et al., 1995). The Ebro basin became closed, limited by the Pyrenees, the Catalan Coastal Ranges and the Iberian Range, and overfilled by non-marine clastic deposits up to its opening to the Mediterranean basin in the late Miocene (e.g., Vergés et al., 2002). Deposition coeval to shortening deformation continued at least to the upper Oligocene times (~24.7 Ma) as determined by Meigs et al. (1996) in the southeastern tipline of the Pyrenean thrust sheets front. However, the amount of deformation recorded by these syntectonic deposits is relatively small to fit thermochronology data, thus suggesting a slightly older age for the end of the major tectonic exhumation. The endorheic history of the Ebro basin is characterized by an internal fluvial network that radially delivered sediments to a large central lake (e.g., Anadón et al., 1979; Arenas and Pardo, 1999). During late Miocene times (e.g., Coney et al., 1996), but well before the Messinian (e.g., Arche et al., 2010; García-Castellanos et al., 2003), the endorheic Ebro fluvial system opened toward the Mediterranean Sea.

From the first pioneering studies of Morris et al. (1998) and Fitzgerald et al. (1999) along the central Pyrenees there has been a

great profusion of scientific results using different thermochronological techniques to decipher the exhumation history of the Pyrenean orogen. Results from apatite fission track (AFT) thermochronology show 1.4 to 3.8 km of orogenic exhumation between 50 and 35 Ma followed by 8.2 to 11.6 km of fast exhumation between ~35 and 32 Ma (latest Eocene and earliest Oligocene). These results were refined by using apatite (U–Th)/He and fission track thermochronometry, indicating that fast rates of exhumation ended at around 30 Ma, but somewhat continued up to the earliest Miocene (Gibson et al., 2007) in the central-western Pyrenees (Jolivet et al., 2007). New apatite (U–Th)/He (AHe) low temperature data combined with previous apatite fission track (AFT) thermochronology of the Maladeta pluton shows again that rapid cooling started at 30–35 Ma but slowed down abruptly at 25–30 Ma followed by low velocity exhumation until ~15 Ma (Metcalfe et al., 2009). Combined magnetostratigraphy and detrital AFT dating on the middle Eocene (~42 Ma) to early Oligocene (25 Ma) La Pobla conglomerates show a poorly defined event at 70–60 Ma followed by a slow cooling event at 50–40 Ma and a faster exhumation period at 30–25 Ma (Beamud et al., 2011). Their results also confirm the late Miocene exhumation event after 10 Ma was linked to the opening of the Ebro basin to the Mediterranean (Beamud et al., 2011). Inverse thermo-kinematic modeling of low-temperature thermochronology data provides new results for the rapid exhumation period (>2.5 km/m.y.) between 37 and 30 Ma followed by much lower rates (0.02 km/m.y.) (Fillon and van der Beek, 2012). To complete this short summary of thermochronology results, the analysis on ages of illite along thrust planes also show similar and complementary results (Rahl et al., 2011). These authors confirm the

age for the Boixols thrust at ~72 Ma during Campanian times and the Gavarnie thrust at ~69 Ma, during Maastrichtian times, which shows the Late Cretaceous displacement age for these early thrusts (Rahl et al., 2011). A renewed motion for the Gavarnie thrust, during the latest Oligocene, has been identified (fault gouge of ~32 Ma) as well as younger ages of thrust displacement along the Llavorsi thrust (~24 Ma; Rahl et al., 2011). The illite ages for these thrust displacements corroborate their proposed ages based on growth strata dating (e.g., Puigdefàbregas et al., 1992; Teixell, 1996) (Fig. 3).

The results discussed above confirm that the major period of orogenic growth of the Pyrenees occurred from Late Cretaceous to Oligocene times with a rapid exhumation period roughly between 37–35 and 32–30 Ma (early Oligocene). The rapid exhumation is related to the emplacement of lower basement thrust sheets beneath the antiformal stack corresponding to the Axial Zone of the chain (e.g., Beaumont et al., 2000). Younger exhumation ages are determined for the latest Oligocene in good agreement with growth strata ages in the foreland basin (e.g., Beamud et al., 2011; Meigs et al., 1996). The end of major Pyrenean orogenic activity along its central part in mid Oligocene times coincides with the time at which most of the interpretations initiate the development of the Betic–Rif system.

3. The Betic–Rif orogenic system

The Betic–Rif orogenic system is part of a larger-scale mountain belt that developed coevally to the opening of the western Mediterranean and is formed by: i) the Apennines, showing a NW–SE structural trend and NNE-vergence; ii) the Tellian fold and thrust, with an overall SSE-directed tectonic transport; and iii) the Betic–Rif system, showing a fairly regular NNW direction of tectonic transport in the External Betics, turning around the Gibraltar Strait to mostly SW-directed thrusting in the External Rif (Figs. 2 and 4). In this section we will focus on the structure and kinematics of the different tectonic domains forming the Betic–Rif system.

3.1. Structure of the Betic–Rif orogenic system

The arcuate Betic–Rif orogenic system shows five major tectonic domains (Fig. 4), which from external to internal are: a) the Guadalquivir and Gharb foreland basins and their corresponding continuation into the Gulf of Cadiz; b) the External Betics, separated in Prebetic and Subbetic showing ENE–WSW trend, and the External Rif with NNW–SSE trend; c) the Gibraltar Flysch Units around the western side of the thrust belt; d) the HP/LT metamorphic complexes of the Internal Betics and Internal Rif; and e) the Alboran back-arc basin occupying the innermost portion of the orogenic system. The Rif segment of the orogen shows an abrupt lateral termination against the Jebha and Nekor faults, which coincide with the SW continuation of the left-lateral Alboran Ridge fault zone (Martínez-García et al., 2011; Platt et al., 2003a) also called the Trans-Alboran shear zone (Stich et al., 2006).

The Guadalquivir and Gharb basins evolved as foreland basins during the Miocene downward flexure of the basement in response to the loading by the thrust and imbricated External Units (Fernàndez et al., 1998; García-Castellanos, 2002; Sanz de Galdeano and Vera, 1992; Zouhri et al., 2001) (Fig. 4). To the west, tectonic units of the Betic–Rif orogen emplaced since the early to the late Miocene, invading the Atlantic Margin region that includes the Gulf of Cadiz and the NW Moroccan margin, and forming the so-called Gulf of Cadiz Imbricate Wedge (Gràcia et al., 2003; Gutscher et al., 2002; Iribarren et al., 2007; Maldonado and Nelson, 1999). The front of the central–western part of the External Betics is formed by Triassic evaporites, unconformably overlain by an incomplete and discontinuous succession of Mesozoic to Neogene deposits. The External Betics form an intricate system of tectonic thrust imbricates and chaotic bodies emplaced in a marine depocenter,

which includes numerous olistoliths and olistostromes defining a large deformed tectono-sedimentary unit (e.g., Azañón et al., 2002).

The External Betics are mostly formed by two large paleogeographic domains with different stratigraphies attained during the Jurassic opening of the Ligurian–Tethys corridor (e.g., García-Hernández et al., 1980; Gibbons and Moreno, 2002; Vera, 1988, 2004; Vilas et al., 2003) (Fig. 4). The proximal and shallower Prebetic domain was located along the south-eastern margin of the Iberian plate while the deeper Subbetic domain was located farther SSE above thinner crustal regions corresponding to the north-western parts of the Ligurian–Tethys domain. Local submarine Jurassic and early Cretaceous volcanic and subvolcanic mafic rocks are present in the deeper sedimentary domains of the Subbetic (e.g., Vera et al., 1997). A second period of rifting during the early Cretaceous gave rise to SSE-dipping low-angle faults and the formation of half grabens within the eastern side of the Prebetic domain (e.g., Vilas et al., 2003). Below the Jurassic rocks, the upper Triassic evaporites constitute the present detachment level of the Betic fold-and-thrust system. The Late Cretaceous to Tertiary sedimentary successions display the progressive changes related to the passage from passive to the active margin at the onset of Africa convergence (e.g., Vera, 1988, 2000).

The Gulf of Cadiz Imbricate Wedge is a thick (> 11 km in its internal and eastern part) and a tongue-shape tectonic unit that occupies a large area of the Atlantic margin in both NW Africa and SW Iberia (e.g., Gutscher et al., 2002; Iribarren et al., 2007) (Fig. 4). Existing data from this accretionary wedge did not show a clear tectonic grain (Flinch et al., 1996; Frizon de Lamotte et al., 2008; Gutscher et al., 2002; Zizi, 1996) although recent swath bathymetry of the seafloor and multichannel and high-resolution seismic data highlight coherent NW, W and SW vergent anticlinal folds (Gutscher et al., 2009a). Gutscher et al. (2002) defined the Gibraltar accretionary wedge as active above a narrow east-dipping subducting oceanic slab. The principal deformation structures look fossilized by a thick deposition in late Tortonian times along the lateral boundaries of the system (Gràcia et al., 2003; Iribarren et al., 2007; Maldonado and Nelson, 1999; Zitellini et al., 2009) although high resolution seismic data suggests active deformation of the accretionary wedge toward the more internal boundaries of this accretionary wedge (Gutscher et al., 2009b). The late Tortonian fossilization of the front of the Betic–Rif fold-and-thrust belt is also observed in seismic lines crossing the External Betic Units in the Guadalquivir basin (e.g., Berástegui et al., 1998). The post-Tortonian deformation observed in localized areas of the Gulf of Cadiz Imbricate Wedge (Gutscher et al., 2002, 2009b), however, suggests that this region is still active. Recent neotectonic numerical modeling mostly shows NE–SW thrust faults and WNW–ESE right-lateral strike-slip faults in the Gulf of Cadiz in agreement with GPS data (Cunha et al., 2012).

The Gibraltar Flysch units of the Betic–Rif orogenic system are only cropping out along its western side, from near Malaga in Spain to Alhucemas in Morocco (Fig. 4). The Gibraltar Flysch units are formed by siliciclastic rocks ranging from upper Jurassic to early–late Burdigalian age (e.g., de Capoa et al., 2007; Luján et al., 2006; Rodríguez-Fernández et al., 1999; Serrano, 1979). The Gibraltar Flysch units are presently deformed as an accretionary prism tectonically sandwiched above the External Subbetic fold-and-thrust system and below the HP metamorphic rocks of the Internal Betics (e.g., Bonardi et al., 2003; Crespo-Blanc and Campos, 2001). The Flysch units along the Rif segment of the orogen show equivalent tectonic relationships (e.g., Michard et al., 2006). The siliciclastic foredeep turbidites of the Flysch units (Mauretianian) were supplied by the stacked Internal Units of the Betic–Rif system since the earliest Miocene and up to the late Burdigalian (de Capoa et al., 2007). Although the folding of these turbiditic systems occurred after their late Burdigalian deposition (e.g., Balanyá et al., 1997; Crespo-Blanc and Campos, 2001) the strong association with uplift and erosion of the Internal Betic–Rif units possibly records the onset of collision. Both the front of the Subbetic fold-and-thrust belt and the front of the

Gibraltar Flysch Unit were fossilized by Langhian sediments (Gràcia et al., 2003).

The Internal Betics are constituted by 3 main tectono-metamorphic complexes that are tectonically stacked, from bottom to top they are: the Nevado-Filabride, the Alpujarride and the Malaguide (Fig. 4). Although we mainly describe the Betic segment of the orogen, the Rif segment shows very similar characteristics (e.g., Michard et al., 2002). The Nevado-Filabride Complex is composed of 3 main tectono-metamorphic units affected by high-pressure and low-temperature metamorphism under eclogite and/or blueschists facies (e.g., Martínez-Martínez et al., 2002). Their metamorphic conditions vary from 1.2 to 1.6 GPa and 320°–450 °C, corresponding to pressure depths of ~35–42 km, to eclogitic conditions of 2.0–2.2 GPa and 650°–700 °C resultant from depths of ~60–65 km (e.g., Augier et al., 2005; López Sánchez-Vizcaíno et al., 2001; Puga et al., 2000). The contacts between different units within the Nevado-Filabride Complex are gently dipping ductile-shear zones with westward sense of shear (e.g., García-Dueñas et al., 1988; Martínez-Martínez et al., 2002). The Alpujarride Complex crops out extensively in both the Betic–Rif fold belt and the Alboran basin. It is composed of several 2 to 5 km-thick units of Paleozoic to middle-upper Triassic rocks showing strong differences in the metamorphic gradient across the contacts between units. Their pressure–temperature trajectories indicate a single cycle of burial–exhumation with high-pressure metamorphic peaks ranging from 0.7 GPa and 340 °C (~21 km depth) to 1.1 GPa and 570 °C (~33 km depth) (e.g., Azañón and Crespo-Blanc, 2000), followed by a rapid isothermal retrograde decompression (e.g., Comas et al., 1999; Rossetti et al., 2005). In the Internal Rif units the pressure–temperature trajectories are slightly higher, reaching ~1.6 GPa in pressure (e.g., Michard et al., 2006). In the western Betics and Rif, the Ronda and Beni Bousera mantle peridotites were emplaced between two structurally highly metamorphic Alpujarride units. The Malaguide unit is formed of low-grade metamorphic Paleozoic basement and a Permo-Triassic sedimentary cover, followed by a kilometer-thick discontinuous Jurassic to lower Miocene sedimentary succession with no metamorphism (e.g., Lonergan, 1993; Martín-Martín et al., 1997). This unit was intruded by Oligocene and lower Miocene volcanism (Turner et al., 1999). The contact with the underlying Alpujarride units represents a large normal fault although this apparent displacement can be interpreted as resulting from the Alpujarride emplacement as a tectonic wedge between the non-metamorphic Malaguide units above and the Nevado-Filabride below (e.g., Lonergan and Platt, 1995). The large variations in metamorphic conditions from one unit to the next as well as the abrupt jumps in metamorphic degrees across contacts propitiated different tectono-metamorphic interpretations concerning the evolution of the Internal Betic complexes. Nonetheless, there is an increasing consensus in relating the HP/LT metamorphic rocks cropping out in most of the orogenic systems surrounding the western Mediterranean to partial subduction, and linking their exhumation to the flow generated in the subduction channel facilitated by the creation of space caused by the slab roll-back (e.g., Jolivet et al., 2003).

The Alboran basin is a back-arc basin occupying the inner side of the Betic–Rif system and formed mainly during the early Miocene (e.g., Comas et al., 1992, 1999; García-Dueñas et al., 1992; Maldonado et al., 1992; Martínez-García et al., 2011; Platt and Vissers, 1989; Watts et al., 1993) (Fig. 4). The Western Alboran basin formed around 27–25 Ma and is floored by HP Alpujarride metamorphic basement units that were affected by high grade metamorphic conditions before late Oligocene–earliest Miocene times (Soto and Platt, 1999). These basement rocks share a common origin, history and timing with the exposed Alpujarride units in the Betics. The eastern Alboran basin appears to be younger than the Western Alboran basin forming at 12–10 Ma (Booth-Rea et al., 2007). The extensional evolution of the Alboran basin was accompanied by broad magmatism and several volcanic episodes (e.g., Doblas et al., 2007; Duggan et al., 2004; Lustrino et al., 2011). The extension in the Alboran basin was followed by a compressional

reorganization since the late Miocene to Holocene times (Chalouan et al., 1997; Comas et al., 1992, 1999; Watts et al., 1993). The tectonic processes occurring in the basin during the Pliocene and Quaternary times are to a large degree responsible for the present seafloor physiography of the basin (e.g., Ballesteros et al., 2008; Bourgois et al., 1992; Gràcia et al., 2006; Martínez-García et al., 2011; Woodside and Maldonado, 1992).

3.2. Kinematics of the Betic–Rif orogenic system

The kinematic data used in our reconstruction of the Betic–Rif fold belt take into account the folding and thrusting sequence based on balanced cross-sections and the shortening estimates across the orogen whereas the along-strike variations of the vertical axis rotations in the Betic–Rif system are discussed hereinafter.

Few regional balanced cross-sections have been constructed across the Betic–Rif orogen and only some of them have been restored. In this section we focus on the structural cross-cutting relationships of the different thrust sheets displayed in published regional transects across the Betic–Rif system including both the Internal and the External units. These time relationships between different thrust sheets of the Betic–Rif system are crucial for the kinematic model presented here (see location of cross-sections described in this section in Fig. 4). These cross-sections were constructed parallel to the tectonic transport direction for each segment of the Betic–Rif orogen based on the thrust-related slip vectors (e.g., Platt et al., 2003a) but also perpendicular to the folding trends.

