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Extension in a convergent tectonic setting: a lithospheric view on the Alboran system of SW Europe

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ABSTRACT. The Betic Cordillera of southern Spain forms a clear example of a collisional orogen that has undergone large-scale late-orogenic extension while convergent motion of the bounding plates continued. The orogen provides a unique opportunity to study the tectonics of the system at different lithospheric levels. At shallow levels in the crust, fault-bounded intramontane basins, formed during the early to middle Miocene, contain coarse continental sediments heavily affected by normal faulting, followed by a less deformed late Miocene marine succession. Extension was accommodated by coeval shortening in thin-skinned fold-and-thrust belts in the periphery of the system, and much of the region has now subsided to form a large marine basin, the Alboran basin. The thermal and deformational record of these processes is preserved in rocks from deeper crustal levels in the internal zone of the Betic Cordillera. These rocks were metamorphosed down to 50 km depth and are now exposed beneath major low-angle detachment zones that separate them from heavily faulted low-grade rocks above. Cooling ages of associated mylonites indicate that these detachments were active during the early to middle Miocene. Peridotite massifs in the western Betics emplaced in the early Miocene provide coherent outcrops of subcontinental upper mantle that allow insight in coeval processes in the mantle lithosphere. The peridotites record evidence for exhumation in several stages from asthenospheric depths to the surface. Early stages of exhumation probably occurred during Mesozoic rifting. Cooling at mid-lithospheric depths possibly reflects early crustal thickening, followed by extension and subsequent heating. A sudden rise of ambient temperatures in the mantle rocks by about 400 °C suggests loss of most of the underlying lithosphere and ascent of asthenosphere, whilst the final stages of exhumation in early Miocene time reflect extensional collapse. All of these phenomena can be explained by some form of removal of the lithospheric root beneath a Paleogene collisional orogen, leading to large-scale extension followed by thermal subsidence of the center of the system. The processes inferred here for the Alboran region are in all likelihood not unique, as many similarities can be identified with the geology of the Tibetan Plateau, but also with domains in the Variscan and the Pan-African orogenic belts where extensional processes and associated LP/HT metamorphism and magmatism can be shown to equally have occurred in a convergent tectonic setting.

KEYWORDS: Late orogenic extension, Betic Cordillera, Alboran, convective removal

1. Introduction

The Betic Cordillera of southern Spain forms the northern branch of an arc-shaped mountain belt that continues across the Strait of Gibraltar in the Rif Mountains of Morocco, surrounding the western end of the Mediterranean or Alboran Sea (Fig. 1). The arc forms the western termination of the Alpine orogenic system of southern Europe and developed during, and partly in response to, late Mesozoic to Tertiary convergence between Africa and Iberia. It is difficult to locate a plate boundary between these two continents, however. Instead, the present-day seismicity is scattered over the whole region (Buforn and Udías, 1991). The arc and the Alboran Sea thus seem to form a domain of distributed intracontinental deformation comparable in some respects to those found elsewhere in the Alpine-Himalayan system.

The Betic Cordillera is commonly divided into an external and an internal zone. The external zone consists of nonmetamorphic Mesozoic and Tertiary sediments of the former southern margin of Iberia, strongly shortened by thin-skinned thrusting and folding during the Miocene (García-Hernandez et al., 1980; Banks and Warburton, 1991; Platt et al., 2003a). The internal or Betic Zone to the south is made up of Paleozoic and Mesozoic rocks, most of which have been penetratively deformed under a variety of metamorphic conditions and are now exposed in elongate ranges typically 15 to 30 km wide,

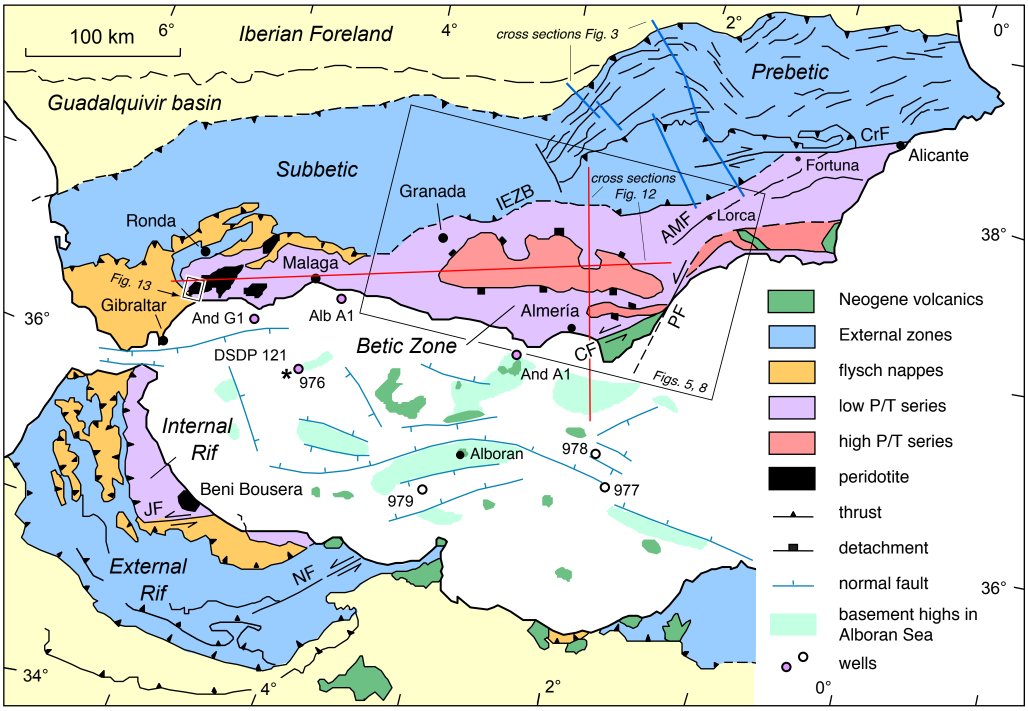

Figure 1. Sketch map of the Alboran Sea and surrounding chains, slightly modified after Vissers et al. (1995). High P/T series include glaucophane schist, eclogite, high-P greenschist and amphibolite facies rocks of Nevado Filabride Complex. Low P/T series include unmetamorphosed and low to intermediate P/T rocks of Internal Rif and the Alpujarride and Malaguide Complexes of the Betic Cordillera. Wells in Alboran Sea after Comas et al. (1999): star – DSDP 121, circles - Alb A1 (Alboran A1), Andalucia G1 (And G1), Andalucia A1 (And A1), nos. 976 to 979 represent ODP Leg 161 sites. Filled circles: wells in which basement has been drilled, open circles: no basement drilled. Abbreviations are AMF Alhama de Murcia Fault, CF Carboneras Fault, CrF Crevillente Fault, IEZB Internal External Zone boundary, JF Jehba Fault, NF Nekkor Fault, PF Palomares Fault.

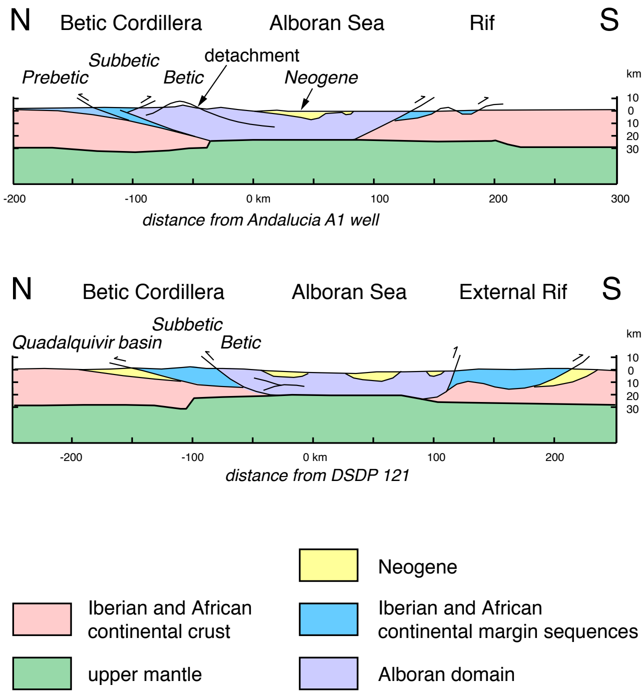

Figure 2. Cross sections across the Alboran Sea, after Watts et al. (1993) showing a markedly low crustal thickness around 20 km or less in the center of the Alboran system.

trending roughly parallel to the belt as a whole. These ranges are separated by intramontane basins showing a highly variable record of continental and marine sediments of Neogene and Quaternary age. In the western Betics, mantle peridotites occur tectonically interleaved among metamorphic rocks of crustal origin.

The internal or Betic Zone of the Betic Cordillera is only a small part of what is often called the Alboran Domain. The Alboran Domain comprises the internal zones of the Betic Cordillera and Rif and the crust underneath the Alboran Sea. The presently marine part of the Alboran Domain is characterized by thin (13-20 km) continental crust (Fig. 2; Banda et al., 1983; Watts et al., 1993; Comas et al., 1999) and a rugged subsea morphology. Some areas contain as much as 8 km of Neogene sediment beneath up to 1300 m of water (Watts et al., 1993), whereas the Alboran ridge is locally emergent and exposes volcanic rocks on Alboran Island. The region is characterized by elevated heat flow (Albert-Beltrán, 1979) and is underlain by anomalously low-velocity mantle (Banda et al., 1983; Blanco and Spakman, 1989; Spakman and Wortel, 2004). The area now occupied by the Alboran Sea was a marine basin during the Mesozoic and includes the sites of deposition of the continental margin sequences of the external Betic and Rif mountains, as well as the so-called flysch sequences that now occur as allochthonous thrust sheets between the internal and external zones mainly in the western part of the Betic-Rif arc. This marine basin, which may have been in part oceanic, closed progressively during the Paleogene, and by the end of the Oligocene it had been largely replaced by an emergent collisional orogen (Platt and Vissers, 1989). The present basins in the Alboran Sea, together with the onshore Neogene basins, were created by extension and subsidence in the early to middle Miocene (Watts et al., 1993), accompanied by mixed mafic to silicic volcanism (Hernandez et al., 1987). Until the end of the Tortonian, the internal Betic Cordillera would not have been obviously distinguishable from the rest of the Alboran Domain: the whole region consisted of elongate mountainous islands surrounded by marine basins. From Messinian time onward the Betic Cordillera became progressively uplifted, whereas the present Alboran Sea continued to subside. During this period the whole region was affected by a combination of strike-slip and compressional tectonics (Watts et al., 1993; Meininger and Vissers, 2006).

Classical studies of the Betic Cordillera focussed on the occurrence of nappe structures in the mainly metamorphic rocks of the Betic Zone and intuitively interpreted these structures in terms of crustal shortening (Brouwer, 1926; Fallot et al., 1960; Egeler and Simon, 1969). The Neogene basin sediments were considered to be posttectonic (e.g., Egeler and Simon, 1969). More recently, several authors have tried to interpret the Betic-Rif arc in terms of plate tectonic processes, involving the westward motion of an “Alboran plate” (comprising the internal zones of the Betic and Rif mountains together with the floor of the Alboran Sea) between Africa and Iberia (e.g., Andrieux et al., 1971; Leblanc and Olivier, 1984). These interpretations emphasized the importance of major strike-slip faults, shown in Fig. 1, such as the dextral Crevillente fault in the external Betic Cordillera (De Smet, 1984) and the sinistral Jebha and Nekkor faults in the Rif (Leblanc and Olivier, 1984). These ideas have been difficult to reconcile with the evidence that the Alboran Sea basin is young and extensional in origin. This has stimulated a variety of explanations, including the hypothesis of mantle diapirism developed by Van Bemmelen (1969) and Loomis (1975), which they associated with the emplacement of the Ronda and Beni Bousera peridotite massifs and the gravitational emplacement of the Betic nappes, and more recently the hypothesis of back arc spreading behind a westwardly migrating subduction zone (Torres-Roldán et al., 1986; Royden, 1993; Lonergan and White, 1997; Gutscher et al., 2002, Wortel and Spakman, 2000; Spakman and Wortel, 2004). On the basis of a growing data set of map, kinematic, seismic, and age data, the role of large-scale extension in the Neogene evolution of the Betic Zone and Alboran Sea has become widely recognized (Aldaya et al., 1984; Comas et al., 1992; Galindo-Zaldivar et al., 1989; García-Dueñas et al., 1988; Jabaloy et al., 1993; Platt et al., 1983; Platt, 1986; Platt and Behrmann, 1986; Watts et al., 1993, Martínez-Martínez and Azañón, 1997), and increasing emphasis has been placed on late orogenic extensional collapse as an explanation of these phenomena (Dewey, 1988; Platt and Vissers, 1989).

This paper summarizes previous and new stratigraphic, structural, and metamorphic data for each of the main structural levels exposed in the Betic Zone. We review in turn the shallow structure of the system, the partly metamorphosed (mid-crustal) rocks, and the peridotites of the western Betics which provide a record of the tectonic history of the upper mantle beneath the Alboran system, emphasizing structural and time relationships between these different levels in the Betic lithosphere. We then discuss the data in the light of current hypotheses for the tectonics of the Betic Cordillera and Alboran Sea.