A cross-section through the central Rif by Michard et al. (2002) shows the structural relationships between different tectonic units with large but unconstrained thrust displacements. From the foreland to the hinterland the Gharb foreland basin is overthrust by the External Rif units constituted by the Rides préifaines, the Prerif and the Mesorif, which in turn are overthrust by the Intrarif units. This imbricate system is thrust by the Maghrebien Flysch units that are overthrust by the Dorsale carbonate unit (thrust sheet). The Internal Rif units override all the previous Rif cover thrust sheets, each one corresponding to separated paleogeographic domains. These Internal Rif units are constituted from top to bottom by the less-metamorphic Ghomaride units (Malaguide in the Betics) and by the HP/LT metamorphic Sebtdites units (Alpujarride in the Betics), which include the tectonic slivers of Beni Bousera peridotites (Ronda peridotites in the Betics). A typical forelandward propagation of thrusting would imply a large-scale progressively younger age for the emplacement of the thrust sheets from the Internal Rif units to the External Rif units and foreland.

Several cross-sections illustrate the Betic segment of the Betic–Rif orogen (Fig. 4). Although they are coincident in most of the major tectonic trends, there are significant differences regarding the position of the Flysch unit and the contact between the External and the Internal Betics, as discussed below. The western Betic structure is illustrated along several NW–SE trending cross-sections in the area from the foreland to the hinterland and summarized in a conceptual cross-section (Fig. 5). According to these sections, the Guadalquivir Allochthon deposited on top of the carbonate Prebetic unit, infills the foreland basin of the Betics. This foreland basin is thrust by the External Subbetic unit (e.g., Berástegui et al., 1998), which is detached above the Triassic evaporites and internally folded and thrust (e.g., Crespo-Blanc, 2007). According to Crespo-Blanc and Campos (2001) and to Mazzoli and Martín-Algarra (2011), the Flysch units thrust over the Subbetic Unit but show an opposite structural position in Platt et al. (2003a). Nevertheless, all interpretations show the Subbetic thrust sheet on top of the Prebetic carbonate successions (e.g., Balanyá et al., 2007). The Nieves unit is formed by a Triassic to Oligocene carbonate succession thrust over the Flysch units but displays a large overturned syncline intensely metamorphosed in the footwall of the Ronda peridotites (e.g., Martín-Algarra, 1987; Mazzoli and Martín-Algarra, 2011). These peridotites are tectono-metamorphically

sandwiched within HP-LT metamorphic Alpujarride slivers piled up beneath the non-metamorphic Malaguide Complex. This complex is nevertheless capped by upper Oligocene–early Miocene breccia deposited after the Malaguide emplacement but before the tectonic wedging of the HP metamorphic slices beneath it (e.g., Martín-Algarra et al., 2000). The geometry of the central-eastern Betic fold-belt follows the same structure that was previously described for the central Rif and western Betics. Across this segment, the Prebetic unit thrusts over the foreland basin deposits of the Guadalquivir Allochthon and is in turn overthrust by the Subbetic unit as already documented in García-Hernández et al. (1980). The main difference between transect interpretations concerns the External–Internal boundary that Banks and Warburton (1991), Platt et al. (2003a) and Frizon de Lamotte et al. (2004) show with a different structure than that described above. According to these studies, the Subbetic unit thrusts over the External–Internal boundary on top of a basal thrust with a relatively large displacement and showing an internal, SE-directed imbricate back-thrust system. In these interpretations, the stack of Internal Betics formed a tectonic wedge transported toward the north-west above the Triassic evaporites and below the Subbetic units. The cross-sections across the Betic–Rif fold-belt show a rather systematic disposition of the different allochthonous tectonic units building up such a complicated thrust system.

Although not totally established, the syntectonic deposition during thrusting suggests a forelandward evolution in both the Betic and Rif orogenic segments (Fig. 5). Assuming this sequence of thrusting, the order of thrust sheet emplacement would correspond to the Malaguide–Ghomaride thrust sheet, the Flysch thrust sheet, the Subbetic thrust sheet and the Prebetic deformation, which may displace its front on top of foreland basin deposits. Departures from this scheme would be the emplacement of Alpujarride HP metamorphic rocks between the Malaguide on top and the Nieves unit in the western Betics as documented above. The Malaguide emplacement on top of the Nieves unit is fossilized by the deposition of upper Oligocene–early Miocene breccia in the western Betics (e.g., Martín-Algarra et al., 2000) and by early Oligocene conglomerates in the Sierra de la Espuña basin in the central–eastern Betics, which contain Malaguide clasts (e.g., Lonergan and Mange-Rajetzky, 1994; Serra-Kiel et al., 1996). These tectono-sedimentary relations indicate that the Internal Betics were thrust over the External Betics by early Oligocene times and thus right before the onset of the Alboran extension. The fossilization of the Guadalquivir Allochthon during the latest Tortonian times constrains the late major thrust displacement in the Guadalquivir foreland basin.

Shortening estimations for the External Betics vary from 64 to 85 km, according to García-Hernández et al. (1980) to about 100 km according to Banks and Warburton (1991) and to 196 ± 67 km proposed by Platt

et al. (2003a) once corrected for the true shortening direction along the Jaén cross-section. Thrusting and folding in the External Betics occurred on top of the main contractional detachment along the upper Triassic evaporitic level (Keuper), which also reactivated in diapiric structures during the Tertiary (e.g., Crespo-Blanc, 2007; García-Hernández et al., 1980; Luján et al., 2003; Moseley et al., 1981; Rondeel and Gaag, 1986). At the larger scale the upper Triassic evaporites represent the decoupling horizon separating the stratigraphies of the External and Internal units in both the Betic and the southern Rif segments, as already formulated by Wildi (1983) and recently supported by Chalouan et al. (2008).

4. Geological and geophysical key constraints on the kinematic model for the Betic–Rif orogenic system

The kinematic model across the central Betics proposed in this work is based on some key considerations that refer to: a) the large-scale plate reconstructions addressing the Africa–Iberia boundary during the Mesozoic and Cenozoic; b) the spatial distribution of HP/LT metamorphic complexes related to the Alboran and Algerian basins; c) the role of the mechanical stratigraphy separating the External from the Internal Betic–Rif units; and d) the timing for the onset of compressional deformation along the Betic–Rif system. In general, these considerations are based on geological and geophysical constraints, although some assumptions are made when data are not conclusive enough.

4.1. Large-scale plate reconstructions addressing the Africa–Iberia boundary during the Mesozoic and Cenozoic

All recent plate reconstructions indicate a strongly segmented Ligurian–Tethys oceanic basin at ~85 Ma, due to the transtensional motion of Africa relative to Iberia from Jurassic to Late Cretaceous (e.g., Handy et al., 2010; Schettino and Turco, 2010; Stampfli and Borel, 2002) (Fig. 1). All these reconstructions, presenting a fairly good fit between them (see Capitanio and Goes, 2006), display an average separation between Africa and Iberia mainlands of about 350–450 km in a NW–SE direction, including the ones proposed by Rosenbaum et al. (2002a). This calculated width includes the thinned continental margins and the transitional/oceanized crust domains. Nonetheless, the total E–W length of the Betic–Rif segments of the Ligurian–Tethys domain is longer than 700 km (depending on the proposed orientation of the narrow ocean basins), which fits perfectly with the proposed length (depth) of the lithospheric slab beneath Gibraltar. Due to the uncertainty of this Late Cretaceous arrangement we assume original WNW–ESE elongated segments separated by transform faults (Fig. 6). The initial

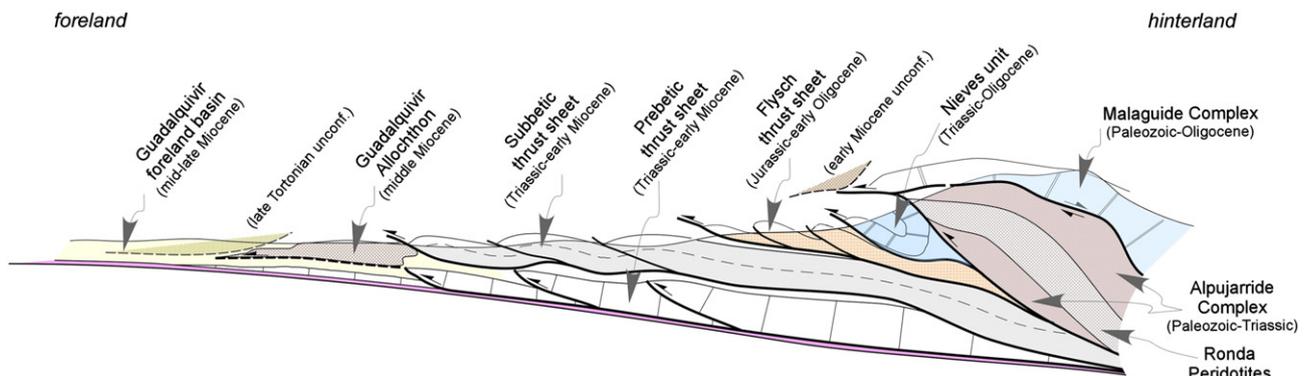


Fig. 5. Conceptual cross-section (not to scale) summarizing cross-cutting and temporal relationships between adjacent thrust sheets from the hinterland to the foreland of the Betic–Rif fold-and-thrust belt. This cross-section is based on the sections delineated in Fig. 4.

configuration of the Africa–Iberia plate boundary could consist of several small segments. These segments would be formed by very thin continental and/or transitional-to-oceanic crust and would be located between the Atlantic Ocean and the Ligurian Tethys. In our reconstruction we use four segments: the easternmost two corresponding to the Betic–Rif and the two western segments to the Gulf of Cadiz and Gorrige Bank. The Gulf of Cadiz and Gorrige segments should be displaced to the NW with respect to the Betic–Rif segment(s) in an echelon disposition as recently determined by Sallarès et al. (2011). The reconstructed length of the proposed Betic–Rif subduction front is of about 400 km, which is roughly equivalent to the present length of the metamorphic Internal Betics. The assumed initial plate-boundary configuration fits with recent plate tectonic reconstructions and with the regional characteristics of the Betic–Rif orogen. In this sense, the proposed kinematic model uses the potential imprint of the Jurassic plate boundary geometry on the Late Cretaceous–Late Miocene development of the Betic–Rif orogenic system. This inherited Jurassic extensional structure may still be studied in the External Betics where NW–SE trending faults with apparent dextral strike slip component (e.g., Socovos and Tiscar faults in Fig. 4; Ruiz Ortiz et al., 2006; Vera et al., 1984) could control the Mesozoic deposition and subsequently the compressional geometry of the fold-and-thrust belt during the Tertiary (e.g., Banks and Warburton, 1991; Platt et al., 2003a). Interestingly, these Prebetic and Subbetic strike slip faults share the subparallel direction and the sense of the lateral displacement of the transform faults separating the Ligurian–Tethys basins proposed in the published plate kinematic reconstructions and in the presented model (Fig. 6). Their original WNW–ESE trends would correspond to the transform faults bounding the segmented basins whereas the roughly NE–SW directions would correspond to the margins of the

Ligurian–Tethys transitional-oceanized crustal domains as discussed later.

4.2. Spatial distribution of HP/LT metamorphic complexes related to the Alboran and Algerian basins

Exposures of the high-pressure metamorphic complexes in the Betic–Rif system (Internal units) of Spain and Morocco and in the Tellian system (Kabylies units) of Algeria show an opposite map view distribution and an opposite tectonic polarity (Fig. 6). The eastern termination of the Internal Betics and the western end of the Great Kabylies can be connected through a line with NW–SE direction (paleo-transform fault), which would be parallel to the present NW–SE faults that compartmentalize the present northern margin of Africa from Tunisia to Morocco (Mauffret, 2007; Strzeczynski et al., 2010; Yelles-Chaouche et al., 2006). These faults could correspond to old Ligurian–Tethys transform faults. The Internal Betics are located to the SE of the NW-verging External Betics fold belt (e.g., Frizon de Lamotte et al., 1991) whereas the Kabylies are located to the NW of the SE-directed Tellian fold belt (e.g., Frizon de Lamotte et al., 2000). The External Betics constituted the SE Iberian margin depositional units and the Tellian and Atlas units that extend to the SE of the Kabylies composed the depositional units of the northern African margin. In this context, the Tellian units west of the Great Kabylie would form the northern margin of the large Atlas basin as well as the northern boundary of Africa (e.g., Frizon de Lamotte et al., 2011 and references herein).

The opposed symmetry of the Internal Betics and Kabylies in both their location and structural position within the orogenic systems may indicate that the Internal Betic complexes could be the

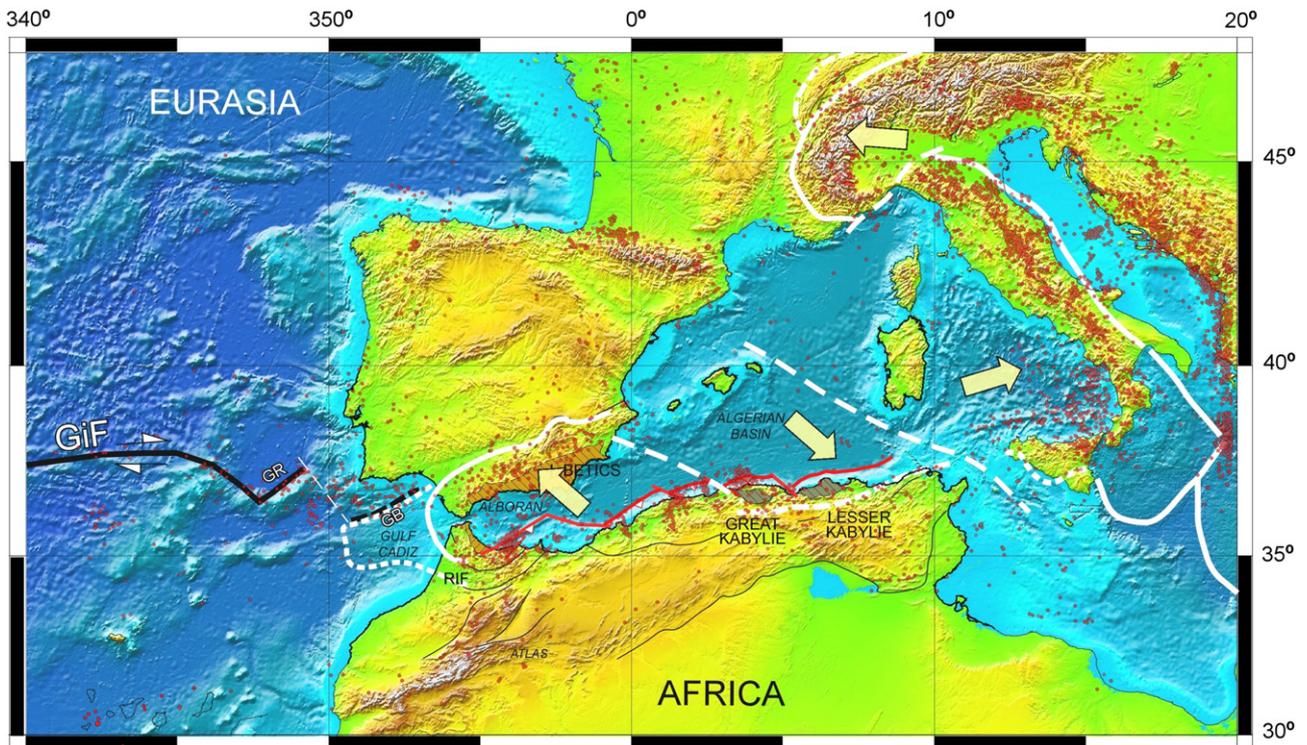


Fig. 6. Topographic map of the western Mediterranean with the location of earthquakes and the distribution of HP metamorphic complexes along the Betic–Rif system (Internal Betics and Internal Rif), and along the Tellian fold belt in northern Africa (Great and Lesser Kabylies). The Internal Betics and the Kabylies show a complementary position across the westernmost Mediterranean. Thick continuous white lines show the geometry of arcuate orogenic systems around the Mediterranean. Two opposed subducting domains are separated by active, inactive or paleo-transform faults (dashed white lines) dividing Alps from the Apennines and Kabylies from Betic–Rif systems. The tectonic inversion of the northern margin of Africa, marked by the red line, is based on Mauffret (2007) and Strzeczynski et al. (2010). The geometry of the Calabrian arc is based on Polonia et al. (2011). GIF: Gibraltar Fault; GR: Gorrige Ridge; and GB: Guadaluquivir Bank.

consequence of SE-dipping subduction processes, in the same way that the Kabylies has been unambiguously interpreted as resulting from the NW-dipping subduction of the Algerian segment of the Ligurian–Tethys ocean (e.g., Frizon de Lamotte et al., 2000; Gueguen et al., 1998; Rosenbaum et al., 2002a; Vergés and Sàbat, 1999) (Fig. 7). This proposed initial SE-dipping subduction along the Betic–Rif segment of the Ligurian–Tethys ocean, as already proposed for the Oligocene–Miocene periods (e.g., Booth-Rea et al., 2005), would represent a change in the subduction polarity across the inferred NW–SE transform fault separating the Algerian from the Betic–Rif segments of the larger Ligurian–Tethys oceanic realm. This interpretation shares some aspects with previously proposed models as Boccaletti and Guazzone (1974), Zeck (1999) and Gelabert et al. (2002). The SE-dipping subduction along the Betic–Rif segment triggering a NW-directed orogenic thrust system is also supported by the principal vergence of thrusting and folding in the External Betics and by the marked parallelism between the trend of these folds and thrusts and the direction of their boundary with the internal metamorphic units along the entire Betic–Rif domain (e.g., Frizon de Lamotte et al., 1991; Platt et al., 2003a).