2. Shallow tectonics of the Alboran system

In this section we inspect the main features of the Betic Alboran system at shallow crustal levels. We summarize the first-order characteristics of the External Zone, of the Internal or Betic Zone, and of the Alboran Sea floor, to arrive at geological constraints on geodynamic interpretations of the Alboran system.

2.1. Structure of the External Zone

In the eastern part of the Betics, the External Zone of the Betic Cordillera includes a shallow marine, carbonate-dominated platform to shelf sequence of the Prebetic Zone and, more internally, deeper marine rocks of the Subbetic Zone. These sediments of Mesozoic and Tertiary age, deposited on the former southern passive margin of Iberia, were strongly shortened by thin-skinned thrusting and folding during the Miocene (García-Hernandez et al., 1980; Banks and Warburton, 1991; Allerton et al., 1993; Platt et al. 2003a) as illustrated in Figs 3 and 4a. A Miocene age of thrusting is evidenced by the emplacement, in the Prebetic zone, of Mesozoic limestones on top of Tortonian sediments (Meijninger and Vissers, 2007), as well as by backthrusting of Subbetic Mesozoic rocks onto rocks of early Miocene age, in turn emplaced on rocks of the Betic Zone (Figs 3, 4b). In the northwestern part of the Lorca basin further east, these lower Miocene sediments, the backthrusted rocks of the External Zone, and rocks of the Internal Zone are unconformably overlapped by middle Miocene (upper Langhian and Serravallian) marine sediments (Geel & Roep, 1999; Meijninger and Vissers, 2006). This constrains the backthrusting to the early Miocene. Thrusting in the eastern part of the External Zone occurred in a NW to NNW direction, but at the scale of the entire Betic-Rif arc thrust directions vary from NW in the Betic Cordillera to W in the Gibraltar region and to WSW and S in the Rif (Platt et al. 2003a, and references therein).

2.2. Intramontane basins of the Betic Zone

The present surface of the Internal or Betic Zone is characterized by a “Basin and Range” type morphology, particularly in the east, with up to 3000 m high mountain ranges of mainly metamorphic rocks,

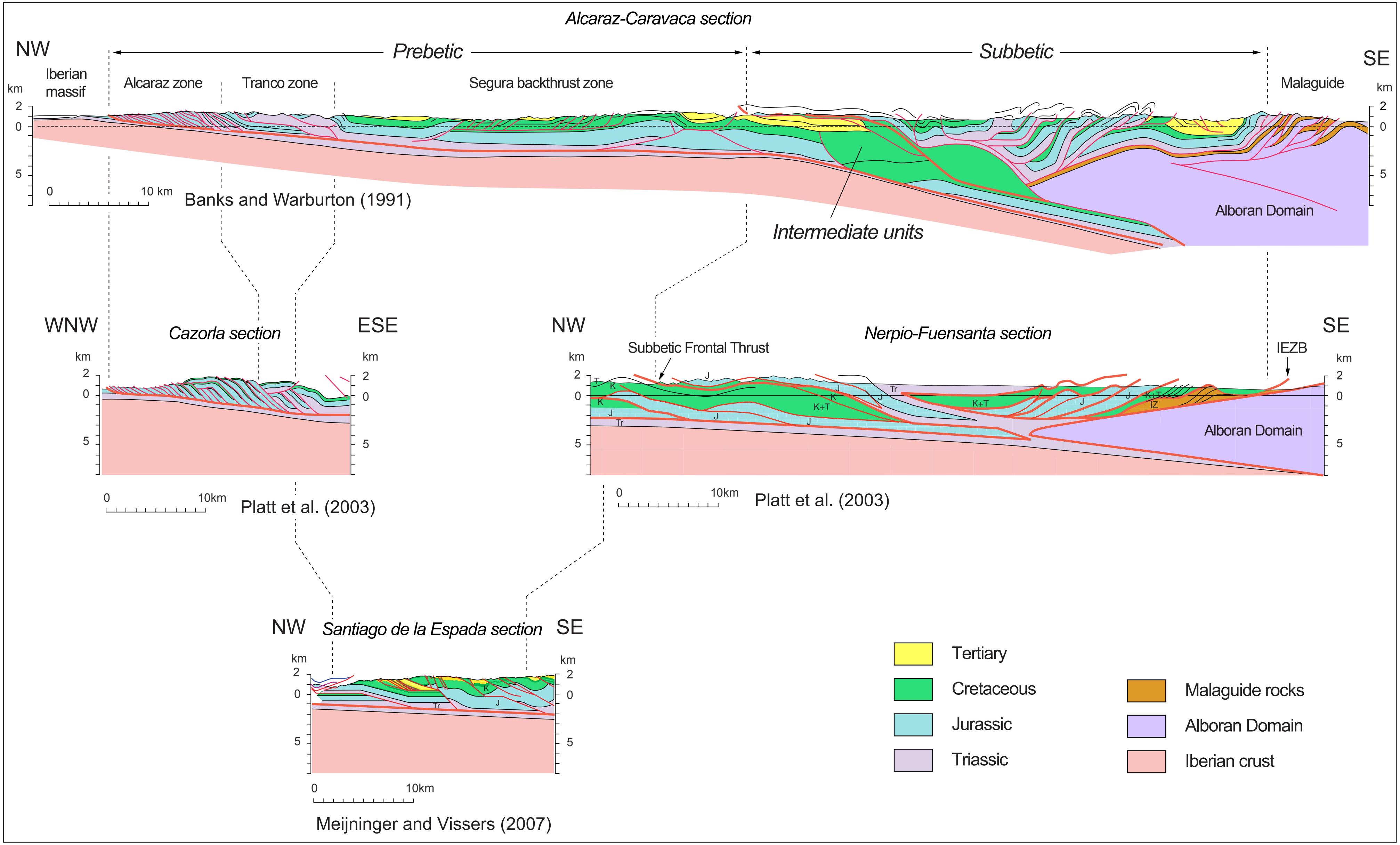

Figure 3. Cross sections across the Prebetic and Subbetic of the eastern Betic Cordillera illustrating its thrust belt structure with foreland (NW-) directed thrusts mainly in the Prebetic and external part of the Subbetic, and abundantly developed SE directed backthrusts near the Internal or Betic Zone. Modified after Banks & Warburton (1991), Platt et al. (2003a) and Meijninger & Vissers (2007).

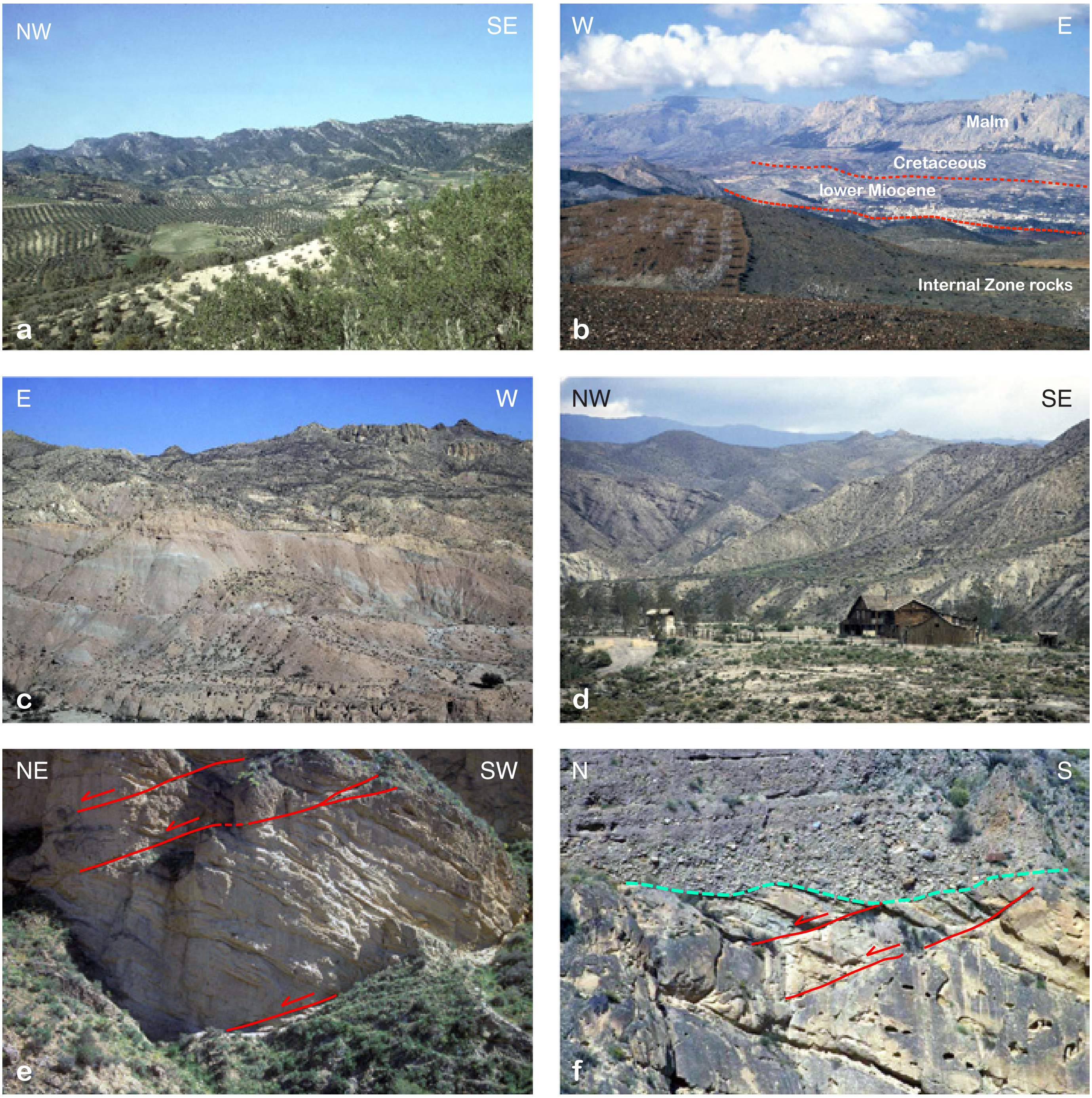

Figure 4. Field aspects of the Betic External Zone and of the Miocene basins in the Betic Zone of SE Spain. (a) Panoramic view of part of the Sierra de Cazorla looking approximately NE, showing repeated imbrication of Jurassic platform limestones in a NW directed thrust stack, see also Fig. 3. (b) Scenery around Velez Rubio looking NNW. An overturned sequence of Malm limestones and underlying Cretaceous rocks are backthrusted (i.e. moved towards the observer) onto poorly exposed lower Miocene sediments, in turn backthrusted against rock of the Internal Zone. (c) Field aspect of Langhian-Tortonian redbeds in the Huercal Overa basin passing upwards into yellow coloured lacustrine and shallow marine sediments, with carbonate reefs upper right (see also Fig. 6). (d) Tortonian marine massflow deposits, turbidites and marls near Tabernas. (e) Exposure of domino-style normal faults in Tortonian marine mass flow sediments east of Huercal Overa. Note relatively shallow dips of normal faults and associated backtilting of the bedding. Height of exposure about 30 m. (f) Exposure in cliff face south of Huercal Overa showing faulted and tilted Tortonian sediments unconformably overlain by Messinian conglomerate. Height of cliff face on photograph approximately 10 m.

separated by Neogene intramontane basins (Fig. 5). The stratigraphy of the basins, illustrated in Fig. 6, is somewhat variable (e.g., Sanz de Galdeano and Vera, 1992), but commonly includes mutually discordant, continental scree and debris flow deposits of probable middle and late Miocene age, deposited in narrow fault-bounded basins. These pass via fan-delta (Mora-Gluckstadt, 1993) and playa-lake sediments with local continental intercalations (Fig. 4c) into Tortonian sequences of marine reefs and submarine fans (Fig. 4d). In some of the basins, the Tortonian sediments are followed by Messinian conglomerates, reef limestones and evaporites, and by Pliocene marls (Fig. 6). All of these sediments are followed by late Pliocene to Recent alluvial fans. The onset of marine sedimentation in the basins seems to have occurred earlier (Burdigalian to Langhian) in the south; the transition to continental sedimentation at the end of the cycle took place earlier (late Tortonian) in the north, with a tendency to be also diachronous from ENE to WSW (Meijninger, 2006)

Many of the basins are markedly asymmetric, with strongly faulted southern margins and unconformable contacts with the metamorphic rocks on the northern margins. This suggests that these basins may have originated as tilted half grabens or be floored by tilted basement blocks. A seismic survey of the Granada basin by Morales et al. (1990) shows half graben structures at a 5 km scale. A detailed structural and sedimentological study of the Huercal-Overa basin by Mora-Gluckstadt (1993) and Meijninger (2006) demonstrates the existence of several pre-Tortonian half graben structures developed at a similar scale, within an overall half graben geometry of the basin as

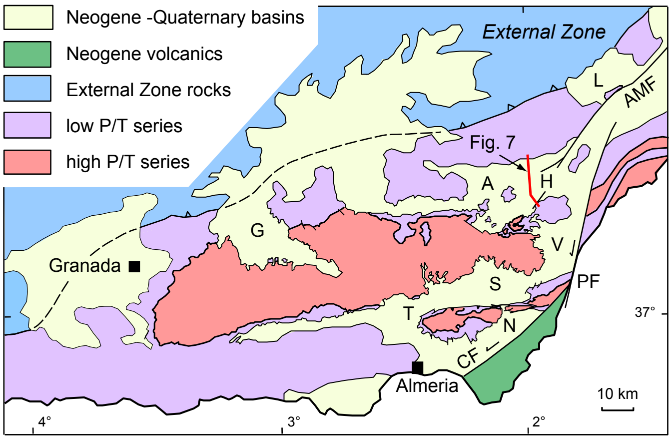

Figure 5. Sketch map of the central Betic Zone showing main locations of Miocene-Recent intramontane basins. Abbreviations are A Albox basin, G Guadix basin, H Huercal Overa basin, L Lorca basin, N Nijar basin, S Sorbas basin, T Tabernas basin, V Vera basin, other abbreviations as in Fig. 1. Line denotes section shown in Fig. 7.

a whole (Fig. 7). Similar structures have been imaged seismically in the Alboran Sea (Watts et al., 1993; Comas et al., 1999). A southward expansion of the stratigraphy occurs in the Sorbas basin towards the heavily faulted contact with the Sierra Alhamilla (Ott d'Estevou and Montenat, 1990), and a same tendency is seen in the Lorca basin, again suggesting a half graben origin of these basins.