4.3. Role of the mechanical stratigraphy in separating the external from the internal Betic–Rif units

Another key observation that has received little attention in the literature is the role of the mechanical stratigraphy and specifically, the role of the upper Triassic evaporites in shaping the Betic–Rif orogenic system (Fig. 4). The interface between the Internal and External Betic–Rif units shows a complete decoupling along the upper Triassic evaporites as is commonly observed in orogenic systems surrounding the Western Mediterranean region and containing these thick upper Triassic evaporites (e.g., Hudec and Jackson, 2007).

The oldest rocks of the tectonic nappes of the Prebetic and Subbetic units (External Betics) are upper Triassic evaporites (e.g., García-Hernández et al., 1980; Lanaja et al., 1987), which produced diapiric processes during the Jurassic and Cretaceous (Crespo-Blanc, 2007; Nieto et al., 1992) and later reactivations during the Tertiary (e.g., Moseley et al., 1981; Roca et al., 2006). In contrast, the stratigraphy of the Internal Betics consists of Paleozoic rocks up to middle–upper Triassic marbles although some evaporites may be present at least in one metamorphic tectonic sliver (e.g., Egeler and Simon, 1969; Puga et al., 2000). According to these observations we propose a plausible restoration in which the Internal Betic units would be stratigraphically and structurally located (at least partly) beneath the External Betic units, at the initiation of Africa–Iberia convergence in Late Cretaceous times. Consequently, during the northern displacement of Africa the External units were thrust and folded above the upper Triassic detachment level whereas the underlying upper crustal rocks and associated cover layers (lower and middle Triassic) were carried down into the subduction zone to depths of several tens of kilometers. These down-going units were exhumed and emplaced as a tectono-metamorphic wedge beneath the Malaguide Complex to form the stacked HP/LT metamorphic slices of the Alpujarride and Nevado-Filabride complexes.

4.4. Timing for the onset of the compressional deformation along the Betic–Rif system

The northern convergence between Africa and Europe started in Late Cretaceous times according to tectonic reconstructions (e.g., Dercourt et al., 1986; Dewey et al., 1989). Typical estimates of convergence between Africa and Eurasia from Cretaceous time to middle Eocene vary from ~500 km (Dewey et al., 1989) to ~100 km (Dercourt et al., 1986). Part of this convergence, up to ~150 km, took place along the Pyrenean orogen across the northern boundary of

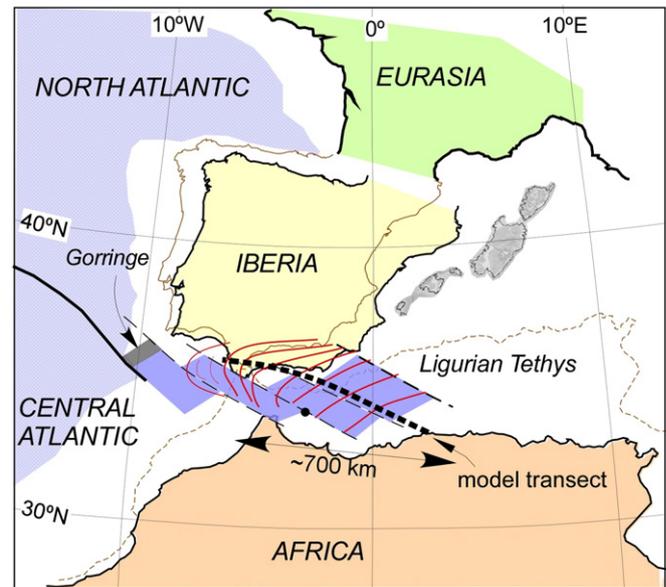


Fig. 7. Map reconstruction of the proposed Ligurian–Tethys domain between Iberia and Africa during Late Cretaceous before the onset of the African convergence. Most reconstructions of this area depict an irregular and segmented geometry for the Atlantic–Alpine–Tethys corridor. The dashed thick line shows the approximate trace of the section used to illustrate the proposed kinematic model depicted in Fig. 9. Thin continuous and dashed brown lines show the present positions of Iberia and Africa, respectively.

Iberia (e.g., Muñoz, 1992). After middle Eocene times, most of the models show similar convergence values of about 120–150 km (Dercourt et al., 1986; Dewey et al., 1989; Roest and Srivastava, 1991) that should be mostly accommodated along the Betic–Rif orogenic system (Vissers and Meijer, 2011). The total convergence between Africa and Iberia is estimated to be about 200 km from Late Cretaceous to late Tortonian and 50 km more after this time period (e.g., Dewey et al., 1989; Mazzoli and Helman, 1994). Similar values of post-Tortonian shortening were estimated from crustal thickness reconstructions (Iribarren et al., 2009). Despite the significant number of Late Cretaceous and Palaeogene indications of active compression tectonics along the eastern Betics the majority of the proposed geodynamic models disregard the pre-Oligocene history.

Compressive tectonic pulses modified the paleogeography of the Prebetic and Subbetic domains causing the restructuring of carbonate platforms, the onset of mixed carbonate–siliciclastic platforms, the deposition of debris flows, the formation of SW–NE troughs and the inversion of extensional faults (e.g., Martín-Chivelet and Chacón, 2007; Reicherter and Pletsch, 2000; Vilas et al., 2003) (Fig. 8). These events were not widespread at the basin scale but affected individual sub-basins starting around Coniacian–Santonian (~85 Ma) and lasting to late Campanian–Maastrichtian and Paleocene–early Eocene periods. An older event back in the middle–late Cenomanian has been also related to transpressive movements along the southern margin of Iberia (e.g., de Jong, 1990) but probably refers to the pre-compression history of the Ligurian–Tethys basin (Fig. 8). To account for these compressive events taking place in the most external part of the SE Iberian margin we assume that Africa–Eurasia convergence initiated already in the Betic–Rif segment of the Ligurian–Tethys since Late Cretaceous times (Coniacian–Santonian boundary at ~85 Ma). The disposition and nature of the HP/LT metamorphic units in the Internal Betics would fit with an initial SE-dipping subduction of the Iberian lithosphere beneath Africa.

The Africa–Europe convergence was initially partitioned across the Pyrenees and the Betic–Rif Iberia boundaries although it became totally accommodated across the southern margin of Iberia from

middle Oligocene onward. More specifically, our kinematic model uses long term estimations of African convergence rates calculated from Rosenbaum et al. (2002a) (Fig. 9). In the first interval, from Late Cretaceous to mid Oligocene, the total convergence rate between Africa and Europe has been equally distributed between the Pyrenees and the Betic–Rif orogenic systems. During the second interval, after mid Oligocene times, the African convergence is fully accommodated across the southern margin of Iberia once the tectonic activity along the Pyrenees is near to ending. For the sake of simplicity, we use long-term uniform convergence rates although changes in the speed at which Africa moved northward are documented particularly between 67 and 52 Ma (slow down of convergence rates as recognized in Cande and Stegman, 2011 and Rosenbaum et al., 2002a). The velocity variations are not considered in the presented kinematic model but must be included in further refinements of the model. Our kinematic model rests on the following major considerations: a) an initial segmented disposition of the Ligurian–Tethys ocean, b) a SE-dipping subduction of the Betic–Rif segments of the Ligurian–Tethys ocean beneath Africa, which are separated from the Algerian segment by a roughly NW–SE paleo-transform fault, c) a total decoupling of the External and Internal Betics along the upper Triassic evaporites that produced the cover fold and thrust system above it and the partial subduction, HP metamorphism and exhumation of cover and basement rocks beneath it, and d) a slow SE-dipping subduction during

Late Cretaceous to mid Oligocene followed by a mid-Oligocene to late Miocene westward slab twisting with coeval fast retreat and lateral tearing.

5. Betic–Rif kinematic evolution model

Our kinematic model shows the evolution of the Betic–Rif orogenic system along an approximate SE–NW transect that turns to more E–W in its northern segment to accommodate the changing directions of the orogen during its progression (Fig. 7). The model is separated in 7 time-steps that summarize the major geodynamic processes occurring in the Betic–Rif orogenic system from Late Cretaceous to present (Figs. 8 and 9). Analysis of older processes related to the earlier transcurrent and transtensional dynamics of the Africa–Iberia plate boundary during Jurassic times (e.g., Puga et al., 1999, 2011; Tubía, 1994) is beyond the scope of this paper.

The seven time steps are represented by means of lithospheric scale sections to show principal geodynamic events occurring for each time step. These time-steps correspond to stages with constraining data although they might also account for processes that could occur somewhat before or/and after. The outlined geometry of the lithospheric mantle through time, although necessary to unravel the entire geodynamic evolution of the Betic–Rif orogenic system, must be taken with caution.

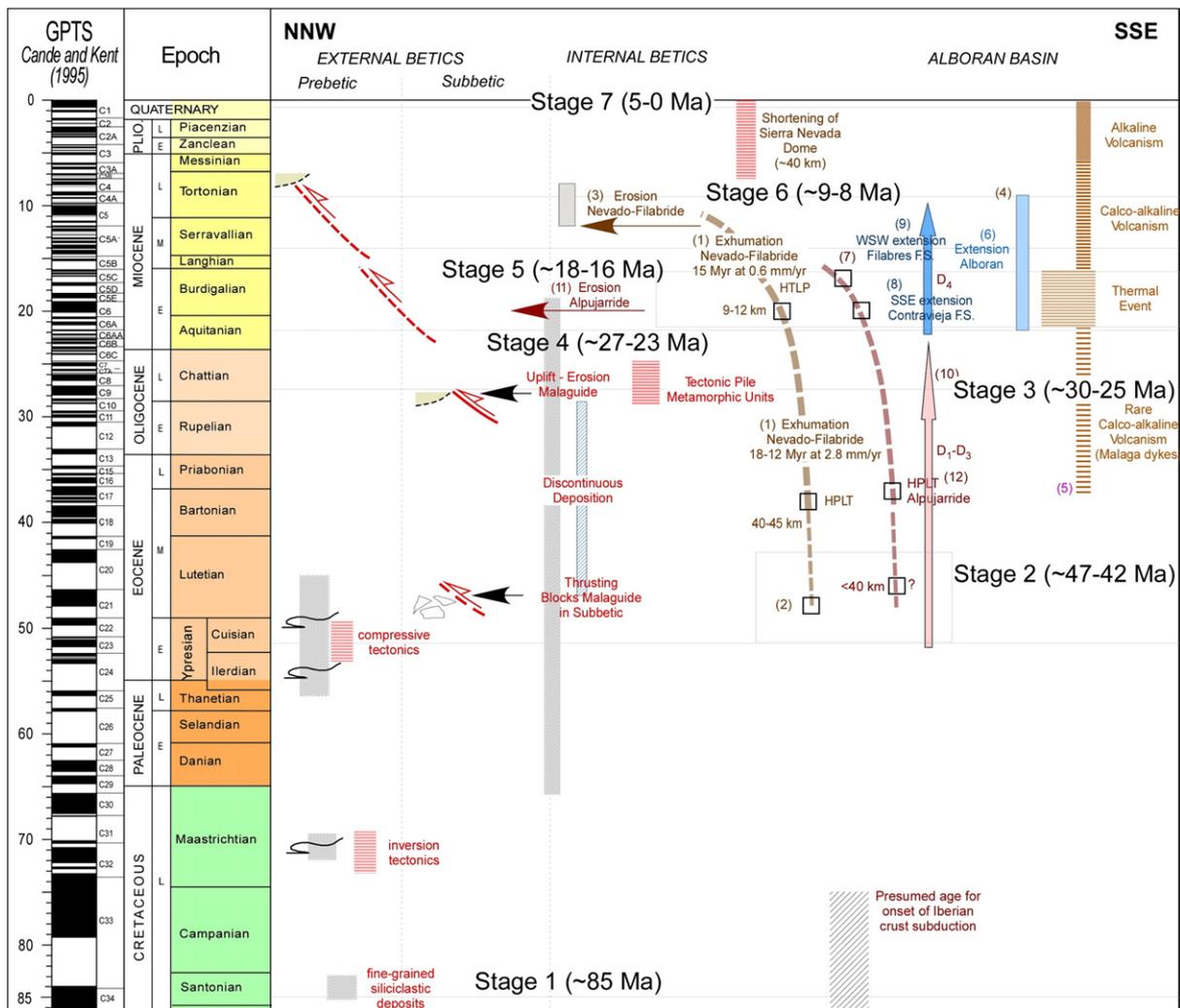
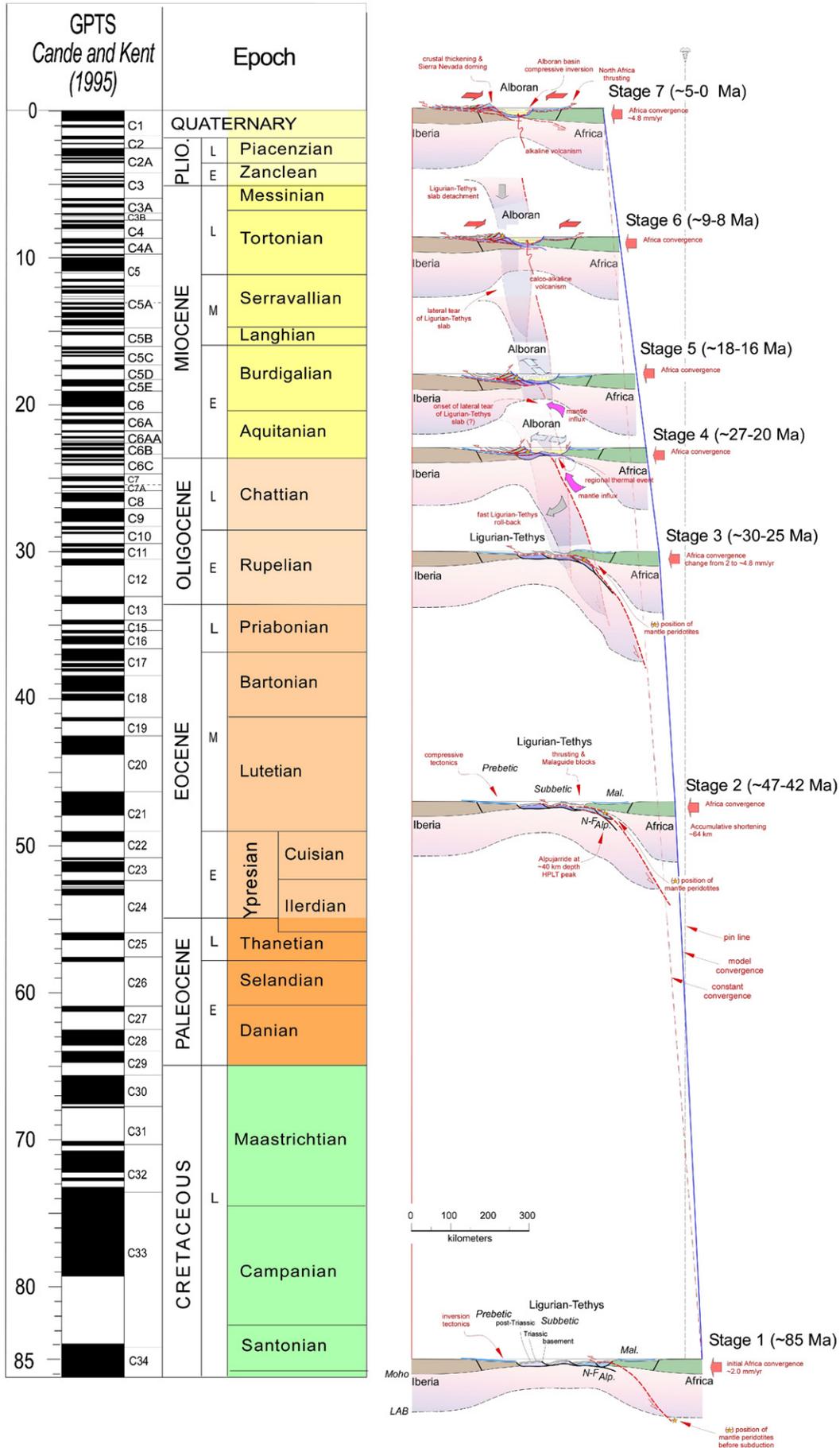


Fig. 8. Condensed chronostratigraphic chart for the Betic segment of the Betic–Rif orogenic system combining different datasets from both External and Internal units. (1) Augier et al. (2005); (2) Monié et al. (1991); (3) García-Dueñas et al. (1992); (4) Zeck et al. (1992); (5) Crespo-Blanc et al. (1994); (6) Lonergan and Mange–Rajetzky (1994); (7) Vissers et al. (1995); (8) Johnson et al. (1997); (9) Comas et al. (1999); (10) Duggen et al. (2004); (11) Rossetti et al. (2005) (12) Monié et al. (1991, 1994); and Platt et al. (2005).



5.1. Stage 1 – model set-up at the onset of the northern convergence of Africa during Late Cretaceous at ~85 Ma

The initiation of the northerly convergence of Africa is established at Late Cretaceous and although earlier phases of compression/transpression were inferred in the External Betics (e.g., de Jong, 1990) we use Santonian (~85 Ma) as the initial configuration of our kinematic model (Figs. 9 and 10). The Late Cretaceous compression in the Betic system is supported by field observations showing punctuated events particularly in the Prebetic units where they have been preserved (e.g., Martín-Chivelet and Chacón, 2007; Vilas et al., 2003). These early compressive events are well-calibrated by the occurrence of planktonic foraminifera in syntectonic sediments encompassing the Late Cretaceous, Paleocene and early Eocene periods, and have been related to the far field stresses away from the active African front (Martín-Chivelet and Chacón, 2007). The first compression event in the Prebetic region is proposed to occur during the late Santonian. The mid Maastrichtian event caused a clear inversion of previous extensional structures, large-scale uplift and coeval sedimentation of large olistholiths in the Prebetic as well as rapid deepening of the Subbetic basin with deposition of siliciclastic turbidites (Chacón, 2002). Palaeocene events are also recognized in the study area but the early Eocene event is correlated to noteworthy compression and erosion in the Prebetic (De Ruig, 1992; Geel and Roep, 1998) and to a deposition of large olistostromes in the Subbetic (Comas, 1978; Vera, 2004).

These early ages of deformation in the Betic–Rif system are in agreement with early compressive events in the Pyrenees (e.g., Déramond et al., 1993; Garrido-Megías and Ríos, 1972; Puigdefàbregas and Souquet, 1986), which were interpreted as due to a large plate reorganization (Dercourt et al., 1986) (Fig. 9). Closer to the Betic–Rif system, the SW region of the Iberian plate (Alentejo basin), was deformed fairly continuously since Late Cretaceous but particularly during the middle Eocene (Pereira et al., 2011). Furthermore, in the Algarve basin, Triassic salt halokinesis related to regional compression started in Late Cretaceous and was persistent through the Tertiary (e.g., Lopes et al., 2006).

Another important point for discussion is the nature of the crust of the Betic–Rif segments of the Ligurian–Tethys basin, which is difficult to ascertain. We reconstructed these crustal domains using the restored width of the Internal Betics (restored length of Alpujarride and Nevado-Filabride complexes; Platt et al., 2003a,b) and taking into account the type of the scarce remnants of this crust (e.g., Bill et al., 2001; Puga et al., 2011). Therefore, the probable composition of these domains would correspond to highly stretched continental crustal blocks (the future metamorphic Nevado-Filabride and Alpujarride units) possibly separated by belts of serpentized peridotites or patches of newly generated oceanized crust. Such composition is observed in the metamorphic ophiolitic and radiolaritic suites tectonically sandwiched in the Nevado-Filabride Complex (Puga et al., 2011; Sánchez-Rodríguez and Gebauer, 2000; Torres-Roldán, 1979) as well as in tectonic units overthrusting the Subbetic units (Bill et al., 2001). These successions have been interpreted as corresponding to remnants of Ligurian–Tethys oceanic crust (Bill et al., 2001) with isotopic dates from middle Jurassic to late Jurassic (~170–164 Ma to ~146 Ma; Hebeda et al., 1980; Portugal-Ferreira et al., 1995; Puga et al., 1995) or even early Jurassic (Puga et al., 2011; Sánchez-Rodríguez and Gebauer, 2000). Due to the mixed continental and oceanic composition of this Ligurian–Tethys crust domain we use the term Ligurian–Tethys transitional/oceanized crust in our reconstruction. It is worth noting that this proposed original configuration for the Ligurian–Tethys basin is similar to that inferred for the Iberian Atlantic margin where

wide regions of very thinned continental crust and exhumed upper mantle have been described (e.g., Pérez-Gussinyé et al., 2006; Péron-Pinvidic et al., 2007). In this highly stretched crustal domain the preservation of a thick lower crust seems difficult.

In our reconstruction the basement blocks forming the floor of the Ligurian–Tethys are partly located beneath the sedimentary cover of the Subbetic, which was decoupled along the upper Triassic evaporites as discussed earlier. These continental blocks (future Alpujarride and Nevado-Filabride units) underwent partial subduction to depths of few tens of kilometers reaching HP metamorphic conditions at middle Eocene times. The exhumation of large areas of HP/LT metamorphic upper crustal rocks implies their decoupling from the underlying thin lower crust (if existing at the onset of subduction) and upper mantle rocks that constituted the majority of the subducted lithospheric slab (Fig. 9). The cover units of the Prebetic and the Subbetic domains, mostly consisting of uppermost Triassic evaporites and thick Jurassic marine successions, would extend on top of the Iberian margin (particularly the Prebetic units) and on part of the Ligurian–Tethys basin (mostly Subbetic units) in our model. The amount of Ligurian–Tethys transition/oceanized crust covered by deep marine deposits is difficult to ascertain but paleogeographic reconstructions based on balanced cross-sections suggest that the present south-eastern outcrops of the Subbetic units could be restored between 100 km and more than 250 km to the south-east according to García-Hernández et al. (1980) and Platt et al. (2003a), respectively. This proposed configuration could still include an area of the Ligurian–Tethys filled by deep marine or/and flysch-type deposits from Jurassic times onward (Flysch unit) (Figs. 4 and 5).

The SE-dipping Ligurian–Tethys subduction model affected the entire lithospheric mantle (Figs. 9 and 10). This scenario seems to offer an interpretation for the deep mantle peridotites originally located at depths of ~140 km and then reaching depths of 85 and 65 km before the onset of the rapid isothermal decompression as has been recently determined for the Ronda peridotites using P–T–t paths (Garrido et al., 2011). The deep positions of the Ronda peridotites at 140 and 85 km depths could be related to extension processes occurring during the early Jurassic opening of the Ligurian–Tethys ocean. Nevertheless, the next step at 65 km depth has been linked to the subduction-related orogenic period (Tubía, 1994). During subduction-related processes the Ronda and Beni Bousera peridotites were exhumed to shallower depths, placed in contact with the Alpujarride units, sandwiched between two of these HP/LT metamorphic slices, exhumed to shallow crustal depths and finally exposed (e.g., Garrido et al., 2011 for Ronda and Afiri et al., 2011 for Beni Bousera).

5.2. Stage 2 – SE-dipping subduction and related HP/LT metamorphism during middle Eocene at ~47–42 Ma

Several HP/LT peak metamorphic ages have been proposed for the different tectono-metamorphic units of the Betic–Rif system. These may correspond to a unique subduction but with diachronous exhumation of different crustal slivers detached from the subduction plane (see Negro et al., 2008 for a review of the timing of tectonic and metamorphic events in the Alboran domain units of the Betic–Rif orogen). The older ages of HP metamorphism in the Nevado-Filabride Complex are middle Eocene in ages between 49 and 42 Ma (Augier et al., 2005 and Monié et al., 1991, respectively) (Figs. 8, 9 and 11). These HP/LT metamorphic ages are similar to those reported for the Sebide–Alpujarride Complex varying from 48 Ma (Platt et al., 2005) and a minimum of 25 Ma (Monié et al., 1991). These HP metamorphic peak ages are thus partly synchronous with the proposed deformation history of the Alpujarride Complex (e.g., Balanyá et al.,

Fig. 9. Crustal kinematic model showing the general evolution of the Betic–Rif system and its timing (see location in Fig. 7). The different steps are based on the chosen time periods, which are constrained by geological and geophysical data as discussed in the text and in Fig. 8. The dashed line on the right side of the model represents a constant rate of Africa convergence from Late Cretaceous to present. The blue line shows the modeled convergence velocity, which increases in the last 30 m.y.

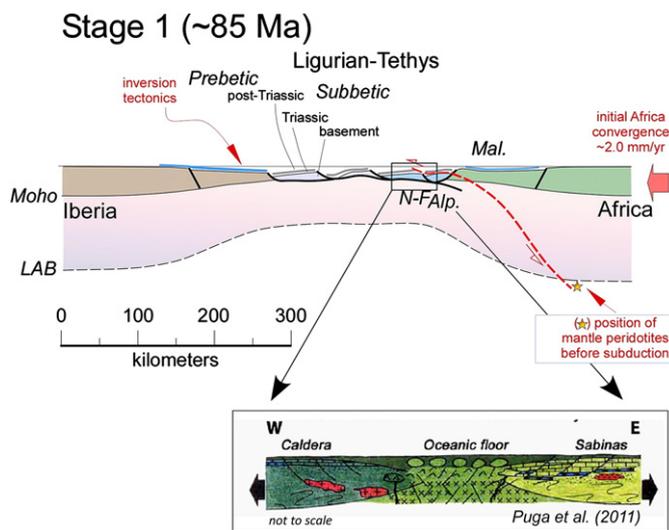


Fig. 10. Stage 1: model set-up at the onset of the northern convergence of Africa during Late Cretaceous at ~85 Ma. The inset illustrates the proposed oceanization of the Jurassic Ligurian–Tethys (from Puga et al., 2011; their Fig. 9).

1997; Platt and Vissers, 1989; Simancas and Campos, 1993). Both metamorphic and tectonic histories are younger in the External Rif, occurring during the Oligocene (Afiri et al., 2011; Negro et al., 2008). Although ages as young as early–middle Miocene have been proposed for the HP/LT metamorphism in the Nevado-Filabride units (López Sánchez-Vizcaíno et al., 2001; Platt et al., 2006) it is very difficult to fit these ages within an integrated kinematic model that combines metamorphism, tectonics, exhumation and deposition histories. There is, however, a general agreement for an Eocene–Oligocene age for the HP metamorphism around the Betic–Rif orogen (see Michard et al., 2006 for a review). The pressure–temperature trajectories calculated for the HP/LT Alpujarride units indicate a single cycle of burial–exhumation with high-pressure metamorphic peaks ranging from 0.7 GPa and 340 °C (~21 km) to 1.1 GPa and 570 °C (~33 km) (e.g., Azañón and Crespo-Blanc, 2000), followed by a rapid isothermal retrograde decompression (e.g., Comas et al., 1999; Rossetti et al., 2005). The different tectono-metamorphic slices of the Nevado-Filabride Complex display metamorphic conditions ranging from 1.2 to 1.6 GPa and 320°–450 °C, which correspond to pressure depths of ~35–42 km to eclogitic conditions of 2.0–2.2 GPa and 650°–700 °C that point to depths of ~60–65 km (e.g., Puga et al., 1999, 2002). In contrast, the Malaguide unit is formed by low-grade metamorphic Paleozoic basement and Permo-Triassic sedimentary cover followed by a kilometer-thick Jurassic to Lower Miocene shallow-water and discontinuous sedimentary successions (e.g., Lonergan, 1993; Martín-Martín et al., 1997). The essentially non-metamorphic Malaguide unit forms the upper plate of the proposed subduction model, which is supported by the character of the present contact with the underlying Alpujarride units. This contact corresponds to a large normal fault interpreted in this work as the product of the tectonic wedging of the Alpujarride units below the Malaguide–Ghomaride units and above the Nevado-Filabride during its shallow-crustal emplacement (e.g., Jolivet et al., 2003). In the Rif, the Ghomaride–Malaguide Complex displays thermal metamorphism along the base of the unit, which is consistent with the underlying HT–LP lower Alpujarride and thus associated to its tectono-metamorphic emplacement as a wedge (Negro et al., 2006).

The principal characteristics of the Internal tectono-metamorphic complexes of the Betic–Rif system are the high variability of metamorphic conditions from one slice to the adjacent one, the apparent synchronous ages of some of the HP/LT metamorphic events in the Nevado-Filabride and Alpujarride Complex, and the occurrence of eclogitic units in the Nevado-Filabride Complex and peridotites in the Sebti–Alpujarride Complex. These features may favor the interpretation of the Betic–Rif

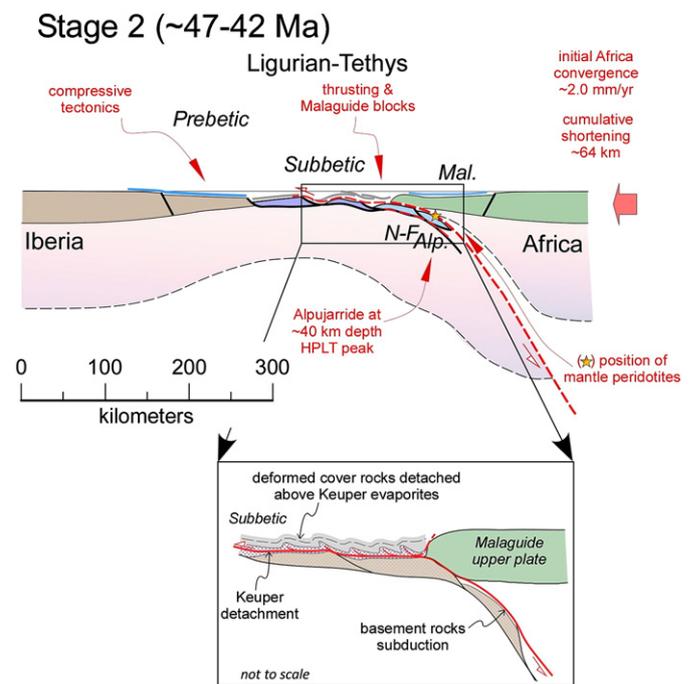


Fig. 11. SE-dipping subduction and related HP/LT metamorphism during middle Eocene at ~47–42 Ma. The inset shows the decoupling along the Upper Triassic evaporites between the folded Ligurian–Tethys sedimentary cover and the stretched continental crust being subducted. This interpretation is important for the tectonic reconstruction of cover and basement in the Betic–Rif orogen.

system growth as due to the subduction and later exhumation of crustal and mantle slices varying in time and space along the boundary between the lower and upper plates (subduction channel) as shown in other subducting-related orogens (e.g., Agard et al., 2009; Monié and Agard, 2009).