Extension directions are about NE-SW in the Granada basin, NNE-SSW in the Huercal-Overa basin, and NE-SW in the Lorca and Fortuna basins. Limited exposures of continental scree breccias in the Huercal Overa basin affected by complicated geometries of successive normal faults suggest that extension may have been considerable (Vissers et al., 1995, Meijninger, 2006), with estimated values for the associated stretching between 50 and 100%. The younger marine sediments of Tortonian age commonly show steeply to moderately dipping normal faults only (Fig. 4e) and the entire succession is covered by an essentially flat-lying and undeformed Messinian conglomerate sealing moderately dipping normal faults in the Tortonian deposits underneath, as illustrated in Fig. 4f (Briend, 1981; Mora-Gluckstadt, 1993; Meijninger, 2006).

Montenat et al. (1987), de Larouzière et al. (1988) and Montenat and Ott d'Estevou (1990, 1999) have suggested that the Neogene basins in the eastern Betic Cordillera are related to a prominent set of N to NE trending sinistral strike-slip faults such as the Alhama de Murcia, Palomares and Carboneras Faults (Figs 1 and 5), which they interpret in terms of N-S Africa-Iberia convergence. The basins, however, are not, in fact, systematically located on releasing bends on these faults, and activity on the faults is essentially late Miocene to Quaternary, i.e., after most of the basins had ceased to be active depocenters. In addition, the areal extent of the basins and the amount of extension suggested by their internal structure appear to require significant crustal thinning during their formation. In a detailed study of the the prominent N to NE trending faults, Meijninger and Vissers (2006) conclude that the Messinian to Recent uplift of the Betic Cordillera as a whole was accompanied by sinistral strike-slip on N to NE trending faults, dextral strike-slip on W to WNW trending faults, and reverse motion on WSW to W trending structures. Some pre-existing horst blocks such as the Sierra Alhamilla were uplifted and thrust over the adjacent basins at this time, and the Neogene sediments were locally folded and overturned (Weijermars et al., 1985; Ott d'Estevou and Montenat, 1990).

In addition to the Neogene sediments, mafic, intermediate, and silicic volcanics of Neogene age occur scattered across the Alboran Domain and adjacent belts of the Betic-Rif system (Torres Roldán et al., 1986; Hernandez et al., 1987; Turner et al. 1999; Duggen et al., 2005). The earliest of these are early Miocene (22-23 Ma) basaltic dikes in the central and western Betics, whilst many of the middle Miocene and Pliocene calc-alkaline to ultrapotassic volcanics in the eastern Betics and eastern Rif seem spatially related to major strike-slip faults (e.g., Montenat and Ott d'Estevou, 1990).

2.3. Structures in the Alboran Sea

The Alboran Sea has been extensively explored during the nineties (Watts et al. 1993; Comas et al. 1999, and references therein) involving seismic studies as well as commercial and scientific drilling. The results of these studies summarized below have been reviewed in detail by Comas et al. (1999). The Alboran Sea has a complex seafloor morphology with several sub-basins, ridges, and seamounts, reaching a maximum depth of 2 km. The 180 km long Alboran Ridge, the most prominent northeast-southwest trending topographic feature of the Alboran Sea, emerges locally to form the volcanic Alboran Island. Multichannel seismic reflection (MCS) profiles show that the major sedimentary accumulation of up to 8 km thick is located in the western part of the Alboran Basin while further east the sediments are probably less than 4 km thick. Available data indicate that the acoustic basement beneath the Alboran Sea is heterogeneous, formed of either metamorphic or volcanic rocks. Metamorphic rocks of the Betic and Rif Chains have been recovered at the bottom of commercial wells offshore Spain and Leg 161 site 976 close to DSDP 121 (Fig. 1; Comas et al., 1999). East of 4° W, most of the residual highs appear to consist of volcanic rocks. As emphasized by Comas et al. (1999), the nature of the true basement in the eastern Alboran region is in fact still unknown, because volcanic edifices may overlie older sediments in places.

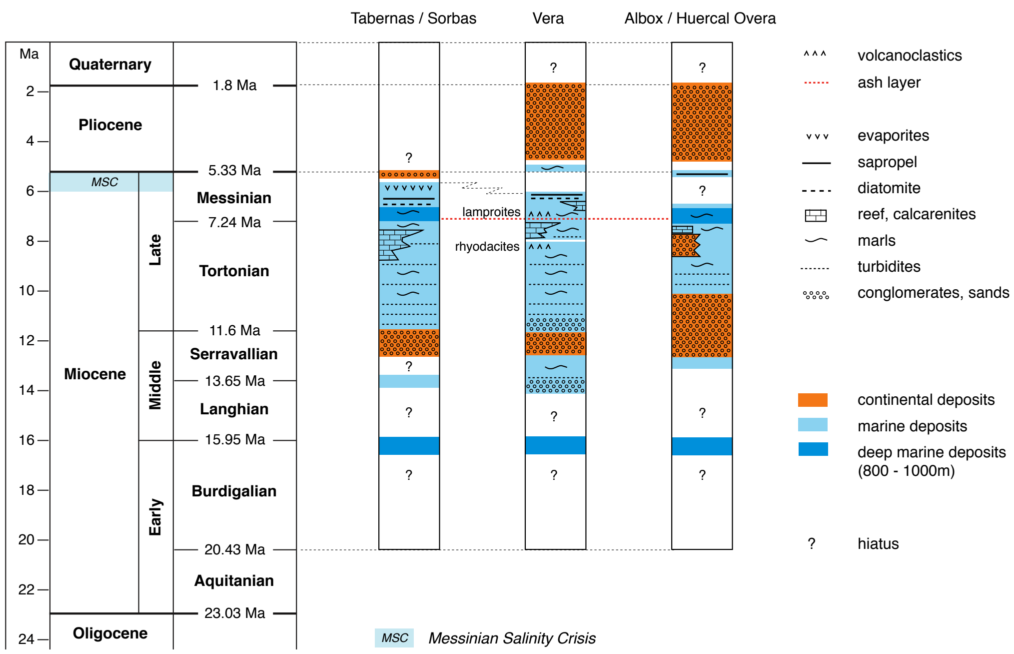

Figure 6. Stratigraphic correlation diagram showing main lithologies in the Tabernas / Sorbas, Vera and Albox / Huercal Overa basins. For locations of these basins see Fig. 5. Note that the stratigraphy is represented on a time scale, not a length scale, to emphasize correlations between coeval deposits as well as the hiatuses in the different basins. Time scale after Lourens et al. (2004).

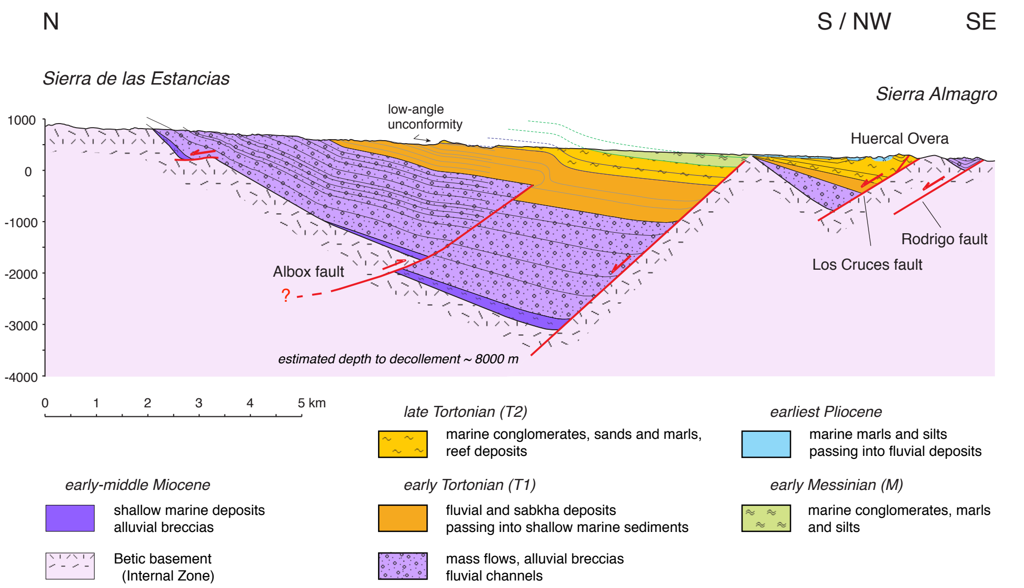

Figure 7. Cross-section of the Huercal Overa basin modified after Meijninger (2006) and Meijninger & Vissers (2006), showing a marked halfgraben geometry. Note thickening of stratigraphic units towards the bounding normal faults indicating synsedimentary stretching of the basin.

The structural pattern of the Alboran Basin is currently explained to result from two consecutive stages, including an early to early–late Miocene extensional stage, followed by a late Miocene to Holocene contractional stage (Comas et al., 1999). Multichannel seismic reflection profiles reveal both extensional and contractional structures. The dominant extensional structures include spectacular halfgraben geometries (Comas et al., 1992; Watts et al., 1993; Comas et al., 1999) at essentially the same scale as in the onshore basins such as, e.g., the Huercal Overa basin. Evidence exists for two stages of extensional faulting: i.e. an earlier Burdigalian–Langhian (~ 17–15 Ma) and a Serravallian–early Tortonian (~ 14–9 Ma) stage. Tectonic subsidence analyses (Watts et al., 1993; Docherty and Banda, 1992, 1995) estimate that rapid extension (initial rifting) continued during the early and middle Miocene, which is in line with the seismic data. Deep structures in the basement imaged from deep-seismic reflection profiles in the northern half of the Alboran Basin have been interpreted as extensional shear zones, presumably associated with major extensional detachments in the Betic Zone (Watts et al., 1993; Comas et al., 1997; see also Fig. 3, upper section).

The seismic studies suggest that extension was completed in the late Tortonian, and that a compressional regime has since affected the whole basin (e.g., Mauffret et al., 1987, 1992; Comas et al., 1992; Maldonado et al., 1992; Watts et al., 1993; Morel and Meghraoui, 1996; Chalouan et al., 1997). Shortening directions appear to have been similar to those described from the onshore regions (e.g., Woodside and Maldonado, 1992).

2.4. Main features of shallow tectonics in the Betic-Alboran system

The above data indicate that the fold-and-thrust belt structure of the External Zone developed during the Neogene well into Tortonian times. The youngest sediments involved in thrusting are Tortonian, while backthrusting must have occurred at the onset of the Langhian. While the External Zone became shortened during the Miocene, the Internal Zone shows abundant evidence for Miocene extension in the form of Neogene intramontane halfgraben basins in which the youngest stretched sediments are Tortonian, sealed in several basins by a Messinian unconformity. Likewise, the Alboran Sea shallow crust is characterized by a horst-and-graben structure filled with Neogene sediments. Seismic and well data indicate that the youngest stretched sediments again are Tortonian, unconformably overlain by Messinian sediments. Both in the onshore basins and in the Alboran Sea there is evidence for post Messinian inversion of earlier extensional structures.

It follows that for much of the Miocene, shortening in the periferal External Zone occurred about coeval with extension in the internal part of the system, and that by the end of the Miocene a compressional regime started to prevail, leading to folding, inversion on previous normal faults and the development of associated strike slip (tear) faults. Regional data suggest compression directions varying between NW-SE and N-S, consistent with both structural (Meijninger and Vissers (2006), seismological (Stich *et al.*, 2003 and Buforn *et al.*, 2004), plate motion (e.g., Dewey et al. 1989; DeMets *et al.*, 1994; Vissers and Meijer, submitted) and dynamic model studies (Negredo et al., 2002). As a result, some extensional grabens of the early and middle Miocene Alboran basin progressively emerged, to become continental basins during the Pliocene.