In our model, the Ligurian–Tethys domain was carried down along a SE-dipping subducting slab with relative slow convergence rates of about 2 mm/yr between Africa and Iberia. Within this configuration the crustal rocks of the Ligurian–Tethys domain attain 30–40 km depths at around middle Eocene times accordingly with the oldest peak metamorphic ages and with D1 major compressive deformation stage (see Rossetti et al., 2005 for a review of different tectonic evolution models). In the model, based on their present relative positions, the Alpujarride units are located to the SE of the Nevado-Filabride ones although the complexity of the metamorphic conditions makes difficult to unravel the precise pre-metamorphic positions of these two tectono-metamorphic crustal domains. In this subduction scenario, however, their similar peak metamorphic ages could indicate that they were very close to, or even beside each other and were emplaced to some extent oblique.

During the middle Eocene, the subduction of upper crustal rocks was limited to depths of few tens of kilometers and thus not deep enough to produce large partial melting and volcanism, which scarcely affected the Malaguide upper plate during the Oligocene and the early Miocene (e.g., Turner et al., 1999). The exhumation of deeper mantle rocks sandwiched within metamorphic crustal slices would imply that these mantle rocks were located at depths of few tens of kilometers at this time period and in contact with the tectono-metamorphic crustal units before their tectonic imbrication.

Several phases of shortening and extensive siliciclastic deposition have been identified in the External Betics from latest Cretaceous to Eocene times (e.g., Martín-Chivelet and Chacón, 2007; Vilas et al., 2003). In the south-eastern Subbetic paleogeographic domains the late Eocene olistostromic deposits were directly supplied from the Subbetic units (Vera, 2000), which deformed at that time. The Malaguide unit (upper

plate) shows discontinuous and shallow marine deposition through most of the Lower Tertiary (e.g., Maaté et al., 2000) and was the source for clasts deposited in the adjacent Subbetic sedimentary basin during Lutetian times. All these early Tertiary events attest shortening related to the push of the African plate. According to our model, cumulative shortening at middle Eocene times may account for up to few tens of kilometers.

5.3. Stage 3 – onset of slab roll-back and exhumation of Ligurian–Tethys HP/LT metamorphic rocks after middle Oligocene at ~30–25 Ma

The significant decrease of the tectonic activity in the Pyrenees at middle Oligocene times approximately coincided with the sudden increase of the rates at which the geological processes occurred across the Betic–Rif orogenic system. This significant geodynamic change along the southern margin of Iberia has been used in the majority of the prevailing models as an indicator of the true onset of the Betic–Rif orogeny. In our kinematic model we also assume these observations by increasing the convergence rate between Africa and Iberia to close to the present ones (~4.8 mm/yr), which is coarsely representative for the Betic–Rif system from ~30 Ma onward (Figs. 9 and 12). This rate however, represents a maximum amount since part of this convergence was accommodated within the Iberian plate, amounting to few tens of kilometers of shortening mostly during Miocene times (e.g., Banks and Warburton, 1991; Casas-Sainz and Faccenna, 2001; de Vicente and Vegas, 2009; Pereira et al., 2011; Vergés and Fernández, 2006).

In the External Betics different compressive events took place during Paleocene, Eocene and Oligocene times (e.g., Martín-Chivelet and Chacón, 2007) coincident with the tectonic episode determined in the Sierra de la Espuña, between the External and Internal Betics, dated as latest Oligocene (e.g., Lonergan and Platt, 1995; Martín-Martín and Martín-Algarra, 2002; Vissers et al., 1995). During mid-Oligocene time (ca. 28 Ma) the Malaguide upper plate showed the most important phase of tectonic activity, becoming the source area for the southeasternmost Subbetic foredeep (e.g., Vera, 2000). The relatively elevated position of the Malaguide upper plate would be sustained by the shallow crustal emplacement of the HP/LT metamorphic units along a subtractive contact beneath it (e.g., Azañón et al., 1997; Lonergan and Platt, 1995) (Fig. 12). Exhumation rates of about 2.8 km/m.y. for the Nevado-Filabride HP/LT metamorphic slices from middle Eocene to middle Miocene (Augier et al., 2005) are coeval with D2–D3 deformation phases (e.g., Rossetti et al., 2005) (Fig. 8). Rossetti et al. (2005) indicate a progressive top-to-the N/NE shear sense of ductile exhumation for the Alpujarride Complex cropping out to the south of the Sierra Nevada region. The metamorphic rocks of this complex, as well as those of the western Betics (e.g., Tubía et al., 1992), have had a long tectonic and metamorphic evolution as discussed in several papers (e.g., Azañón and Crespo-Blanc, 2000; Rossetti et al., 2005). These authors defined three compressive phases related to subduction and exhumation and a post-orogenic extensional phase. Ductile folding, foliation and E–W trending stretching lineations (D2–D3) formed during the top-to-the-N/NE progressive shearing during exhumation subsequent to the HP metamorphic peak (D1) (Figs. 8 and 12).

Scarce calc-alkaline volcanism, mostly concentrated in the Malaga dykes, is dated as Eocene–Oligocene at 38–30 Ma (Duggen et al., 2004). Rare volcanism is also intruding the Malaguide upper plate during the Oligocene and early Miocene (Turner et al., 1999).

5.4. Stage 4 – shift to E-dipping subduction, opening of the Alboran back-arc basin and collision of the orogen during late Oligocene (?)–early Miocene times at ~27–20 Ma

The geodynamic scenario for this time period, which partially overlaps the previous stage, is characterized by multiple events that

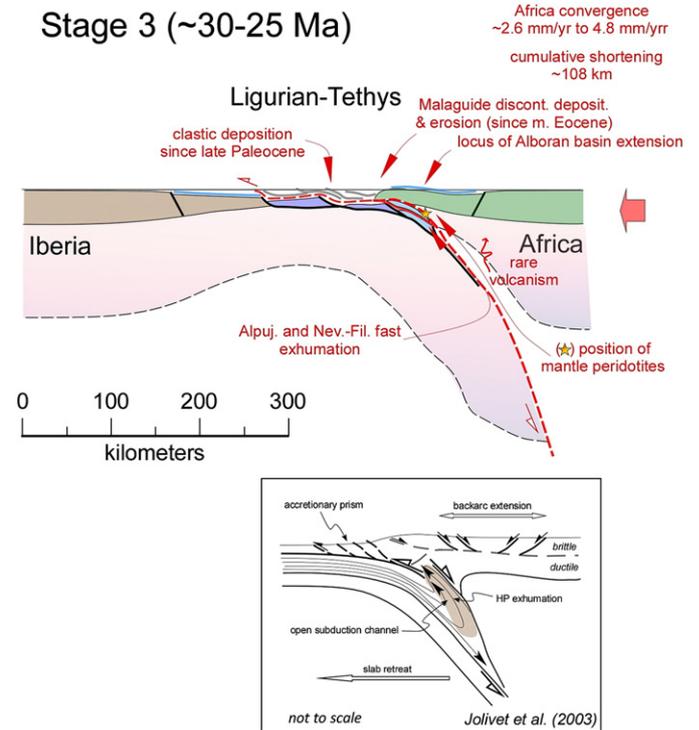


Fig. 12. Stage 3: slab roll-back during middle Oligocene at ~30–25 Ma. The inset shows exhumation of HP metamorphic rocks along the subduction channel during subduction rollback and back-arc extension (from Jolivet et al., 2003; their Fig. 20).

occurred roughly synchronously: a) the opening of the Alboran basin in the inner side of the Betic–Rif orogen; b) the continuous development of the Betic–Rif fold-and-thrust belt along its outer side; and c) the rapid exhumation of HP/LT metamorphic rocks in addition to the flooring of the entire Western Alboran basin with HP metamorphic Alpujarride units (e.g., Platt and Vissers, 1989; Soto and Platt, 1999) (Figs. 8, 9 and 13). Furthermore, the geochemistry of the relatively abundant Miocene Alboran lavas is better explained with an east dipping subduction model (Duggen et al., 2008; Lustrino et al., 2011). The relative ages defined by the cross-cutting relationships among different geologic episodes provide the timing for these distinct events. These partially coeval geodynamic events occurring in the Betic–Rif system during the late Oligocene–early Miocene times, particularly the development of the extensional Alboran basin cannot be explained by the increase of the rate of convergence between Africa and Iberia and must be associated with a deeper geodynamic process, as discussed in Fullea et al. (2010). Yet, our model for this stage shows the Betic–Rif orogenic structure at the end of the selected time step when the different tectono-metamorphic units were nearly stacked ahead of the opening Alboran back-arc basin (Fig. 14).

In our model we interpret both the fast rate of thrust advance along the arcuate outer Betic–Rif system and the fast extension forming the Alboran basin in the inner side of the orogen as produced by a change in direction of the Ligurian–Tethys subduction rollback, from NW-retreating to W-retreating during the late Oligocene–early Miocene times (Fig. 7). The western retreat of subduction has been proposed in the majority of subduction related interpretations for this region to generate the present structure of the Betic–Rif system (e.g., Augier et al., 2005; Gutscher et al., 2002; Jolivet et al., 2008; Lonergan and White, 1997). Augier et al. (2005) proposed a progressive slab retreat to the south (30–22 Ma) and then west starting at ~18 Ma and ending at about 9 Ma whereas Gutscher et al. (2002) proposed that this E-dipping subduction is still active. The change in subduction direction and the consequent rapid opening of the Alboran back-arc basin

(e.g., Martínez del Olmo and Comas, 2008) is caused in our kinematic model by the consumption of a new segment of the Ligurian–Tethys oceanic domain located to the west, which would roughly correspond to the Gulf of Cadiz one in our model. Its consumption by subduction would start or accelerate the opening of the Alboran back-arc basin during late Oligocene (?)–early Miocene times (e.g., Comas et al., 1992, 1999) after or close to the end of the oceanic consumption of its eastern segments. The crustal structure along which the westernmost tip of the Betic–Rif trench changed its direction corresponds in our model to the precursor of the Alboran Ridge fault zone (Martínez-García et al., 2011) separating the Rif from the African triangular crustal domain composed by thinned crust and magmatic intrusions (Booth-Rea et al., 2007) (see black circle in Fig. 7). The change in the direction of subduction retreat from NW to W around the westernmost tip of the trench during the late Oligocene (?)–early Miocene could propitiate the bending of the Betic–Rif orogen and the general diachronous formation of the Betic–Rif system. The subduction related processes were still active after early Miocene times along the western subduction front (Rif and western Betics) while they were more mature along the eastern and central segments of the Betics (e.g., Chalouan et al., 2008).

The Alboran domain shows a complicated structure with its major foredeep located at the Western Alboran basin showing an arcuate geometry parallel to the western side of the Betic–Rif orogen (Comas et al., 1999). The Western Alboran basin shows about 8-km thick sedimentary infill well-dated as early Miocene to Recent. However, the lower deposits

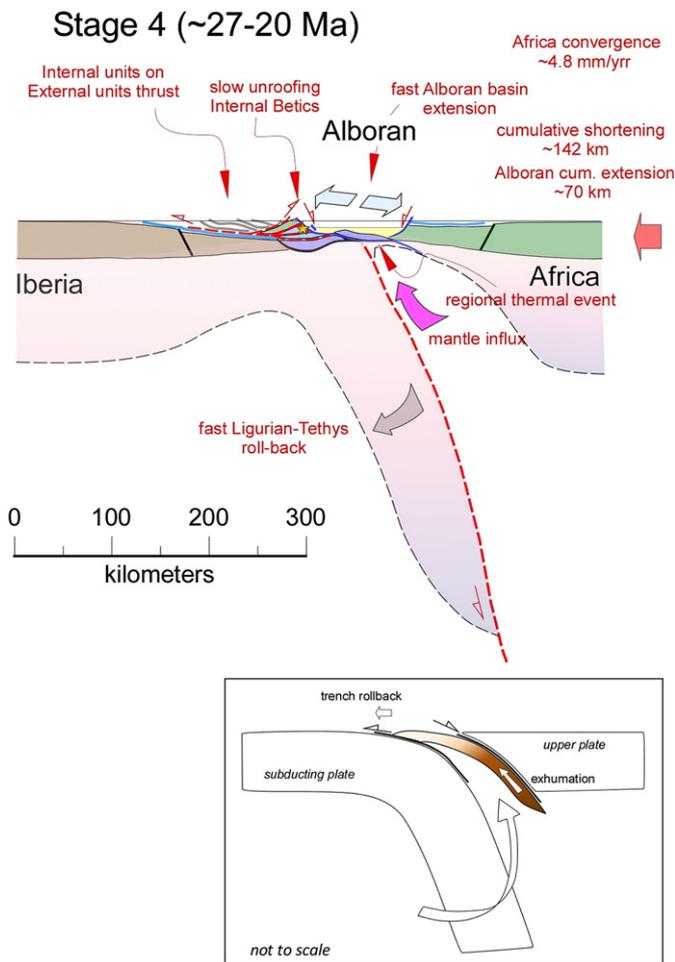


Fig. 13. Stage 4: acceleration of Betic–Rif thrusting and opening of the Alboran back-arc basin during the late Oligocene–early Miocene at ~23 Ma. The inset shows massive exhumation of high grade metamorphic rocks to infill the opening created during back-arc extension behind a fast slab retreat (from Jolivet et al., 2003).

in the deepest part of the basin could be earliest Miocene and even late Oligocene, in agreement with thermal modeling supporting this age for the onset of basin formation (in Comas et al., 1999). The basement of the Western Alboran basin is composed of HP metamorphic rocks equivalent to the Alpujarride Complex (e.g., Platt et al., 1998; Soto and Platt, 1999).

Thermochronological data also propose that these Alpujarride metamorphic rocks were exhumed to reach shallow crustal levels during the late Oligocene. The Alpujarride high grade metamorphic rocks, stacked in slivers of different sizes and metamorphic gradients, could be exhumed along a ductile shear zone (subduction channel) between the down going subducting plate and the upper plate especially under local extensional conditions (e.g., Gerya, 2002; Jolivet et al., 2003; Monié and Agard, 2009). Jolivet et al. (2003) also pointed

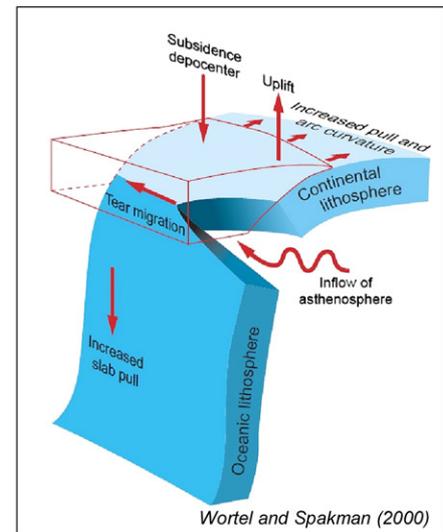
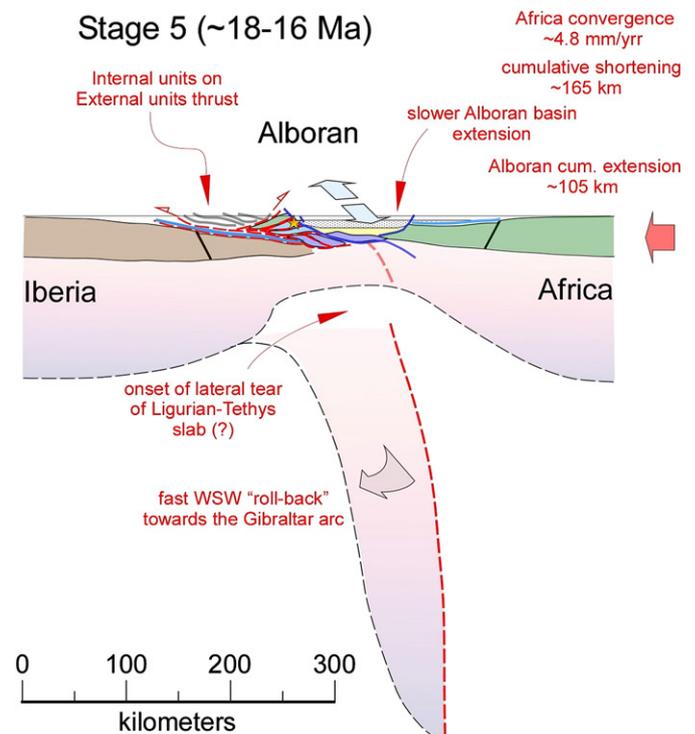


Fig. 14. Stage 5: cessation of major Alboran extension during the middle Miocene at ~16 Ma. The inset shows interpretation of a lateral tear in a subducting slab (from Wortel and Spakman, 2000; their Fig. 4).

out that usually the width of the subduction channel is related to the amount of subduction roll-back that strongly contributes to open the channel, as corroborated by laboratory experiments (Bialas et al., 2011). According to these authors, the fast trench retreat together with the toroidal component of the mantle flow might exhume huge slivers of previously subducted crustal rocks (Fig. 13). This process seems to be a good explanation for the flooring of the Western Alboran basin with HP/HT Alpujarride metamorphic rocks, which occurred synchronously to the stretching of the Internal Betic complexes presently exposed onshore (e.g., García-Dueñas et al., 1988; Martínez-Martínez et al., 2002).