3. Critical observations in the crystalline crustal rocks of the Betic Zone

The Betic Zone is made up of a large number of tectonic units classically grouped into three tectono-metamorphic complexes (e.g., Torres Roldán, 1979), in ascending order, the Nevado-Filabride Complex, the Alpujarride Complex, and the Malaguide Complex. These complexes are distinguished on the basis of a variety of field criteria, including the lithological and metamorphic characteristics of the rocks involved, and strict definitions are impossible mainly due to lithological similarities between rocks of different complexes in combination with a varying metamorphic grade. Perhaps the most important characteristic is that wherever a contact between two of the complexes is preserved, there is almost invariably an abrupt decrease in metamorphic grade upward across it. This implies that these contacts are largely postmetamorphic and emphasizes the fact that there is an overall upward decrease in metamorphic grade. The Nevado-Filabride rocks are pervasively metamorphosed to high-greenschist or amphibolite facies. Alpujarride rocks for the most part show low metamorphic grade, although they locally reach upper amphibolite to granulite facies, in particular in the vicinity of the Ronda peridotite in the western Betics. The Malaguide rocks are virtually unmetamorphosed.

3.1. Nevado-Filabride Complex

The main exposure of Nevado-Filabride rocks is an elongate structural culmination, 125 km long and up to 35 km wide, making up the major part of the Sierra Nevada and Sierra de los Filabres (Figs 1, 8). In addition, Nevado-Filabride rocks crop out in the cores of the Alhamilla and Cabrera ranges to the south, and in the Sierra Almenara further east. They comprise more than 5 km of monotonous graphitic mica-schist and quartzite of probable pre-Permian age, locally intruded by Permian granite (Priem et al., 1966); a probably Permo-Triassic sequence of metamorphosed feldspathic sandstones, carbonates, and silicic meta-igneous rocks; and an association of metabasic rocks, serpentinite slivers and locally unserpentinized harzburgite (Morten and Puga, 1984), marble, and calcareous mica schist. The metabasic rocks are late Jurassic (Hebeda et al., 1980), and the calcareous schist locally contains poorly preserved microfossils of possible Cretaceous age (Tendero et al., 1993). This association of serpentinite slivers, occasional harzburgite, metabasic rocks, calcareous schist and marble has been interpreted as a disrupted ophiolite complex (Bodinier et al., 1987; Puga et al., 1989). The original stratigraphic sequence is repeated in a series of thrust sheets (Fallot et al., 1960; Nijhuis,

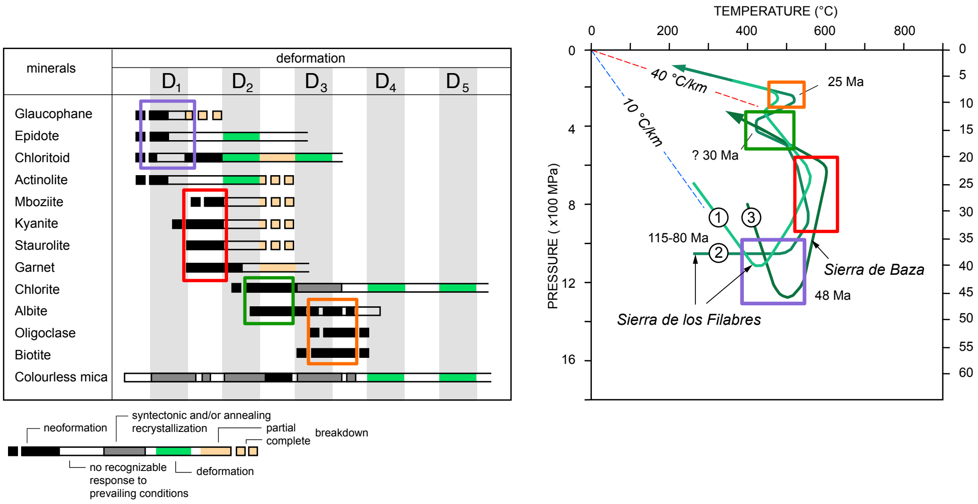

Figure 10. Diagram showing relationships observed in the central Sierra de los Filabres between deformational structures and mineral growth, deformation and breakdown, after Vissers (1981). Four paragenetic groups of minerals are identified indicated by rectangles. Diagram to the right shows interpretation of these parageneses in PT space. PT paths for the Nevado-Filabride Complex compiled after (1) Vissers (1981) and (2) Bakker et al. (1989) for the Sierra de los Filabres, and (3) Gomez Pugnaire & Fernando-Soler (1987) for the Sierra de Baza. Depth scale and geotherms are shown for average crustal density of 2800 kg/m^3.

1964) or fold-nappes (García-Dueñas et al., 1988), but the present great lateral extent of these units (several tens of kilometers) relative to their thickness (a few hundreds of meters) suggests that they have been considerably thinned and modified by subsequent deformation.

At deep levels in the Nevado-Filabride culmination the structure is dominated by a greenschist to amphibolite facies crenulation foliation, locally overprinted by later greenschist facies crenulations and associated foliations (e.g., Vissers, 1981; De Jong, 1991). The dominant crenulation foliation S2 is axial planar to commonly asymmetric F2 folds (Fig. 9a). Occasionally, interference is observed of such folds (labelled F2 in Fig. 9b) with an earlier generation of folds (F1). At high levels in the complex the structure is dominated by a strong subhorizontal foliation, commonly associated with isoclinal folds that may represent strongly attenuated F1 and/or F2 folds, and a stretching lineation that is warped around the regional culminations. This foliation intensifies into a mylonite zone along the upper boundary of the complex described in more detail below, and postdates most of the metamorphic events.

There is evidence in the higher tectonic units of an early HP/LT metamorphism, reflected by locally preserved glaucophane schist and eclogite (De Roever and Nijhuis, 1964; Puga, 1971; Puga and Diaz de Federico, 1978; Vissers, 1981; Gomez Pugnaire and Fernandez-Soler, 1987). An Ar-Ar plateau age on barroisitic amphibole of 48 Ma may date the end of this event (Monié et al., 1991). Structural analyses aimed at relating metamorphic and deformational histories (Langenberg, 1972; Vissers, 1981; Bakker et al., 1989; De Jong, 1991; Jabaloy, 1993) differ somewhat in their detail but mostly show that the high-pressure metamorphic stage is either earlier than or related to the earliest recognizable foliation S1 (Fig. 9c, 10), preserved in most cases as relics in S2 crenulation foliations and within early porphyroblasts. This was followed by thorough overprinting under upper greenschist facies conditions, rising locally to amphibolite facies in the higher tectonic units (De Roever and Nijhuis, 1964; Platt and Behrmann, 1986; Vissers, 1981; Bakker et al., 1989) as evidenced by porphyroblasts of garnet, and occasional kyanite and staurolite, that enclose the S1 schistosity preserved in the crenulations but predate the main dominant crenulation foliation (Fig. 9d, 10). Widespread growth of chlorite and albite mark a stage of extensive midgreenschist facies retrogression. The local growth of late staurolite, oligoclase, and biotite may indicate a late-stage thermal overprint (Nijhuis, 1964; Vissers, 1981; Bakker et al., 1989). PT paths proposed for the complex are shown in Fig. 10.

3.2. Alpujarride Complex

The Alpujarride Complex includes a presumably Paleozoic sequence of graphitic mica-schist and quartzite identical to that of the Nevado-Filabride Complex, Permo-Triassic aluminous phyllite and quartzite, and a thick sequence of Middle to Late Triassic dolomitized platform carbonates which locally enclose mafic intrusives of unknown age. No post-Triassic rocks have been found.

Much of the Alpujarride Complex only shows lower greenschist facies metamorphism, but occurrences of sodic amphibole (Sánchez-Vizcaíno et al., 1991) and carpholite (Goffé et al., 1989) suggest that ambient pressures during this low-grade event may have reached 7 kbar, corresponding to a moderately high PT ratio. Ar-Ar ages from white micas in some of the intensely deformed and recrystallized Alpujarride rocks yield ages close to 50 Ma, which is interpreted as the timing of a main contractional event and associated high P/T metamorphism (Platt et al., 2005). An Ar-Ar plateau age of 25 Ma on phengite from a vein carrying carpholite, chloritoid, and aragonite may date the end of this event (Monié et al., 1991). Locally, however, the grade is much higher, and several areas show staurolite-garnet-kyanite assemblages in metapelites, followed by growth of sillimanite and andalusite (Westra, 1969; Torres-Roldán, 1981). Upper amphibolite to granulite facies metamorphism at pressures up to about 10 kbar is widespread in Alpujarride rocks of the western Betics in the vicinity of the peridotite massifs, and is accompanied by small bodies of anatectic granite and migmatitic gneiss (Westerhof, 1977; Tubía, 1988; Argles et al., 1999). Narrow zones of garnet-alkali feldspar-sillimanite rock adjacent to the peridotite bodies show cordierite-spinel decompression rims around the garnets (Torres-Roldán, 1981). In the same area, metabasic rocks locally carry relict eclogite-facies assemblages suggesting peak pressures of the order of 15 kbar (Tubía and Gil Ibarguchi, 1991). These observations suggest an early medium to high PT ratio metamorphism followed by decompression at constant or rising temperature. Radiometric ages in the range 22-19

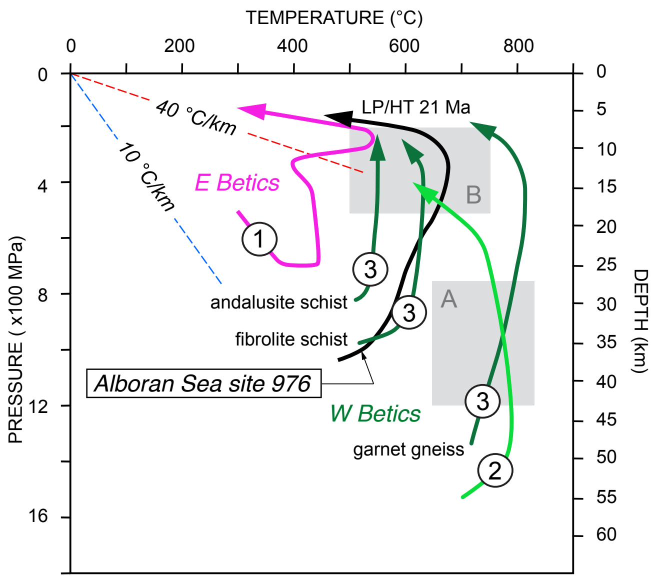

Figure 11. PT paths for rocks of the Alpujarride complex, after (1) Bakker et al. (1989) for the eastern Betics, and after (2) Tubía & Gil Ibarguchi (1991) and (3) Platt et al. (2003b) for the western Betics. PT path for site 976 in the Alboran Sea after Platt et al. (1998). PT fields for the western Betics after (A) Torres Roldan (1981) and (B) Westerhof (1977). Depth scale assumes average crustal density of 2800 kg/m^3.

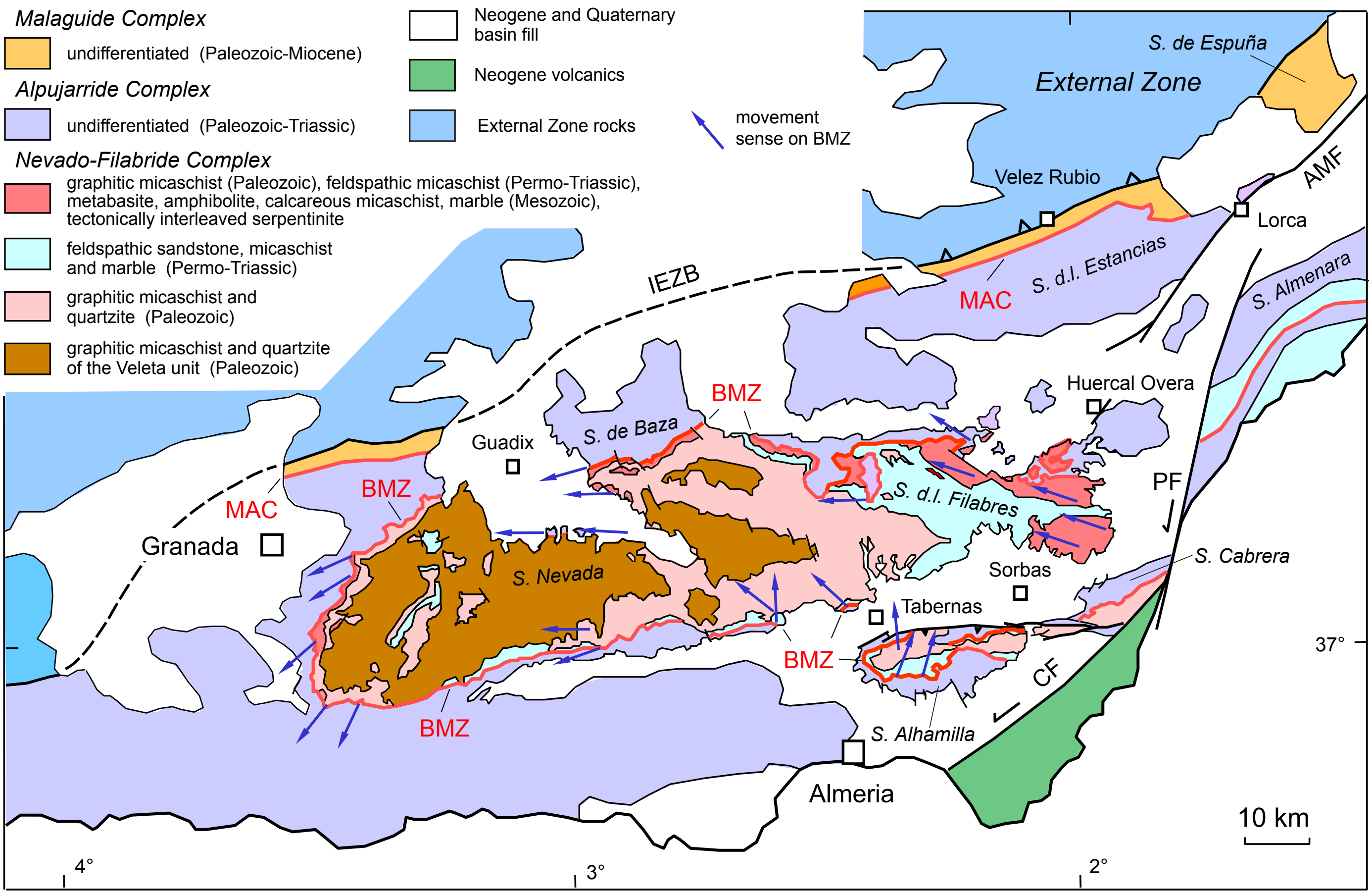

Figure 8. Structural map of the central Betic Zone (location of area in Fig. 1) showing the main Nevado-Filabride culmination, and kinematic data of the Betic Movement Zone. Abbreviations are BMZ Betic Movement Zone, MAC Malaguide/Alpujarride Contact, other abbreviations as in Fig. 1.

Ma (Zeck et al., 1989; Monié et al., 1991; Platt et al., 1998, 2005, Platt and Whitehouse, 1999) suggest that the thermal event occurred late in the tectonic evolution of the Betic Zone. By 21–22 Ma, exhumation had proceeded far enough for the highest grade and originally deepest rocks to cool below about 800ºC, and by 19 Ma exhumation of the high-grade rocks was largely complete (Platt and Whitehouse,1999). Exhumation and decompression occurred very rapidly, because in several areas Alpujarride rocks are overlain by early to middle Miocene sediments (Zeck et al., 1992). PT paths and PT estimates suggested for the Alpujarride Complex are shown in Fig. 11.

Alpujarride rocks show lithological repetitions that are at least in part due to thrusting. In the Sierra Alhamilla, early thrusts and associated folds are south vergent (Platt et al., 1983), but this observation has not been confirmed in other areas. Foliations and lineations associated with NE to NNE directed shear are widely reported from the complex and are commonly interpreted as being related to synmetamorphic thrusting (Tubía and Cuevas, 1986; Tubía et al., 1992). Tubía et al. (1992) suggested that the complex consists of three regionally developed units, the highest of which shows the highest grade and includes large volumes of Paleozoic rocks as well as the peridotite massifs, and the lowest of which shows only low-grade metamorphism and lacks Paleozoic rocks. This issue has been discussed by Vissers et al. (1995) who conclude that the structural pattern in the Alpujarride rocks has been strongly modified by late-orogenic extension, that the highest grade rocks have been exhumed from the greatest depths, and that, during extension, rocks at structurally high levels were affected in an irregular fashion by HT/LP metamorphism.

The complex shows considerable evidence for late extensional deformation. Torres-Roldán (1981) already pointed out that the rapid change in PT conditions within Alpujarride rocks adjacent to the Ronda peridotite requires tectonic thinning of the metamorphic zones. Platt et al. (2003b) suggested that this thinning of the crustal pile above the peridotites of the Ronda region occurred by slip on a series of extensional detachments. In the Sierra Alhamilla, Platt et al. (1983) identified the contact between low greenschist facies Permo-Triassic rocks above and amphibolite-facies Paleozoic rocks below as an extensional fault. In the Sierra de las Estancias (Fig. 8), there is a transition from ductile extension associated with a regional flat-lying foliation and ductile shear bands, to brittle normal faulting. This deformation was accompanied by patchily developed HT/LP metamorphism. The emerging picture of the Alpujarride Complex is that of an originally thick (>20 km) pile of thrust sheets that has been dissected by normal faults and reduced to an aggregate thickness of less than 10 km. HT/LP metamorphism accompanied extension.

3.3. Malaguide Complex

Exposures of the Malaguide Complex are largely limited to the western Betics north of Malaga and to a narrow zone along the internal-external zone boundary (IEZB, Figs 1, 8). It comprises a well-differentiated sequence of Paleozoic clastic and carbonate rocks, Permo-Triassic red-beds, Middle to Late Triassic carbonate rocks and evaporites, a patchily developed sequence of Jurassic to Eocene carbonate rocks, and, in the Sierra Espuña (Fig. 8), a remarkable sequence of foreland basin clastic sediments extending up to the middle Miocene (Lonergan, 1993). The Malaguide rocks show at best sub-greenschist facies metamorphism.

3.4. Major extensional structures in the Betic crust

Alpujarride-Nevado Filabride contact: Betic Movement Zone (BMZ)

The contact between the Nevado-Filabride and Alpujarride units is a major discontinuity involving an abrupt upward decrease of peak metamorphic pressures and a sudden increase in brittle deformation. The contact is decorated by a thick zone of mylonitic and cataclastic rocks indicating large-scale shear under conditions that evolved from ductile to brittle. This shear zone, identified and named as the Betic movement zone (BMZ) by Platt and Vissers (1980), can be traced for some 230 km in an E-W direction from Cartagena to the western end of the Sierra Nevada (Platt et al., 1984; Alvarez, 1987); it coincides with the Mecina Extensional System of Galindo-Zaldivar et al. (1989) and Jabaloy et al. (1992), and in part with the Filabres detachment of Martínez-Martínez and Azañón (1997).

The BMZ has been described in some detail in the Sierra Alhamilla (Platt and Behrmann, 1986), the Sierra de los Filabres (Vissers, 1981; De Jong, 1991), and in several areas of the Sierra Nevada (Martínez-

Martínez, 1986; Jabaloy et al., 1993) and shows the following characteristics (Vissers et al., 1995). The lower boundary of the BMZ is commonly diffuse: S2 crenulation foliations predominating at deeper levels in the Nevado-Filabride Complex gradually become intensified, whilst early (F1 and F2) folds show decreasing interlimb angles with isoclinal geometries close to the contact. Intensification of the foliation in strongly anisotropic micaceous schists is accentuated by the ubiquitous development of single, conjugate, or multiple sets of extensional crenulation cleavages (Platt and Vissers, 1980; Fig. 9e). Quartz-rich rocks become platy and show mylonitic (Fig. 9f) and, locally, ultramylonitic microstructures. Nearly everywhere close to the contact, the mylonite zone includes platy quartz-albite-chlorite schist, meter-scale bands of tourmaline-bearing quartz-feldspar mylonite, and bands of intensely folded and disrupted mylonitic marble. None of these rock types are unique to the BMZ, but the lithological variability indicates that a significant original thickness of rocks, mainly derived from the higher Nevado-Filabride units in the east, was stretched, thinned, and incorporated into the mylonite zone. In addition to small extensional faults bending into gouge layers parallel to the mylonites, larger normal faults in Alpujarride rocks of the hanging wall do in places sole down onto the BMZ (Vissers et al., 1995). The width of the mylonite zone, measured perpendicular to the platy foliation from the first detectable extensional crenulation cleavages to the contact, varies from a few hundred meters to about one kilometer. Ar-Ar dating by Monié et al. (1991) suggests that the mylonites cooled below 350°C at around 16 Ma, i.e., during the late Burdigalian, whilst apatite fission

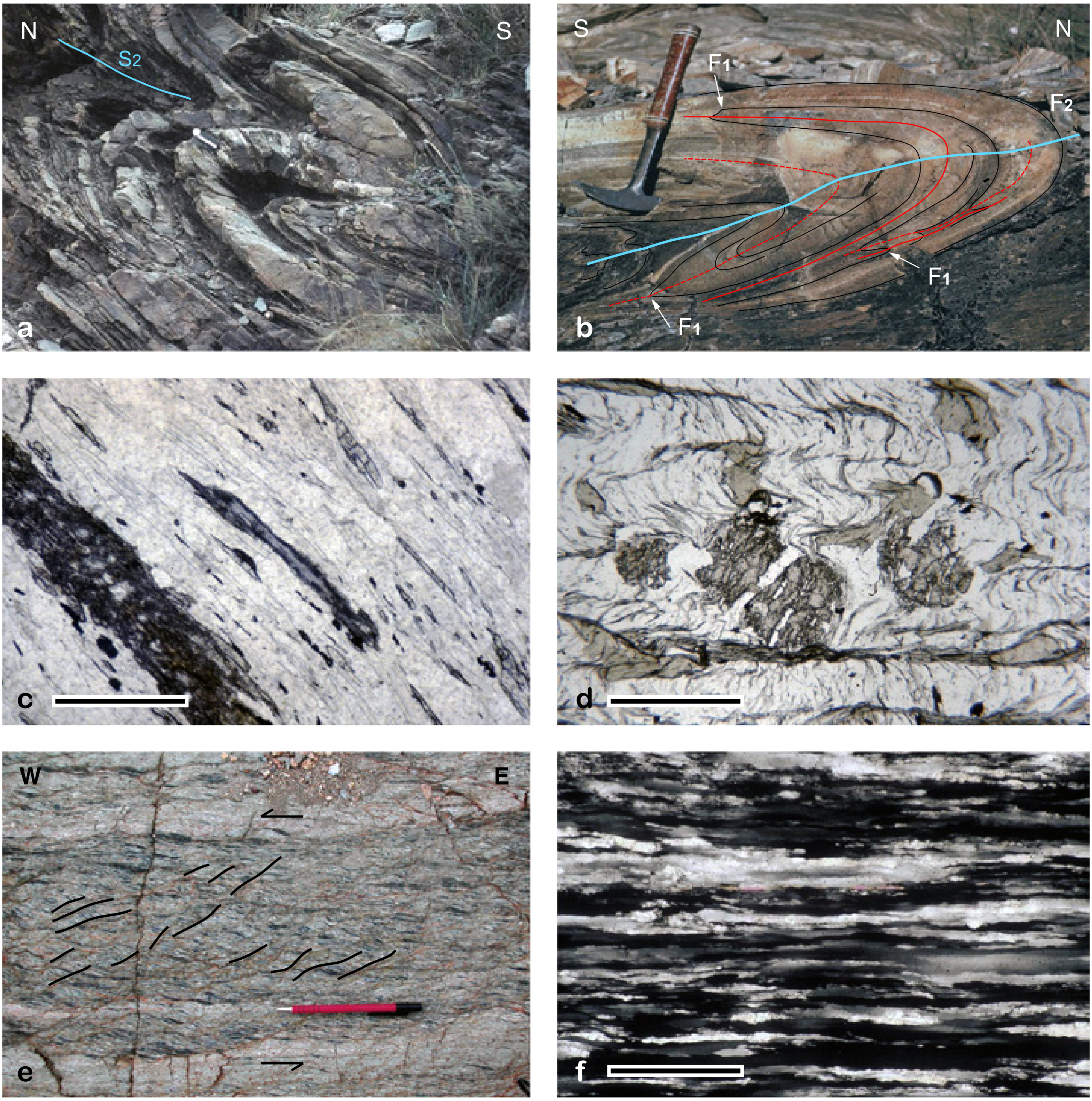

Figure 9. Field aspects and microstructures in rocks of the Betic Zone. (a) Exposure of asymmetric second-generation (F2) fold in dark graphitic micaschist and quartzites of the central Sierra de los Filabres north of Tabernas. (b) Interference structure of first (F1) and second generation (F2) folds in Paleozoic graphitic micaschist and quartzites of the Sierra Nevada south of Guadix. (c) Crystals of glaucophane oriented parallel to the S1 schistosity in Permo-Triassic micaschist, central Sierra de los Filabres. Scale bar 1 mm. (d) Microstructure of garnet chlorite micaschist from graphitic micaschist and quartzites of the central Sierra de los Filabres north of Tabernas. Note straight internal (S1) fabric enclosed in garnets continuous with crenulated microstructure of the matrix dominated by an S2 crenulation cleavage. Scale bar 1 mm. (e) Field aspect of intensely sheared micaschists from the Betic Movement Zone (BMZ) south of Guadix, viewed to the north. The exposure shows ubiquitous west-dipping extensional crenulation cleavage planes indicating a sinistral (top-to-the-west) sense of motion. (f) Microstructure of intensely defomed quartzite from same locality as in (e) marked by extremely streched quartz grains and very limited fine recrystallized material suggesting deformation at relatively low temperatures. Crossed nicols, scale bar 1 mm.