A large and fast extension at shallow crustal levels above the brittle–ductile transition in combination with erosion removed large portions of the overburden as has been determined for the Alpujarride Complex for this time period. The metamorphic Alpujarride Complex has been the source for foreland basin deposits in the central–eastern Betics during the early Miocene times (e.g., Lonergan and Mange-Rajetzky, 1994). The extension affecting the tectonic stack of Internal Betic and Rif units at the onset of the Alboran basin development indicates that these metamorphic units were already piled up at shallow crustal levels (see review in Negro et al., 2008) in agreement with Rossetti et al. (2005). The extent of the Malaguide upper plate unit, which is currently restricted to the foreland-dipping part of the Internal units, and its absence above the Alpujarride basement unit in the Alboran basin, is explained by its tectonic disruption at the beginning of the opening of the Alboran back-arc basin by normal faulting. Only the trailing edge of this unit remained above the Alpujarride outcrops in both the Betics and the Rif segments of the orogen.

In our kinematic model the front of the Betic–Rif fold-and-thrust belt system migrates toward the foreland with similar velocities as the opening of the Alboran back-arc basin at the end of the Oligocene and early Miocene times (Fig. 13). This is in full agreement with the fast period of fold-and-thrust belt advancement (see tectono-stratigraphic charts in Crespo-Blanc and Frizon de Lamotte, 2006; Chalouan et al., 2008). This faster thrust deformation was already proposed as a major Burdigalian tectonic phase, characterized by large tectonic displacements, followed by less important shortening during the Langhian and Serravalian times (Guerrera et al., 1993; Martín-Algarra, 1987). Concordantly, to accommodate the fast tectonic shortening we tentatively imbricate most of the Subbetic units and we initiate their tectonic emplacement on top of the carbonates of the Prebetic unit (Fig. 13). The stacked metamorphic complexes (Internal Betics), along the inner part of the fold-and-thrust belt possibly overrode the SE Iberian margin during the early phases of the Alboran basin opening. This produced continental collision via the emplacement of Malaguide upper plate on top of the Iberian continental margin cover deposits corresponding to the lower plate within the defined SE-dipping subduction geodynamic context. This orogenic collision has been usually documented as taking place, along the strike of the Betic chain, between the rather rigid Alboran block and Iberia (e.g., Soria, 1994; Vera, 2000).

The timing of tectonic activity of the Internal Betic front can be documented locally in the eastern Betics by syn- and post-tectonic deposits (Geel and Roep, 1998), which indicate the end of the collision during the late Aquitanian time (~21–22 Ma). This collision seems to be also recorded at plate tectonic scale in the slowdown of Africa at 20–15 Ma (Missenard and Cadoux, 2012).

During the early Miocene, the Alpujarride basement rocks flooring the Alboran basin were affected by an intense HT metamorphic peak documented in the whole Alboran domain (e.g., Comas et al., 1999; Soto and Platt, 1999) and dated as ~23–21 Ma (e.g., Negro et al., 2006; Platt et al., 2003b) (Fig. 13). Although mostly restricted to the Alboran basin, the high temperature metamorphism also affected some of the Betic–Rif Internal units presently cropping out onto land as clearly identified in the Ghomaride–Malaguide units in the Rif (Negro et al., 2006). This HT metamorphic peak is related to the

concomitant influx of hot asthenospheric mantle material beneath the extending Alboran domain during the subduction roll-back processes (Fig. 13). This hot mantle influx directly affected the already emplaced HP Alpujarride tectono-metamorphic units that formed the basement of the evolving Alboran back-arc basin.

Interestingly, this HT metamorphic event identified at early Miocene time could produce a thermal resetting of some of the thermochronological ages as recently proposed for the Betic–Rif system (see discussions in Lustrino et al., 2011 and Negro et al., 2006). If true, this thermal resetting could explain the cluster of early Miocene thermochronological ages (mostly FT on apatite and zircon) acquired in the metamorphic complexes, the interpretation of which is difficult to reconcile with geological observations as argued in Zeck (1996). This author reports exhumation ages of the metamorphic units (based on detrital thermochronology) that are younger than their proposed ages of deposition. However, some of these correlations were made along the strike of the Betic chain without considering the diachronous formation of the Betic–Rif system, younging to the west, and thus with potential younger ages of exhumation of metamorphic units in the same direction.

The fast lithospheric extension and thinning at the onset of the Alboran back-arc basin together with the upward flow of hot asthenospheric material during the latest Oligocene and early Miocene triggered the production of a large amount of different igneous episodes, mostly crustal derived, from 23 to 18 Ma (e.g., Lustrino et al., 2011; Wilson and Bianchini, 1999) (Figs. 8 and 13).

The complete exhumation of subducted middle and upper crustal rocks to shallow crustal levels by early Miocene times strongly point to a subsequent evolution of the subducting slab only constituted by lithospheric mantle rocks. This late evolution of the subducted or delaminated lithospheric mantle since early Miocene is consistent with the observed present geometry of the Ligurian–Tethys subducted slab beneath the Betic–Rif system (e.g., García-Castellanos and Villaseñor, 2011; Wortel and Spakman, 2000) and with the crustal and lithospheric thicknesses determined along the orogenic system (e.g., Fullea et al., 2010; Torne et al., 2000) as documented in the next stages of the Betic–Rif evolution (Fig. 9).

5.5. Stage 5 – end of major Alboran extension during the middle Miocene at ~18–16 Ma

A large and more than 1-km thick olistostromic unit (the Guadalquivir Allochthon) deposited during Langhian and Serravalian times along the front of the Betic fold-and-thrust belt (e.g., Martínez del Olmo et al., 1999; Perconig, 1960). This middle Miocene massive deposition, which filled the Guadalquivir shallow-marine foreland basin, reflects the considerable denudation of the thrust system, which was characterized by the imbrication of Subbetic units on top of the Prebetic carbonate units, mostly during the previous stage of orogenic build up in early Miocene (Figs. 9 and 14). Toward the west, the Gulf of Cadiz Imbricate Wedge evolved, at least partially during this time period, in front of the migrating Betic–Rif arcuate orogen as a result of the tectonic thickening of the sedimentary cover allocated in the Ligurian–Tethys oceanic crustal segment that forms the present Gulf of Cadiz (e.g., Gutscher et al., 2002, 2009a,b; Iribarren et al., 2007; Maldonado and Nelson, 1999; Medialdea et al., 2004).

The hinterland of the Betic thrust system was characterized by the rapid exhumation of the Nevado-Filabride HP/LT metamorphic units, which reached shallow crustal levels at rates varying between 0.5 and 1.8 mm/yr, depending on the applied temperature gradient (e.g., Augier et al., 2005; Johnson et al., 1997; Negro et al., 2008; Reinhardt et al., 2007; Vázquez et al., 2011). The Nevado-Filabride Complex shows prominent E–W trending stretching lineations interpreted as related to E–W extension with top-to-the-west displacement from 18 to 8 Ma (middle–late Miocene) (Augier et al., 2005; Johnson et al., 1997). E–W

directed stretching lineations and top-to-the-west displacement from the Rif units are related to the same extensional period by Negro et al. (2008). The fast exhumation and final emplacement of the Nevado-Filabride at shallow crustal levels (beneath the Alpujarride) caused its exposure and erosion at the end of the middle Miocene (~12 Ma) with the coeval clastic supply to the piggy back basins ahead of the Internal Betics front (e.g., Lonergan and Mange-Rajetzky, 1994). These basins were further carried on top of the NW-propagating Betic thrust system.

Extensional tectonics affected both the Alboran basin as well the Betic Internal units by regional top-to-the WSW-directed normal faults, which were active during the early Miocene times (e.g., Galindo-Zaldívar et al., 1989; García-Dueñas et al., 1992) from at least 18 to 8 Ma (Johnson et al., 1997; Martínez-Martínez and Azañón, 1997) or up to 6 Ma (Vázquez et al., 2011). Structural and metamorphic evolutions of the eastern external Rif are also in good agreement with these data (Negro et al., 2007, 2008).

In the Alboran basin, however, during the late Burdigalian–early Langhian time, post-extension clastic deposits overlapped the extensional system (e.g., Martínez del Olmo and Comas, 2008), constraining the upper boundary of this late Oligocene and early Miocene extensional tectonics at about 15 Ma. As discussed in previous time steps, the emplacement of the Internal Betics took place before the major extensional phase as the normal faults cut across the stack of metamorphic units and partially or totally reactivate the previous tectono-metamorphic contacts. This is in agreement with regional field observations along the Internal–External boundary, which is fossilized by Burdigalian–Langhian clastic deposits in the eastern and western Betics (e.g., Geel and Roep, 1998; Mazzoli and Martín-Algarra, 2011, respectively) (Fig. 5).

Volcanism was active in the Alboran basin presenting a keel-shaped distribution, which has been directly related to lithospheric structure at depth (Duggen et al., 2008; Lustrino et al., 2011). Alkaline rocks dated from ~12.8 to 8.7 Ma crop out in the center of the Alboran basin whereas calc-alkaline volcanic rocks ranging from ~15 to 6 Ma are distributed around the alkaline ones (these volcanic events occurred between stages 5 and 6, Figs. 8 and 9). This distribution of volcanic rocks has been interpreted as the result of the roll-back and steepening of a narrow remnant of oceanic slab causing mantle delamination (Duggen et al., 2004, 2008). Nonetheless, it could be alternatively interpreted as the product of lateral tearing of the Ligurian–Tethys lithosphere subducted beneath the Betics commencing in its eastern side (e.g., Wortel and Spakman, 2000), which possibly caused thermal thinning of the lithospheric mantle (Fig. 14). The distribution of alkaline volcanism in the center of the Alboran basin fits with the highly attenuated crust and lithospheric mantle that still persists (e.g., Dünder et al., 2011; Fulla et al., 2007, 2010; Soto et al., 2008; Torne et al., 2000). This lithospheric thinning, linked to the lateral tear of the subducting lithosphere, could trigger further regional exhumation but mostly along the Internal Betics directly located above the subducting slab. The onset of this uplift maybe constrained between middle Miocene and Tortonian (Figs. 14 and 15).

5.6. Stage 6 – tightening of the Betic–Rif orogenic system during the late Miocene after 9–8 Ma

During the late Miocene, both radial compression along the outer Betic thrust system and late-stage extension along the Alboran back-arc region rapidly declined (Figs. 9 and 15). The data show that Serravallian deposits covered the oldest extensional system coeval to the formation of the Alboran basin (Martínez del Olmo and Comas, 2008) whereas Tortonian deposits fossilized the top to the WSW-directed fault system (Crespo-Blanc et al., 1994) (Fig. 5). The front of the Betic thrust system, along the tip line of the Guadalquivir Allochthon was also fossilized around the late Tortonian (Berástegui et al., 1998;

Vera, 2000). This age of fossilization is also recognized along the off-shore continuation of the front of the Betic–Rif thrust system in the Gulf of Cadiz (Gràcia et al., 2003; Iribarren et al., 2007), although younger compressional events could deform this domain after 8 Ma (e.g., Gutscher et al., 2002; Zitellini et al., 2004, 2009). The front of the External Rif seems also fossilized during late Tortonian times (Negro et al., 2007, 2008).

The end of extension in the hinterland and major compression along the front of the arcuate Betic–Rif orogenic system at late Tortonian times coincides also with the proposed near end of both calc-alkaline volcanism, in the eastern side of the Alboran Sea at about 9–8 Ma (e.g., Lustrino et al., 2011; Wilson and Bianchini, 1999), and major paleomagnetic vertical-axis rotations at around 7.8 Ma (Krijgsman and Garcés, 2004) although it has been proven that some of these rotations continued to early Pliocene times in the Betic basins (Mattei et al., 2006).

After late Tortonian times the NNW convergence of Africa vs. Iberia was once more the most important active mechanism compressing the margins of Africa and Iberia (Fig. 14). The Alboran basin was squeezed along the western Africa segment, whereas the Algerian basin margin initiated underthrusting beneath the continental crust of Africa (e.g., Déverchère et al., 2005; Mauffret, 2007). Shortening in the Alboran basin reactivated strike-slip faults suitably-oriented normal faults (e.g., Maillard and Mauffret, 2011; Martínez del Olmo and Comas, 2008; Martínez-García et al., 2011). The Betic fold belt was mostly deformed in its hinterland along the Internal units by doming of the Sierra Nevada and other smaller antiformal structural highs like the Sierra de Alhamilla. The doming of the Sierra de Alhamilla initiated in late Tortonian time (Weijermars et al., 1985) and seems to be still

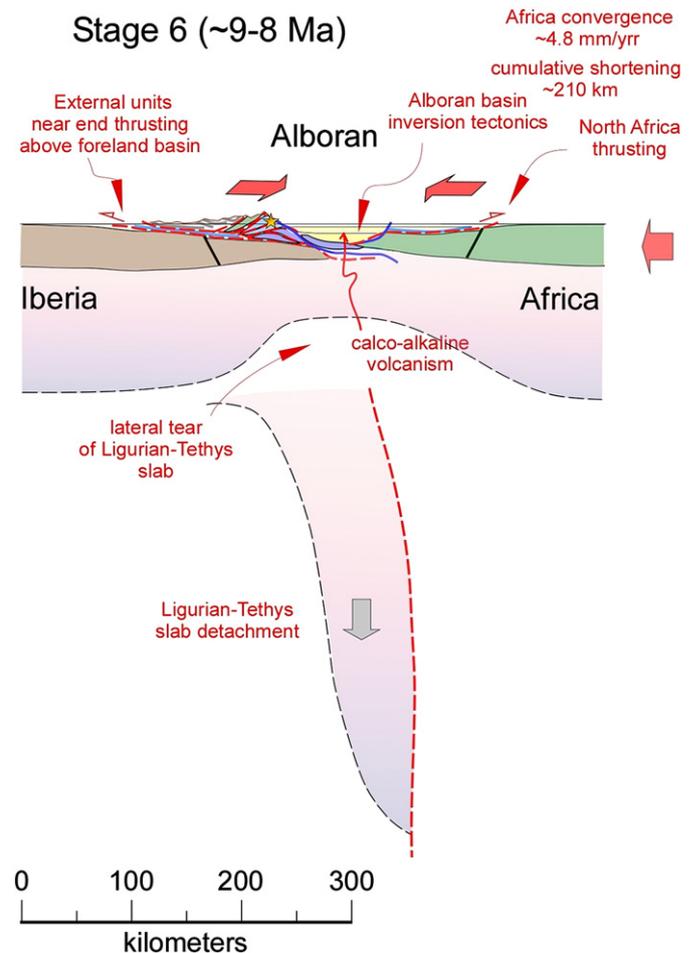


Fig. 15. Stage 6: Betic–Rif compressive tightening during the late Miocene after 9 Ma.

active (Rodríguez Fernández and Martín Penela, 1993). The growth of topography shaping the Betic Cordillera mostly initiated after late Tortonian times as determined by paleogeographic reconstructions (e.g., Andeweg, 2002; Braga et al., 2003; Geel and Roep, 1998; Sanz de Galdeano and Alfaro, 2004; Weijermars, 1991; Weijermars et al., 1985). This growth of topography after 7 Ma resulted in the uplift of numerous small, mildly deformed intramontaneous basins, in a concentric pattern around the Sierra Nevada dome (Iribarren et al., 2009).

Other evidences of tectonic shortening are given by the halokinetic successions in the Algarve basin showing the reactivation of diapiric structures from the late Tortonian to the Messinian (Lopes et al., 2006) and from the Messinian in the central part of the Gulf of Cadiz (Flinch et al., 1996).