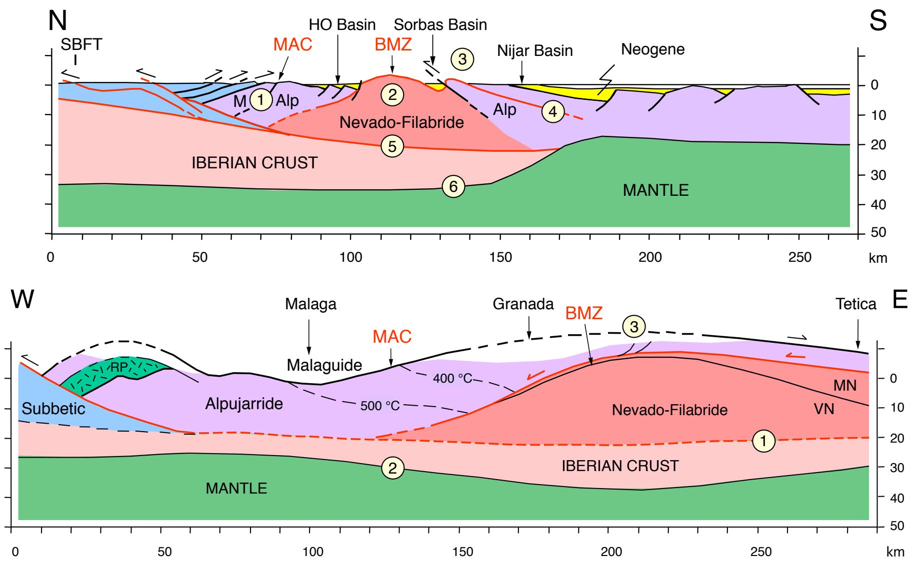

Figure 12. Schematic crustal sections across the central Betic Cordillera and neighbouring regions, slightly modified after Vissers et al. (1995). Location of sections shown in Fig. 1. (a) N-S section. Note that some features shown are not precisely on the line of section. Subbetic sector after Allerton et al., (1993), Alboran Sea sector after Watts et al. (1993). Abbreviations are Alp Alpujarride, M Malaguide, HO basin Huercal-Overa basin, SBFT Subbetic frontal thrust, other abbreviations as in Fig. 8. Notes: (1) Malaguide moved ENE into plane of section relative to the Alpujarride along the MAC, (2) Alpujarride moved W out of the plane of section relative to the Nevado-Filabride along the BMZ, (3) BMZ was folded and thrust in the late Miocene in the Sierra Alhamilla (Platt et al., 1983), (4) Reflector visible in offshore seismic line may be BMZ (Watts et al., 1993), (5) Position of top of underthrust Iberian crust not certain, (6) Moho geometry after Watts et al. (1993). (b) Schematic E-W crustal section along the Betic Cordillera. Abbreviations as in (a) with MN Mulhacen nappe, VN Veleta nappe, RP Ronda peridotite. Notes: (1) Betic rocks moved roughly NW to WNW relative to Iberian crust, (2) Moho depth after Banda et al. (1983), (3) Normal faults in Sierra de Baza sole down onto BMZ.

track ages indicate that the underlying Nevado-Filabride rocks cooled below ~70-100°C between 12 and 9 Ma, suggesting exhumation during the Tortonian (Johnson et al., 1997).

The significance of the BMZ is not unambiguous. In view of the duplication of Paleozoic and Mesozoic strata, the original contact between the Alpujarride and Nevado-Filabride Complexes must have been contractional, and the present contact has classically been considered a thrust (e.g., Fallot et al., 1960; Egeler and Simon, 1969). It is now accepted to be extensional over its entire trajectory for reasons reviewed in detail by Vissers et al. (1995), including a sudden upward decrease in metamorphic grade, excision of the higher Nevado-Filabride units in the direction of transport, and its evolution from ductile to brittle. Kinematic data from the BMZ mylonites (Fig. 8) show an arcuate pattern of transport directions, swinging from around N in the Sierra Alhamilla and southeastern Sierra Nevada, to WNW in the central Sierra de los Filabres, via W in the northern Sierra Nevada to SW in the western Sierra Nevada. These observations are consistent with the results of a similar study by Jabaloy et al. (1993). It follows that the dominant sense of shear on the BMZ is westward. The only possible correlation of rocks across the contact is between the amphibolite facies marble, granitic gneiss, schist, and metabasic rocks of the higher Nevado-Filabride units in the eastern Sierra de los Filabres with a similar lithology in the Alpujarride of the Ojen region in the western Betics. This would imply a displacement of more than 100 km (Fig. 12, E-W section). Given the evidence for substantial extension in both Nevado-Filabride and Alpujarride complexes and the evidence described above for variations in shear direction in both space and time, it may be better to view the BMZ as a detachment horizon between two crustal levels that were deforming to some extent independently, and that has itself been substantially extended.

Malaguide/Alpujarride Contact (MAC)

The contact between the Alpujarride Complex and the overlying Malaguide Complex in the Velez Rubio region (Fig. 8) is marked by a steeply north dipping zone, up to 20 m thick, of calc-mylonite derived from Alpujarride carbonate rocks, overlain by fault gouge derived from Malaguide Paleozoic greywacke. The steep orientation is a consequence of Neogene tilting along the adjacent internal-external zone boundary (Lonergan et al., 1994; Lonergan and Platt, 1995). The fault gouge generally shows a chaotic pattern of lineations indicating reactivation in several phases of motion. The calc-mylonite is extremely fine-grained and contains abundant clasts of limestone and dolomite, many of which show fragmented or recrystallized tails. The stretching lineation in the mylonites is subhorizontal, trending ENE, and the sense of shear is consistently dextral (i.e., top-to-the-east-northeast). Related deformation can be traced downward and southward into the underlying Alpujarride phyllites and quartzites, where it is represented by a foliation, ENE trending stretching lineation, and abundant 10-cm-scale shear bands with a top-to-the-east-northeast shear sense (Platzman and Platt, 2004). The shear bands range from ductile-type shear bands which contribute to the intensification of the foliation and lineation, to more discrete brittle shear bands marked by fiber-lineated quartz veins and striated brittle fracture surfaces. These structures overprint earlier folds and ductile fabrics in the Alpujarride rocks and show a consistent orientation and shear sense.

Like the Alpujarride/Nevado-Filabride contact, the Malaguide/Alpujarride contact must have originated as a contractional structure, but several lines of evidence suggest that at least in the Velez-Rubio area, the contact has been reactivated as a normal fault. Arguments are similar to those in favour of an extensional nature of the BMZ (see also Platt, 1986; Aldaya et al., 1991; Vissers et al., 1995) of which a marked decrease upward in metamorphic grade is the most obvious: everywhere along the contact, Malaguide rocks showing at most sub-greenschist facies metamorphism are in direct contact with greenschist facies Alpujarride phyllites. Illite crystallinity data from the Sierra-Espuña suggest that this difference in grade corresponds to a temperature contrast of about 100°C (Lonergan, 1993). A significant pressure difference is also likely because the Alpujarride rocks in the northern Sierra de las Estancias locally carry carpholite pseudomorphs (Goffé et al., 1989), indicating pressures of about 7 kbar. Apatite fission track ages suggest exhumation along this contact at about 23 Ma (Johnson, 1993). Together with the distorted shape of the Malaguide/Alpujarride contact, these data suggest that extensional motion on the MAC may have ceased somewhat earlier than on the BMZ. The opposing sense of shear on the two contacts contributes to a picture of large-scale, roughly coaxial horizontal extension in the Betic crust (see E-W section, Fig. 12).

3.5. Summary of main features in the Betic crust

The Nevado-Filabride rocks show a multistage metamorphic history from early high-pressure to lower pressure conditions. Adjacent Alpujarride units in the central and eastern parts of the Betic Zone are quite low-grade, but locally show low-pressure high-temperature facies series. The two complexes are separated by a major crustal-scale shear zone (BMZ) surrounding a 100 km scale elongate structural dome. Rocks below and above this shear zone show a distinct pressure gap indicating that the BMZ should be an extensional detachment. Transport directions vary between N and NW directed in the east to W and SW directed in the west, indicating overall E-W extension in Betic middle crust. Geochronology indicates a Miocene age of low-greenschist facies ductile shearing at around 16 Ma, while apatite fission track ages suggest exhumation between 12 and 9 Ma.

The contact of the Malaguide complex with the underlying Alpujarride rocks (MAC) is marked by a distinct difference in metamorphic grade, with almost non-metamorphic Malaguide rocks overlying Alpujarride greenschist facies phyllites and micaschists along a mylonitic contact that evolved from ductile to brittle. Geochronology suggests Miocene exhumation along the MAC at 23 Ma, hence extension on this contact

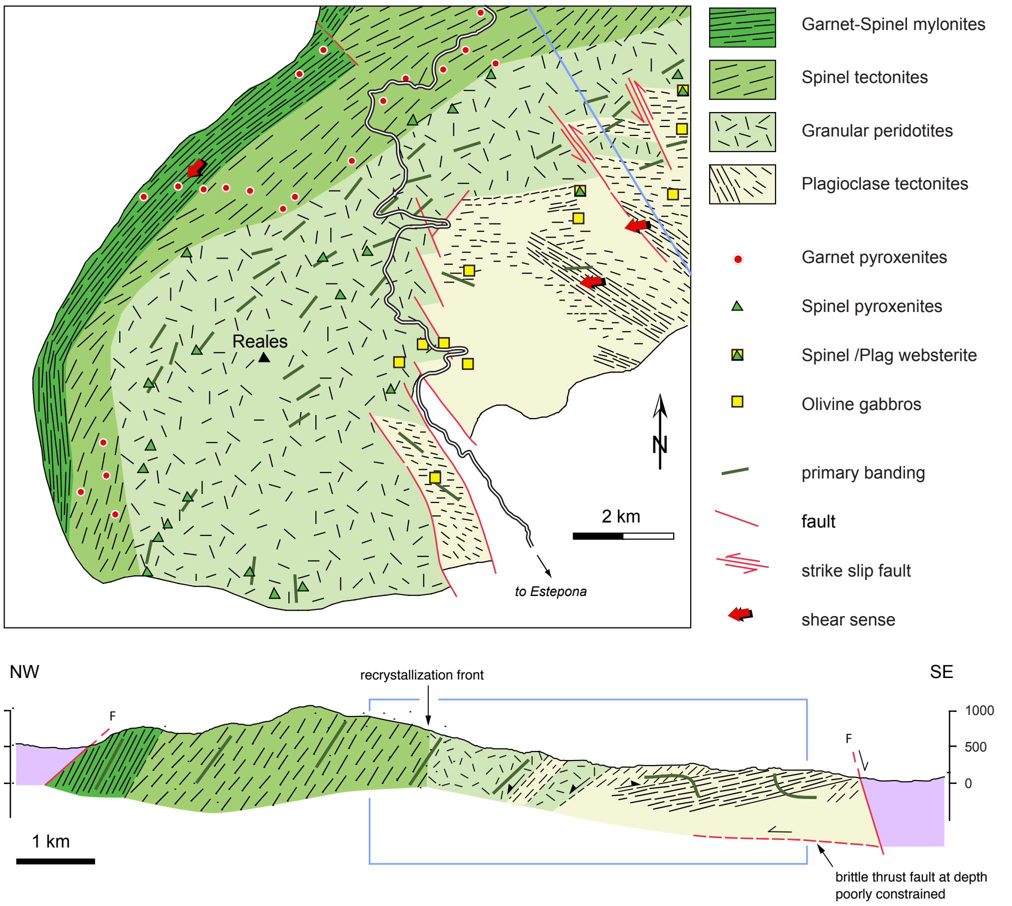

Figure 13. Structural map and cross section of the Ronda peridotite of the western Sierra Bermeja (location of area shown in Figure 1). Note that cross section is partly outside the map area, and is presented at a larger scale than the map, to allow more detail. Domain in rectangle represents part of the section seen along the (blue) section line in the northeast corner of the map. For further explanation see text.

may have ceased earlier than on the BMZ.

4. The Alboran upper mantle: the Ronda peridotite record

In the western Betic Zone and internal Rif of northern Morocco, upper mantle peridotites with Neogene emplacement ages (Priem et al., 1979; Zindler et al., 1983) are exposed amidst mostly high-grade crustal rocks (Figs 1 and 12). Graphite pseudomorphs after diamond in these peridotites (Pearson et al., 1989; Davies et al., 1993) indicate a deep lithospheric or asthenospheric origin. In addition, several of the peridotites show a structural and compositional heterogeneity reflecting successive stages of uplift and eventual emplacement. The Neogene emplacement ages suggest that this heterogeneity results at least partly from geologically young processes in the western Mediterranean mantle.

A marked feature of the Ronda peridotite body is its roughly concentric pattern of all three peridotite facies (i.e., garnet-, spinel- and plagioclase-peridotite facies (Obata, 1980), in an essentially coherent outcrop of some 300 km^2. A structural study of this massif (van der Wal, 1993; van der Wal and Vissers, 1993, 1996) shows that the three facies domains are related to and coincide with the progressive development of three different structural domains (Fig. 13). These are (1) foliated garnet-bearing peridotites in the northwestern part of the massif, (2) coarse-grained granular spinel peridotites in the central part, and (3) porphyroclastic plagioclase tectonites in the southern and eastern part. Note that the term porhyroclastic is used here as in mantle geology studies to describe a coarse-grained, foliated mantle peridotite containing deformed and partly recrystallized grains of the constituent minerals such as olivines and pyroxenes in a somewhat finer-grained matrix of these same minerals.

The garnet-bearing peridotites of the northwestern part of the massif (Fig. 13) include foliated Ariegite-facies spinel lherzolites (i.e., spinel lherzolites enclosing layers of garnet pyroxenite), referred to as spinel tectonites (Fig. 14a). They represent the oldest deformational structure preserved (van der Wal, 1993). Their microstructure is dominated by elongate olivines, elongate orthopyroxene porphyroclasts surrounded by recrystallized orthopyroxenes and clinopyroxenes (Fig. 14b), and oriented trains of spinel. Porphyroclast core compositions indicate high temperatures (1110±65 °C, labelled R1) at spinel-peridotite facies conditions, while the mineral chemistry of the recrystallized pyroxenes suggests dynamic recrystallization at much lower temperatures (810-900 °C, labelled R2) consistent with progressive cooling during deformation (Fig. 15).

Towards the margin of the peridotite body, the tectonite foliation bends over low angles into a mylonitic foliation of garnet-spinel mylonites (Figs 13, 14c) defining a mylonitic shear zone of a few hundred meters width (see also Precigout et al., 2007). A few similar but much narrower mylonite zones occur within the massif. The mylonites contain extremely stretched clasts of orthopyroxene (Fig.14d), elongate spinel, and spinel rimmed by a fine-grained kelyphitic assemblage with bulk compositions of a pyropic garnet, suggesting deformation near the transition from spinel to garnet peridotite (18-22 kbar, Fig. 15). This is supported by the stable occurrence, in the mylonitic matrix, of garnets derived from extremely stretched garnet-pyroxenite layers. The mineral chemistry of recrystallized orthopyroxene grains suggests temperatures around 830-880°C (labelled R3 in Fig. 15) during mylonitization.

Towards the core of the massif, the porphyroclastic microstructures as well as occasional narrow mylonite zones are overprinted by intense recrystallization and development of granular peridotites (Fig. 14e). The boundary between the spinel tectonites and the granular rocks, currently known as the Ronda recrystallization front (van der Wal & Vissers, 1993, 1996), lies oblique to the trend of the spinel tectonite foliation (Fig. 13). The boundary is not sharp but occurs gradually over about 200 m. In sections across this transition, orientations of pyroxenite dykes, compositional banding, and olivine petrofabrics seen in the spinel tectonites remain essentially constant, whilst the microstructure progressively evolves, via strain-free olivine microstructures (Fig. 14f) enclosing elongate pyroxenes and trains of spinel, to a coarse granular peridotite. This structural continuity and the evidence for concurrent growth of olivine, pyroxene, and spinel suggest that the granular rocks developed as the result of extensive recrystallization affecting the spinel-bearing tectonites and local garnet-spinel mylonites. Mineral chemistry based thermobarometry by van der Wal et al. (1993) suggests temperatures in the granular domain of 1090±70 °C (labelled R5 in Fig. 15). The associated pressures were inferred from the progressive transformation of garnet pyroxenites at

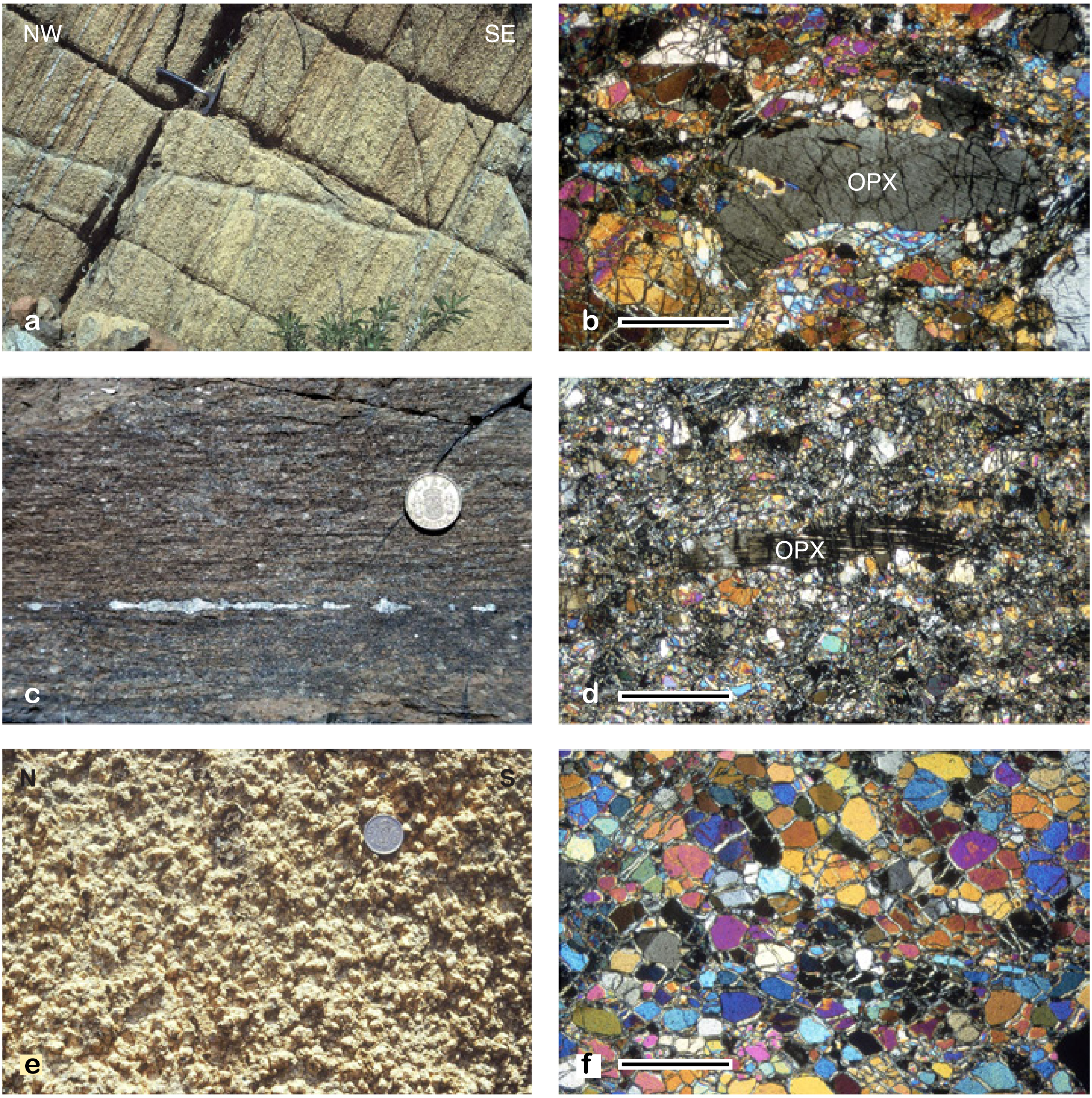

Figure 14. Structures and microstructures in the Ronda peridotite of the western Sierra Bermeja. (a) Field aspect of spinel tectonite foliation with garnet pyroxenite layers parallel to foliation, note hammer for scale. (b) Photomicrograph of spinel tectonite microstructure showing elongate orthopyroxene (OPX) dispersed in olivine matrix. Crossed nicols, scale bar 2 mm. (c) Field aspect of garnet-spinel mylonites with strongly boudinaged garnet pyroxenite bands, diameter of coin 2 cm. (d) Photomicrograph of garnet-spinel mylonite microstructure, showing elongate stretched orthopyroxene (OPX) in fine grained mylonitic matrix. Crossed nicols, scale bar 2 mm. (e) Field aspect of granular peridotite, diameter of coin 2 cm. (f) Annealed olivine microstructure from transition zone between sheared spinel tectonites and granular peridotites. Crossed nicols, scale bar 2 mm.

the transition into spinel pyroxenites (Seiland subfacies, i.e., spinel lherzolites enclosing spinel pyroxenites), indicating substantial exhumation to pressures of 8-14 kbar (Fig. 15). Van der Wal (1993) and van der Wal and Vissers (1993) inferred that exhumation of the Ronda rocks proceeded under relatively cool conditions between 770 and 880 °C (shown in Fig. 15 as path towards conditions labelled R4), after which the outer and presumably higher part of the Ronda body remained relatively cool (dashed grey trajectory in Fig. 15) while the deeper part of the body became heated and affected by melting and melt-rock interaction processes (solid grey trajectory from R4 to R5).

Van der Wal & Bodinier (1996) have shown that the recrystalization front results from extensive melt-rock interaction and melt-assisted recrystallization affecting the foliated spinel tectonites and occasional mylonites at spinel-facies conditions. Van der Wal & Vissers (1993) obtained rather low temperatures (770-880 °C) in the spinel tectonites close to the front, but renewed thermobarometry by Lenoir et al. (2001) of similar samples has yielded much higher temperatures of around 1050-1100 °C. In addition, Lenoir et al. (2001) concluded that the development of the recrystallization front occurred at temperatures in the range 1180-1225 °C and pressures near those of the ariegite-seiland (a-s) subfacies boundary (field R5L in Fig. 15), i.e. at higher temperatures but also higher pressures than inferred by van der Wal & Vissers (1993). Lenoir et al. (2001) consequently interpreted the Ronda recrystallization front as a lithospheric melting front.

The melting and melt-rock interaction processes were followed by peridotite emplacement into the crust along high-temperature plagioclase-facies shear zones developed in the deeper parts of the spinel-facies granular peridotites. Most plagioclase tectonites show north directed shear (i.e., north directed displacement of the hanging wall), but a later set of narrower plagioclase tectonite shear

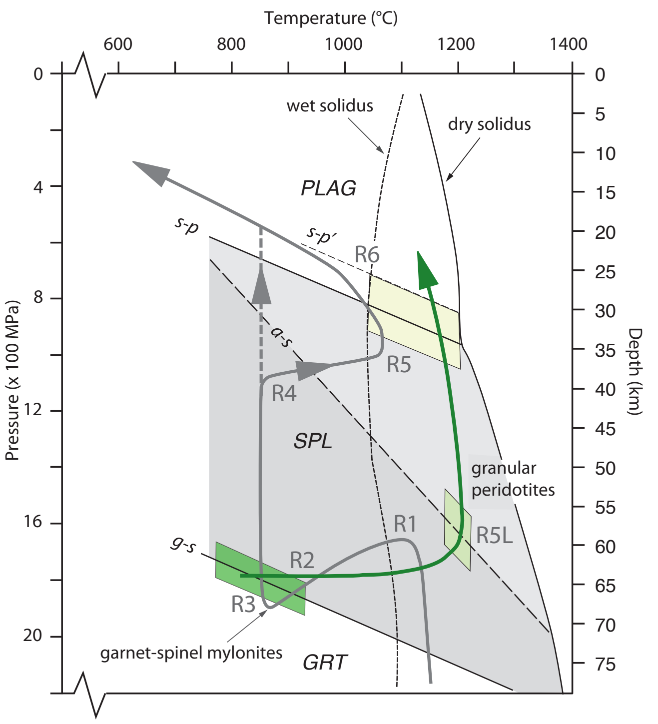

Figure 15. PT diagram showing inferred uplift trajectory of the Ronda peridotite. Peridotite solidus, garnet-spinel (g-s) reaction curve and ariegite-seiland subfacies boundary (a-s) after van der Wal & Vissers (1993), spinel-plagioclase peridotite boundary for fertile lherzolite (s-p) and depleted lherzolite (s-p') according to Borghini et al. (2009). GRT, SPL and PLAG denote garnet, spinel and plagioclase peridotite facies, respectively. Light grey PT path according to Van der Wal & Vissers (1993), with R1 early spinel-facies equilibration, R2 development of spinel tectonites, R3 garnet-spinel mylonites, R4 conditions in spinel tectonites close to recrystallization front, R5 granular spinel peridotites, R6 emplacement-related plagioclase-facies shear zones. Dashed decompression path from R4 indicates relatively cool exhumation of the Ronda mylonitic margin as opposed to coeval heating in the granular peridotites towards R5. Dark green PT path inferred by Platt et al. (2003b) for rocks at the Ronda recrystallization front. Shaded field near R2 and R3 represents equilibrium conditions during development of spinel tectonites and garnet-spinel mylonites, shaded field labelled R5L shows conditions for the granular peridotites as estimated by Lenoir et al. (2001), shaded field at s-p boundary between wet and dry solidus represents plausible conditions for the plagioclase-facies shear zones. Depth scale assumes average crustal density of 2800 kg/m³. For further explanation see text.

zones show south and west directed displacements (Fig. 13). The syntectonic plagioclase assemblages indicate that the shear zones developed during exhumation of the peridotite toward crustal levels (R6 in Fig. 15). Cooling ages of hornblende bearing leucocratic dikes (~22 Ma; Priem et al., 1979) transecting the plagioclase tectonites suggest that this stage of exhumation of the peridotite proceeded till the lower Miocene.