The origin of the late Miocene and Pliocene potassic magmas in SW Spain and NW Africa has been interpreted as the product of partial melting of metasomatized subcontinental mantle lithosphere from earlier subduction-related processes (Duggen et al., 2004) or related to delamination and coeval asthenospheric mantle upwelling (Duggen et al., 2005). These sub-lithospheric geodynamic scenarios are discussed in Fullea et al. (2010) and Jiménez-Munt et al. (2011) (Fig. 15).

Although not yet constrained, we also agree with the interpretation indicating that one portion of the vertical movements observed in the Internal Betic zone after late Tortonian times could be related to the westward lateral tearing of the steeply dipping subducted Ligurian–Tethys slab as proposed by Wortel and Spakman (2000). The down-pull of the subducting Ligurian–Tethys lithosphere could produce the subsiding topography of the Betic orogenic system up to the late Miocene (García-Castellanos, 2002). The geometry of the lithospheric mantle beneath the Betic–Rif system during Tortonian would be essentially the same as at present, consisting of a relatively steep high-velocity slab descending down to ~660-km as imaged by regional and global tomographies (e.g., Blanco and Spakman, 1993; García-Castellanos and Villaseñor, 2011; Gutscher et al., 2002) and recent seismicity (e.g., Fernández-Ibáñez and Soto, 2008; Pedrera et al., 2011; Ruiz-Constán et al., 2011). In our model we initiate the lateral tear of the subducting slab as a consequence of the continental collision and subsequent slab roll-back decline during early Miocene (Figs. 14). However, the timing of this lateral lithospheric tear is not constrained and could start later as discussed above (Fig. 15).

5.7. Stage 7 – Pliocene–Quaternary evolution of the Betic–Rif orogenic system

Although available data seem to indicate that most important Betic–Rif orogenic processes were almost finished at late Miocene times, there has been a significant tectonic activity since then to shape the present-day topographic expression of the orogen, which mostly grew in the last few million years (e.g., Braga et al., 2003; Iribarren et al., 2009; Sanz de Galdeano and Alfaro, 2004) (Figs. 9 and 16). Tectonic processes in the Betic–Rif orogen are still active as demonstrated by the large amounts of shallow and deep historical seismicities along the Iberia–Africa plate boundary (e.g., Ruiz-Constán et al., 2009). The N to NNW convergence of Africa against Europe triggered the tectonic inversion of normal faults within the Alboran basin (Martínez del Olmo and Comas, 2008) and the reactivation of compressive structures in the Betic–Rif fold and thrust belt.

However, one of the most recent and remarkable results regarding the deformation at shallow crustal levels comes from the GPS data, which show a general clockwise rotation of the velocity field in the westernmost Betics and Rif. The Rif segment experiences a SW relative motion with respect to stable Africa with present rates between 3.5 and 4.0 mm/yr (e.g., Koulali et al., 2011; Pérez-Peña et al., 2010; Stich et al., 2006). This motion is explained by the SW escape of the Rif segment of the chain parallel to the Alboran Ridge fault zone (e.g., Fernández-Ibáñez et al., 2007; Maldonado et al., 1992) and to the presently left

lateral Nekor Fault onshore (e.g., Asebriy et al., 1993) (Figs. 4, 6 and 7). This active fault zone limits the large triangular thinned continental African crust intruded by arc magmatism (Booth-Rea et al., 2007) impinging Iberia and deforming also the Western Alboran basin (e.g., Booth-Rea et al., 2007; Fernández-Ibáñez and Soto, 2008; Martínez-García et al., 2011; Morel and Meghraoui, 1996; Stich et al., 2003, 2006; Tahayt et al., 2008). The northern limit of the Alboran Ridge is a thrust fault that changes to right-slip fault along the Yusuf Fault (Alvarez-Marrón, 1999; Martínez-García et al., 2011). Further east of the Yusuf Fault the Algerian basin crust is interpreted to be underthrusting beneath the highly irregular northern African boundary (e.g., Déverchère et al., 2005; Mauffret, 2007). The south-western escape of the Rif segment with respect to Africa also explains the crustal thickening imaged beneath the region with maximum thickness values of 34–36 km determined by numerical modeling based on potential fields (Fullea et al., 2007, 2010) and 35–44 km based on the analysis of P-to-S converted waves in teleseismic receiver functions (Mancilla et al., 2012). Interestingly, this crustal thickening is not substantiated by a significant high topography but coincides with a prominent lithospheric mantle thickening striking SW–NE and affecting the Gibraltar region (e.g., García-Castellanos and Villaseñor, 2011) and the NW Moroccan margin (Fullea et al., 2007, 2010; Jiménez-Munt et al., 2011) (Fig. 18A).

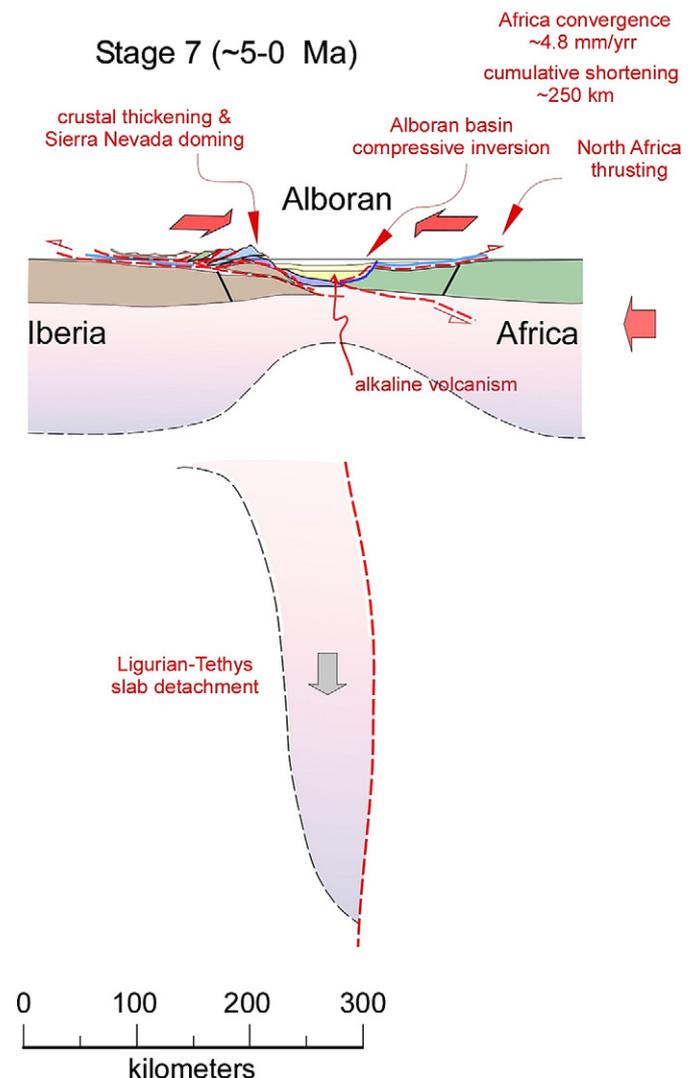


Fig. 16. Stage 7: recent evolution and present-day structure of the Betic–Rif orogen and Alboran basin.

The presence of deep earthquakes (deeper than 620 km) suggests that the cold lithospheric slab located beneath the Betic Cordillera experiences small movements (Buforn et al., 1991; Pedrera et al., 2011; Ruiz-Constán et al., 2009) and could be the cause for uplift and subsidence of the Gibraltar Strait during the latest Miocene (e.g., García-Castellanos and Villaseñor, 2011; Krijgsman et al., 2010). In our model the final length of the subducted lithospheric slab is approximately consistent with the depths proposed by the seismic results and by the depth of the studied earthquake focal mechanisms (Figs. 9 and 16).

The final arrangement of the Betic–Rif orogenic system is directly related to the total consumption of the westernmost segments of the Jurassic Ligurian–Tethys domains between Africa and Iberia and thus of its original size and geometry. Our initial map at Late Cretaceous times, which is based on independent studies, shows a reconstructed Betic–Rif segment of the Ligurian–Tethys transitional/oceanized crustal region of about 200,000 km² (including half of the former Gulf of Cadiz crustal segment; Fig. 7), which is coarsely equivalent to the present size of the subducted lithospheric mantle placed beneath the Betic–Rif orogen as imaged by recent tomographic studies (Blanco and Spakman, 1993; Gutscher et al., 2002; Spakman and Wortel, 2004) thus supporting our proposed kinematic model.

6. Discussion

This section is organized to discuss the most significant geodynamic implications of the presented model, which is characterized by an initial SE-dipping subduction zone along the Betic–Rif segment of the Ligurian–Tethys domain since the Late Cretaceous. We first examine how the currently accepted models fit with the main characteristics of the westernmost Mediterranean region in terms of present-day lithospheric structure, metamorphism and volcanism. Secondly, we discuss the main differences between our model and the currently accepted models in terms of kinematic evolution, including discussion and potential interpretation of vertical axis rotations along the Betic–Rif fold-and-thrust belt. Finally, we discuss the proposed change in subduction polarity across the Betic–Rif and Algerian domains bringing examples from the Western Mediterranean and SE Asia, and the position of the paleo-suture between Africa and Iberia.

6.1. Current models and present-day lithospheric structure

Recent geophysical models based on seismic data, tomography and potential fields show a strong asymmetry in the crustal and lithospheric mantle structures along the Gibraltar arc region. Tomography models (e.g., Bijwaard and Spakman, 2000; García-Castellanos and Villaseñor, 2011; Spakman and Wortel, 2004) and SKS anisotropy analysis (Diaz et al., 2010) draw an arcuate mantle slab restricted below the Betic–Rif orogen, dipping toward the E in the Gibraltar Strait and turning to the SE and S beneath the Betics (Fig. 17). The slab terminates abruptly beneath the eastern tip of the Betics and detaches vertically along a tear zone beneath the central-eastern Betics. In addition, integrated modeling of the crustal and lithospheric structures (e.g., Fullea et al., 2007, 2010; Torne et al., 2000) shows also a strong asymmetry in the crustal structure of the south-Iberia and north-Morocco margins. The northern Moroccan margin, east of the Internal Rif units, is characterized by a smooth crustal thinning toward the Alboran basin whereas the southern Iberian margin presents a much sharper thinning. Recent seismic analyses based on receiver functions also show a sharp decrease of the crustal thickness from the central Rif with values of 35–44 km to 22–30 km near the Morocco–Algeria border (Mancilla et al., 2012). These variations have been interpreted as the result of the tear and detachment of the lithospheric mantle slab beneath the Iberian segment of the Betic–Rif orogen (Figs. 9 and 17).

Among the models proposed to explain the kinematic evolution of the westernmost Mediterranean, the one receiving a wider consensus

is that envisaged by Rosenbaum et al. (2002b) and Rosenbaum and Lister (2004) and also supported by many other authors (e.g., Booth-Rea et al., 2007; Duggen et al., 2004; Gutscher et al., 2002; Spakman and Wortel, 2004) (Fig. 18B). All these studies propose that the Alboran domain underwent a large westward drifting driven by a subduction rollback. Important assumptions of the model are: i) the Alboran domain with HP–LT rocks formed far to the east of its present position through orogenic compression and subsequent extensional collapse driven by subduction rollback with traveling distances between 100 and 800 km; ii) the trench associated with slab-rollback turns clockwise around a pivotal or anchor point located in the present eastern termination of the Betics and a total trench rotation of ~180° in the Betic segment; and iii) the trench and the accretionary wedge displaced westward forming a continuous subduction front along the Betics, Rif and Maghrebides.

Models invoking a westward retreat of the accretionary wedge as the main mechanism forming the arc predict a fairly symmetric crustal geometry of the Alboran basin and its margins, and result in a narrow east-dipping subducting slab (e.g., Duggen et al., 2004, 2005; Gutscher et al., 2002). Alternative models with a more autochthonous component were proposed by Faccenna et al. (2004) and Gueguen et al. (1998). According to these models, the Oligocene subduction trench extended continuously along the whole Iberian Mediterranean margin from the present Gibraltar arc to the Alps with a NW-dipping polarity. Gueguen et al. (1998) do not provide details on the formation of the Gibraltar arc whereas Faccenna et al. (2004) propose that the progressive south-western slab retreating and its rupture beneath the eastern tip of the Internal Rif domain at ~15 Ma would have caused a return mantle flow accelerating the westward trench retreat and its clockwise rotation. A limitation of the model rises in explaining the present configuration of the slab beneath the Betic–Rif belt and the distribution of HP–LT metamorphism.

6.2. Geodynamic implications of an initial SE-dipping subduction beneath the Betic–Rif orogenic system

The kinematics of the Betic–Rif orogenic system proposed in our model and based on an initially SE-dipping subduction differs largely from the currently accepted models (Fig. 18). An important difference is the initial configuration of the Ligurian–Tethys domain during the Late Cretaceous as depicted in Fig. 1C. According to our model and based on recent independent plate tectonic reconstructions, the Iberian and African margins were strongly segmented as a result of the transtensive tectonics during the early Jurassic. Remnants of this segmentation have been inherited in the present structure of the Betic–Rif orogenic system and the Western Mediterranean basin. In most of the previous models, the late Tertiary compressive tectonics is solved by a continuous NW-dipping subduction extending all along the Iberian margin (e.g., Faccenna et al., 2004; Gueguen et al., 1998) or from the present-day eastern end of the Betics to the Alps (e.g., Rosenbaum and Lister, 2004; Rosenbaum et al., 2002b). Booth-Rea et al. (2007) consider an initially segmented margin but with NW-dipping subduction affecting only the easternmost segment of the Ligurian–Tethys domain with the Betic–Rif segment resting unaffected as in the Rosenbaum's model.

Fig. 18 compares the migration of the subduction front resulting from our model and from Rosenbaum et al. (2002b) using the same Ligurian–Tethys reconstruction between the relative positions of Africa and Iberia for the Late Cretaceous times. In our model the consumption of the Ligurian–Tethys oceanized basins occurs parallel to the reconstructed crustal segments instead of occurring perpendicular to them. All the other previous models propose a displacement perpendicular to the Jurassic–Cretaceous direction of extension and parallel to the strike of the subducting slab requiring the concurrence of lithosphere tearing and slab detachment in both the SE Iberian and N African margins. Differences between our model and previous models

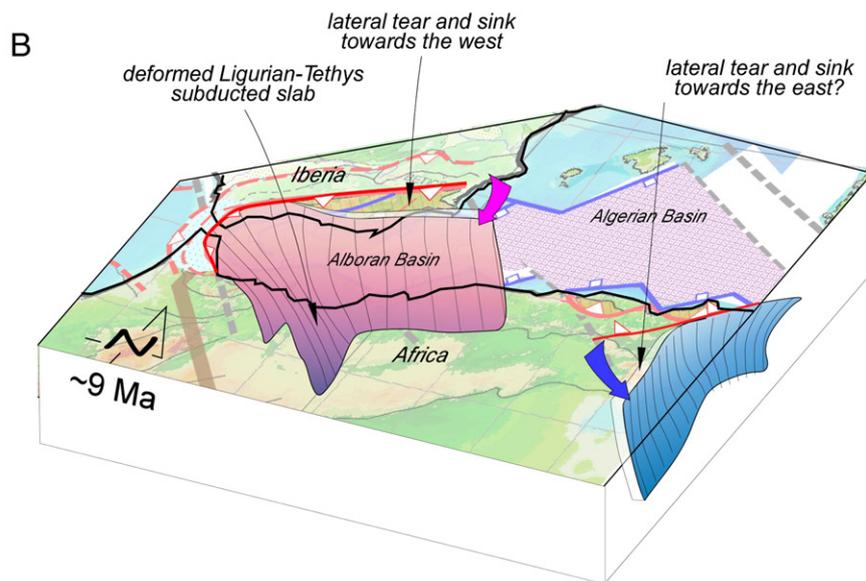
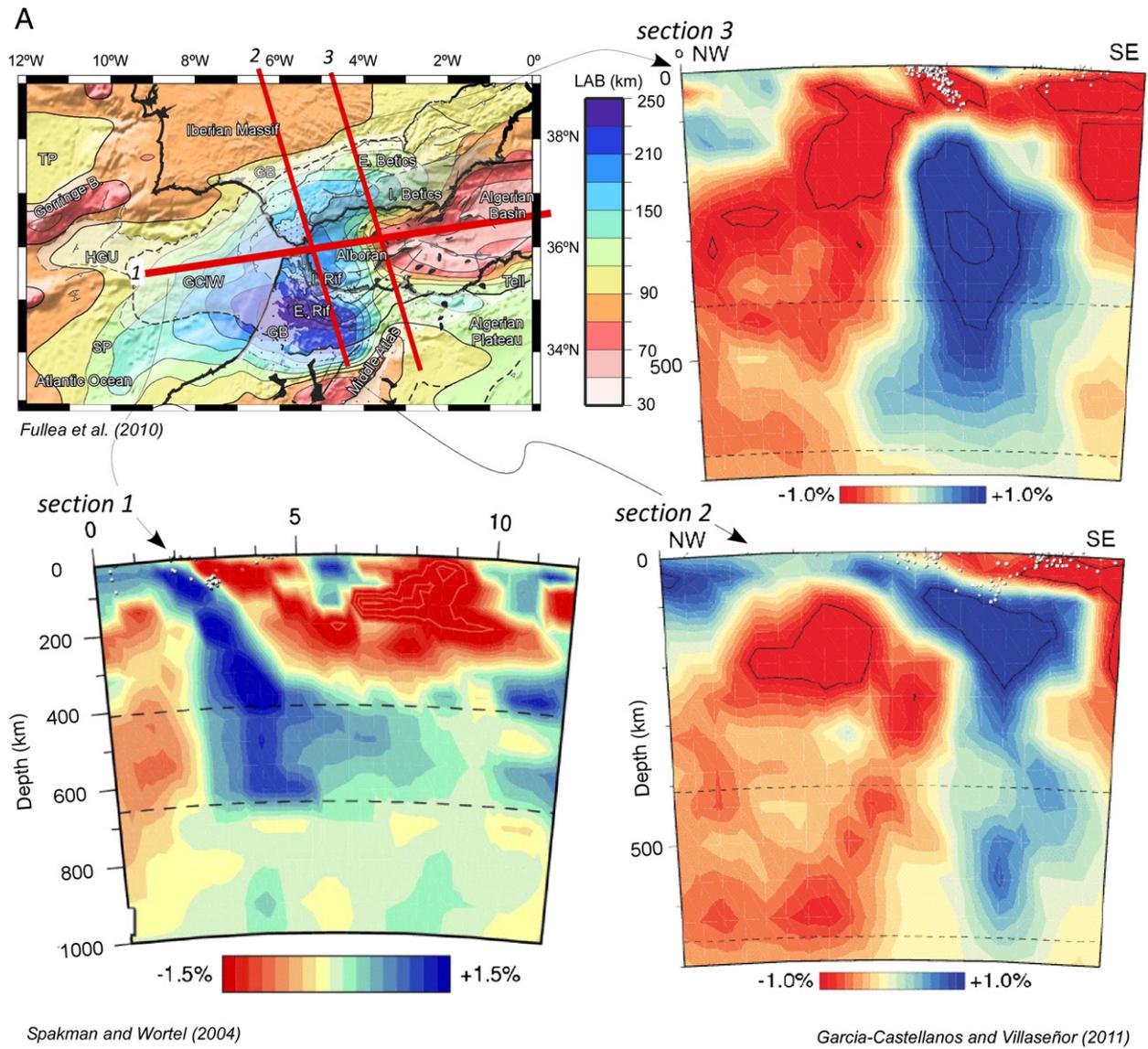


Fig. 17. A. Map of lithosphere thickness for the Betic-Rif system (Fullea et al., 2010; their Fig. 6B) with location of three seismic tomographic transects. These image the 3D geometry of the cold lithospheric slab hanging beneath the orogen, which is continuous along its westernmost part (section 1, Spakman and Wortel, 2004; their Fig. 2.6A), and along section 2, (García-Castellanos and Villaseñor, 2011, their Fig. S1-4), but is interrupted beneath the eastern and central Betic (section 3). B. Cartoon showing the geometry of the subducted lithospheric slab beneath the Betic-Rif system.

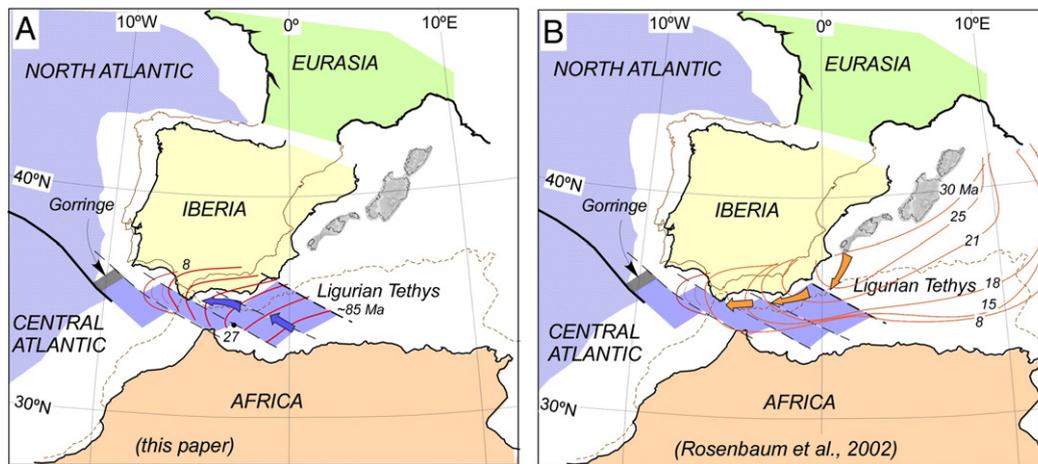


Fig. 18. A. Sketch map showing the progressive advance of the upper plate front above the initial SE-dipping and later E-dipping subducting Ligurian–Tethys slab, according to our kinematic model since Late Cretaceous. The change in the direction of subduction occurred at around 27 Ma, synchronous to the onset of Alboran basin formation. B. Similar sketch map showing the advance of the upper plate in the majority of the models reconstructing the western Mediterranean evolution during the Neogene (based on Rosenbaum et al., 2002b; their Figs. 11–18).

are not only referred to the path of the advancing subduction front but also on the timing of the subduction-related orogenic processes: Late Cretaceous in our model and late Oligocene in the currently accepted ones. Timing differences are also evident in the proposed evolution of the early Miocene Alboran basin. In our model the basin formed when the Betic–Rif orogenic front surpassed its Oligocene–Miocene position (pointed by a black circle in Fig. 18A). Therefore, the Alboran back-arc basin formed comparatively close to its present position whereas in some of the currently accepted models it traveled for several hundred kilometers after its initiation with very little internal tectonic distortion. Such proposed large displacement of the Alboran back-arc basin since its initiation in earliest Miocene (on top of the Alboran Block) would have produced and equivalent extension behind it, migrating and becoming progressively younger toward the west, which contrasts with the middle–late Miocene formation of the Alghero–Balearic basin (Booth-Rea et al., 2007). Last but not least, the commonly accepted model for the Betic–Rif orogenic system requires at least one major transform fault along the northern limit of the system to accommodate the differential displacement between the far-traveling Alboran Block (Internal Betics) and the External Betics that are considered the depositional sequences of the SE Iberian margin (see Leblanc and Olivier, 1984; Sanz de Galdeano, 2008 and references herein). This major orogenic transform fault would have propagated to the WSW and being younger in the same direction, which is not supported by field observations.

Paleomagnetic vertical axis rotations are also important to constrain the Betic–Rif orogenic system although they are still under debate. We briefly discuss the existing results (see compiling results in Chalouan et al., 2008; and Platt et al., 2003a) and compare them with proposed rotations from our reconstruction. Vertical axis rotations, mostly clockwise (CW) in the Betic segment and counter clockwise (CCW) in the Rif segment of the orogen, have been used to confirm the geodynamic model of a rigid Alboran Block producing the rotation of the Iberian and NW African cover units during its westward displacement. Our model also accounts for large-scale CW rotations of about 35–40° in the Betics and CCW rotations up to 80–100° in the Rif to acquire the final curved geometry of the orogenic system (Fig. 18). The calculated rotations seem difficult to reconcile with the already published results although the latter have a large variability that makes problematic their interpretation (e.g., Allerton et al., 1993; Fernández-Fernández et al., 2003; Osete et al., 2004; Platzman, 1992; Platzman and Lowrie, 1992; Platzman et al., 2000). Most of these results, were analyzed largely from sites in the Jurassic and Cretaceous carbonates of the Subbetic units close to the

boundary with the Internal Betics. These units record a long and composite geologic history encompassing Jurassic and Cretaceous rifting and then subduction-related orogenic processes since the Late Cretaceous and thus with foreseeable complex rotation and magnetization histories as documented in Villalaín et al. (1994). All these factors can be a strong handicap for the proper and unique interpretation of the paleomagnetic data. Furthermore, the results from 13 complete paleomagnetostratigraphic Neogene successions in the intramontane basins of both the Betics and Rif show no rotation after late Tortonian time (Krijgsman and Garcés, 2004). Conversely, Mattei et al. (2006) analyzed late Miocene and Pliocene intramontane basin deposits showing vertical axis rotation related to Africa–Iberia compression that would still produce further orogenic bending.

The distribution of volcanism has been also used to give support for specific geodynamic models. The model presented by Duggen et al. (2004 and 2005) concludes that the westward slab rollback caused continental-edge delamination of subcontinental lithosphere associated with the upwelling of a mantle plume contaminated with sublithospheric mantle. Therefore, calc-alkaline volcanism is derived from metasomatized subcontinental lithosphere whereas alkaline volcanism is derived from sublithospheric mantle contaminated with a plume material. Booth-Rea et al. (2007) give support to Duggen's model and propose that the westward migration of the Gibraltar accretionary wedge (some hundred km) gave rise to a magmatic arc with different crustal domains that from west to east are: thin continental crust modified by arc magmatism, magmatic-arc crust, and oceanic crust. There is a general agreement in considering the widespread calc-alkaline volcanism as related to subduction processes. However, the observed geochemical affinities of volcanic rocks are not restricted to the Alboran basin but extend to a much wider region including the whole Western Mediterranean, the Atlas Mountains and the Massif Central. Therefore the geodynamic interpretations claiming for the delamination of subcontinental lithosphere and mantle plume contamination proposed for the Alboran basin can admit other mechanisms. Although in our model we cannot rule out partial mantle delamination affecting the western Betics and Rif, its occurrence is not forced by the conceptual model.

6.3. Lateral change of subduction polarity and position of the Africa–Iberia paleo-suture

In our interpretation, the Betic–Rif segment of the Jurassic Ligurian–Tethys ocean is consumed by a SE-dipping subduction beneath Africa whereas further east, the former Algerian segment is consumed by a

NW-dipping subduction beneath Iberia. The lateral change of the subduction polarity is assumed to occur across the ~NW–SE trending paleo-transform fault joining the metamorphic complexes of the Betics and the Kabylies, in a direction subparallel to the Pierre Fallot Fault (or N-Balearic Fault; Fig. 2). The surface expression of this proposed transform fault would have been obliterated during the middle–late Miocene evolution of the Algero–Balearic oceanic basin (Booth-Rea et al., 2007) (Fig. 6). Changes in subduction polarities are common in narrow and fragmented oceanic and continental domains squeezed in between large converging continental plates. Although it is not the aim of the paper to make an exhaustive review of present and fossil examples of lateral changes in subduction polarity, we briefly document few cases from the Western Mediterranean and SE Asia.

The convergence of Iberia and Apulia toward Eurasia during Eocene–Oligocene times was solved with a lateral change in subduction polarity, that is: dipping northward beneath the Pyrenees and southward beneath the Alps (e.g., Handy et al., 2010; Stampfli and Borel, 2002; Stampfli et al., 1998). The southern end of the Neogene SE-dipping subduction of Europe beneath the Alps and the W-dipping subduction beneath the Apennines also implies a lateral change of subduction polarity along a transform fault at least during the Oligocene–Miocene times (Boccaletti et al., 1974; Molli, 2008; Rehault et al., 1984; Vignaroli et al., 2008) (Fig. 6). There are several subduction polarities in the intricate region of SE Asia, where multiple small continental blocks are separated by oceanic domains (e.g., Metcalfe, 2011; Pubellier et al., 2004; Villeneuve et al., 2010). The W-dipping Seram subduction is connected with the E- and S-dipping Celebes subduction along the E–W trending eastern segment of the Palu strike slip fault (e.g., Govers and Wortel, 2005) and the east-dipping North Moluccas subduction is directly merging to the large W-dipping Philippine subduction (Pubellier et al., 2008). At a larger scale the Pacific–Australian plate boundary also shows a change in subduction polarity along New Zealand (e.g., Lamb, 2011). To the north, the Pacific Plate is subducted along a NW-dipping plane whereas to the south the Australian Plate subducts along a SE-dipping plane. These two opposite subductions are linked by the Alpine fault of New Zealand, which is a major continental transform fault (Scholz et al., 1979). In summary, invoking a lateral change of subduction polarity in the westernmost Mediterranean is a plausible working hypothesis.

It is well accepted that the present-day Africa–Iberia plate boundary is a diffuse limit as indicated by the seismicity distribution in the area (e.g., Álvarez-Gómez et al., 2011 among others). According to our model however, the extinct plate boundary between Iberia and Africa would have been located between the External Betics forming part of the Iberian plate margin and the Malaguide Complex belonging to the African plate (this suture has been depicted by a slightly thicker discontinuous red line in the reconstructions in Fig. 9). In this scenario the tectonic stack of HP/LT metamorphic units represent the partially subducted highly stretched Ligurian–Tethys transitional/oceanized crust that was subsequently exhumed to shallow crustal levels between Africa and Iberia during the buildup of the Betic–Rif orogenic belt.

7. Conclusions

The new forward kinematic model at lithospheric scale for the Betic–Rif orogenic system proposes an initial slow SE-dipping subduction from Late Cretaceous to middle Oligocene that shifted to a faster E-dipping subduction since this period. These SE- and E-dipping subduction related orogenic scenarios can explain in a straightforward manner the spatial and temporal distributions of shortening along the external fold and thrust belt (External Betic–Rif), the burial, exhumation and exposure history of HP/LT Alpujarride and Nevado–Filabride metamorphic complexes (Internal Betic–Rif) and the opening of the Alboran back-arc basin along the inner part of the arcuate Betic–Rif system.

The initial configuration of the Betic–Rif orogenic system evolved in the SW limit of the highly segmented Ligurian–Tethys transitional/

oceanized crustal domain formed along the transtensional corridor connecting the Atlantic and the Tethys oceans during middle Jurassic. These segmented basins were probably separated from the Ligurian–Tethys Algerian oceanic basin by a NW–SE trending trench-to-trench paleo-transform fault separating two opposed subduction polarities (SE-dipping for the Betic–Rif and NW-dipping for the Algerian segments).

The slow initial SE-dipping subduction of the Ligurian–Tethys realm beneath the Malaguide upper plate unit since 85 Ma is sufficient to subduct Alpujarride and Nevado–Filabride rocks to few tens of kilometers of depth to produce HP/LT peak metamorphic conditions in middle Eocene times.

The late Oligocene (?)–early Miocene rather synchronous multiple crustal and subcrustal processes comprising the collision along the Betic front, the fast exhumation of the HP/LT metamorphic complexes, the opening of the Alboran basin, its flooring by HP Alpujarride rocks and its subsequent HT metamorphic event can be explicated by the fast NW- and W-directed roll-back of the Ligurian–Tethys subcrustal lithospheric slab (subducted crustal rocks were entirely exhumed at shallow crustal levels).

Adiabatic exhumation of HP metamorphic crustal units occurred first along the lower–upper plate subduction boundary and then by regional exhumation once tectonically stacked since early–middle Miocene onward.

The late westward migration of the Betic–Rif orogenic system up to the late Miocene possibly caused the lateral tearing of the subducted Ligurian–Tethys subcrustal slab.

During the last ~9–8 m.y. the evolution of the Betic–Rif system was again predominantly governed by the NW convergence of Africa against Eurasia. This convergence produced the squeezing of the Miocene Betic–Rif orogenic system by the impingement of Africa, clearly observed by the present seismicity distribution and by the SW escape of the Rif fold belt with respect to Africa. At subcrustal depths the large, steep and cold Ligurian–Tethys subcrustal lithospheric body was deformed by its own sinking dynamics though some tectonic squeezing cannot be disregarded.

The final step of our lithospheric model coarsely fits the present geometry of the subcrustal body beneath the Betic–Rif system imaged by seismicity, tomography, and numerical models based on potential fields.

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