Van der Wal and Vissers (1993) interpreted the observed cooling during the development of the spinel tectonites and garnet-spinel mylonites and the plausible increase in pressure toward the garnet peridotite field in terms of (Paleogene) subduction-related thickening. With the aim to reconcile the structural and geothermometric data from the peridotites with those seen in the crustal envelope, Platt et al. (2003b) reconsidered the significance of the spinel tectonites and garnet-spinel mylonites. They suggested that, instead of subduction-related lithosphere thickening, the low-temperatures obtained from the garnet-spinel mylonites may equally be consistent with the onset of lithospheric extension, and that the mylonites may represent an extensional ductile shear zone deforming at temperatures that remained relatively low as a result of continuous cooling against a hanging wall of previously thickened crust. Recent structural work (Precigout et al., 2007) in addition suggests that the spinel tectonites and garnet-spinel mylonites may form one heterogeneous shear zone system, such that the spinel tectonites may equally represent extensional deformation in the upper mantle. On the basis of this alternative interpretation, and using the van der Wal & Vissers (1993) estimates for the garnet-spinel mylonites in combination with the Lenoir et al. (2001) thermobarometric results for the granular peridotites, Platt et al. (2003b) propose a P-T path for the Ronda peridotites involving substantial heating (~400 °C, from 800 to 1200 °C) followed by rapid (almost adiabatic) exhumation (Fig. 15, green trajectory).

The peridotites are underlain by high-grade gneisses (Lundeen, 1978) and migmatites beneath a faulted contact marked by extensive brecciation and gouge development, indicating final emplacement under brittle, upper crustal conditions (Fig. 13, cross section). Likewise, a 10 m scale zone of fault gouge separates the garnet-spinel mylonites of the northwestern margin from an overlying sequence of high-grade gneisses which, away from the peridotites, pass over short distances into lower-grade andalusite-facies pelites previously described as the Ronda contact aureole (Loomis, 1972; Westerhof, 1977; Torres Roldán, 1981).

5. Discussion

Any explanation for the Neogene tectonic evolution of the Betic Cordillera and Alboran Domain must account for the following observations.

1. The predominantly extensional deformation of the region during the Neogene took place in a context of overall convergence between Africa and Eurasia. A variety of plate motion paths have been proposed, and all of them show slow N to NW directed convergence in the Alboran region during the Neogene. The analysis by Vissers and Meijer (submitted) for example shows about 200 km of northward motion of Africa relative to Iberia between 83 and 19 Ma, followed by 50 km of WNW directed motion up to the present day (Fig. 16). It is clear therefore that extension in the region was not caused by motions of the bounding plates but was driven by some process acting within the Alboran system itself.

2. Extension and subsidence in the Alboran Domain (i.e., including the internal zone of the Betic Cordillera) occurred coevally with thrusting and shortening in the external zones of the Betic and Rif mountains. Extension in the Alboran Sea must have started at some time prior to the Burdigalian and continued at a reduced rate through the middle and possibly into the late Miocene (Watts et al., 1993). Onshore, the extensional detachments forming the BMZ and MAC were active in early Miocene time, and extension in the Neogene basins continued through the middle Miocene into the Tortonian. Meanwhile, thrusting of the flysch domains along the internal-external zone boundary around the Betic-Rif arc began in Burdigalian to Langhian time (~17-15 Ma; Wildi, 1983; Hermes, 1978; Martín-Algarra, 1987; Olivier, 1984; Lonergan et al., 1994). Kinematic indicators along this boundary suggest NW convergence in the eastern Betics (Lonergan et al., 1994) and WNW in the western Betics (Platt et al., 2003a), and they swing to the S in the Rif (Platzman et al., 1993; Platt et al., 2003a). NW directed convergence continued

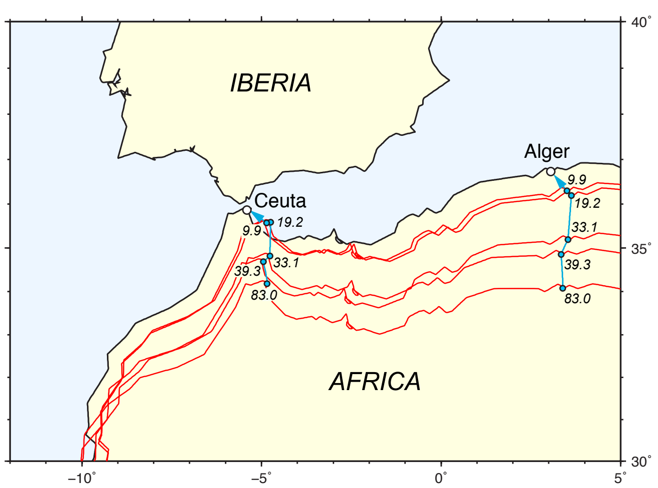

Figure 16. Motion of Africa relative to Iberia since the late Cretaceous. Total reconstruction poles for Africa and Iberia with respect to Europe after Vissers & Meijer (submitted). Africa coastline shown for reference for 83.0 Ma (anomaly 34, Late Cretaceous), 39.3 Ma (anomaly 18, Middle Eocene), 33.1 Ma (anomaly 13, Oligocene), 19.2 Ma (anomaly 6, early Miocene), and 9.9 Ma (anomaly 5, late Miocene).

in the Subbetic during the middle Miocene (Banks and Warburton, 1991), and WNW convergence in the Prebetic possibly continued into the Messinian (Frizon de Lamotte et al., 1991; Meijninger and Vissers, 2007). Similarly, W to WSW directed thrusting continued in the external Rif through into the late Miocene (Frizon de Lamotte et al., 1991; Platt et al., 2003a). During this process, extended rocks of the Alboran Domain were thrust onto the Iberian margin along the internal-external zone boundary (Fig. 1, 12): these rocks have effectively been translated from the extending into the compressional domain.

3. Extension in the Alboran Domain was largely confined to the site of former crustal thickening, suggesting a causal relationship. This is clearly visible from the distribution of the extensional basins in relation to the internal zones of the Betic-Rif arc (Figs 5 and 8). A few late Neogene basins do occur in the external zones: some of these may be deformed remnants of old flexural basins (Van der Beek and Cloetingh, 1992), and a few are genuinely extensional (e.g., the Ronda basin) and reflect minor amounts of post-thrusting extension in the external zones.

4. Extension in the Alboran Domain was accompanied by the exhumation of substantial bodies of mantle peridotite at high temperature, by a regional metamorphic event that locally reached very high temperatures, and by scattered mixed mode volcanism.

5.1. Tectonic Hypotheses for the Betic/Alboran Region

Several hypotheses have been proposed that call on changes in the underlying mantle structure to explain the pattern of Neogene deformation in the Alboran region. Royden (1993) and Lonergan and White (1997) have suggested back arc extension driven by slab rollback to explain the pattern of Neogene deformation in the Alboran region. Seismic tomography by Gutscher et al. (2002) and Spakman and Wortel (2004) shows the presence of a southeast to east dipping curved slab underneath the Betics, leading these authors to suggest that extension and thinning in the Betic-Rif belt and Alboran Domain was induced by west-directed rollback and steepening of a subducting oceanic plate. Two problems surround this hypothesis. First, there is no clearcut geological evidence for the existence of oceanic lithosphere in the region at any time during the Neogene. The local occurrence of thin slivers of E-MORB basic rocks, however, may indicate that the Flysch Trough units in the western Betics were floored by oceanic crust (Durand-Delga et al., 2000; Booth-Rea et al., 2007, and references therein). Secondly, slab roll-back by itself cannot account for the both extreme and rapid heating at shallow depths documented in the metamorphic crustal envelope of the Ronda peridotites (Platt et al., 2003b) and in the metamorphic rocks at Site 976 in the Alboran Sea (Comas et al., 1999), nor for the high temperatures implied by the melt-rock interaction and melt impregnation processes leading to the development of a lithospheric melting front in the Ronda massif.

As opposed to the concept of slab roll-back driving trench retreat, three in a sense similar but differently formulated hypotheses may account for the geological observations in the Alboran system. These include detachment of a subducting slab (Blanco and Spakman, 1993) and either delamination (Channell and Mareschal, 1989; Seber et al., 1996) or convective removal (Platt and Vissers, 1989) of part or all of the lithospheric mantle beneath the Alboran region (Fig. 17). Each of these processes involve the separation and sinking of dense lithospheric mantle from beneath the zone of convergence, thereby removing a downward load. The result should be first an increase in surface elevation and potential energy of the region, which could cause it to extend, even though the bounding plate motions continue to be convergent (England and Houseman, 1989; Platt and England, 1994). Secondly, the ensuing ascent of asthenospheric mantle to shallow levels close to or at the base of the crust should lead to a thermal pulse causing magmatism and metamorphism. The three hypotheses differ mainly in the mechanical properties assumed for the lithosphere and the amount that is removed. The concept of slab detachment (Wortel and Spakman, 1992, 2000; Davies and von Blanckenburg, 1995) requires the subduction of a rigid, presumably oceanic plate that is likely to detach, in particular once collision has occurred. In view of the evidence for continental collision in the Alboran region as early as the Eocene and the lack of genuine evidence for Tertiary oceanic crust, Platt and Vissers (1989) and Vissers et al. (1995) preferred to think in terms of the deformation of continental lithosphere along a pre-existing lithospheric discontinuity. Delamination (Bird, 1979; Channell and Mareschal, 1989) assumes that the subcontinental mantle lithosphere in a region of continental collision will behave in slab-like fashion and will peel off along the Moho. The consequence will be the juxtaposition of asthenospheric mantle with the base of the crust, which is likely to cause immediate and widespread melting of the lower crust. Convective removal of lithosphere draws on a concept developed by Houseman et al. (1981) that treats the lithosphere as a conductive thermal boundary layer to the convecting mantle, with the lower part of the lithosphere behaving as a viscous fluid that is only intermittently involved in convection. Thickening results in an increase of the Rayleigh number of the fluid lower lithosphere and hence of the rate at which it is involved in convective motion, resulting in its removal and the reestablishment of a thermally stable boundary layer thickness.

5.2. An Integrated Hypothesis for the Betic/Alboran Region

The Alpine history of the Alboran region began with the left-lateral oblique rifting event that accompanied the separation of Africa and Eurasia during the Jurassic (e.g. Dewey et al., 1989), creating passive continental margins on the African and Iberian sides of the opening Neotethys basin. This basin was probably floored in part by oceanic crust, but contained one or more partially submerged continental platforms, now represented by the various Paleozoic basement complexes in the internal Betic and Rif mountains. The early uplift of the diamond-bearing peridotites of the Ronda and Beni Bousera massifs from depths of around 150 km into the spinel peridotite stability field (i.e., mid-lithospheric levels) may well have occurred during this phase of extension, as suggested by the recovery of zircon ages around 160 Ma (Sánchez-Rodríguez, 1998).

Plate motion analyses suggests that significant convergence between Africa and Iberia began at about 51 Ma (Dewey et al., 1989) or as early as 83 Ma (Vissers and Meijer, submitted). Stratigraphic evidence (Lonergan, 1993) and radiometric dates on the high P/T metamorphism (Monié et al., 1992) suggest that convergence began or was already underway by late Eocene time, although earlier dates have been suggested (e.g., De Jong, 1991). This resulted in the

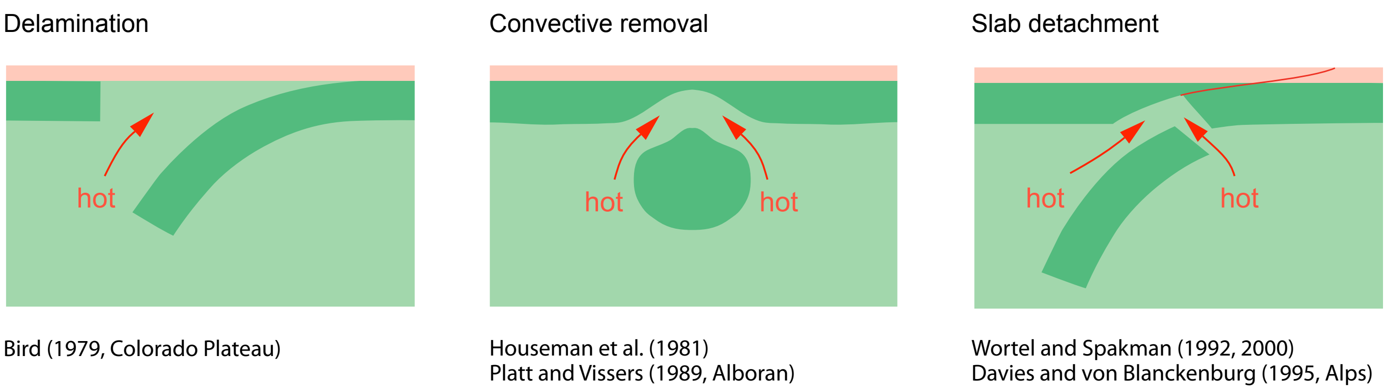

Figure 17. Cartoon illustrating three ways in which the lithosphere-scale structure may change at geologically short time scales. Note that delamination, convective removal and slab detachment each imply that cold and gravitationally unstable heavy lithosphere sinks into the underlying mantle, and that hot asthenospheric mantle replaces this cold lithosphere at shallow levels. For further explanation see text.

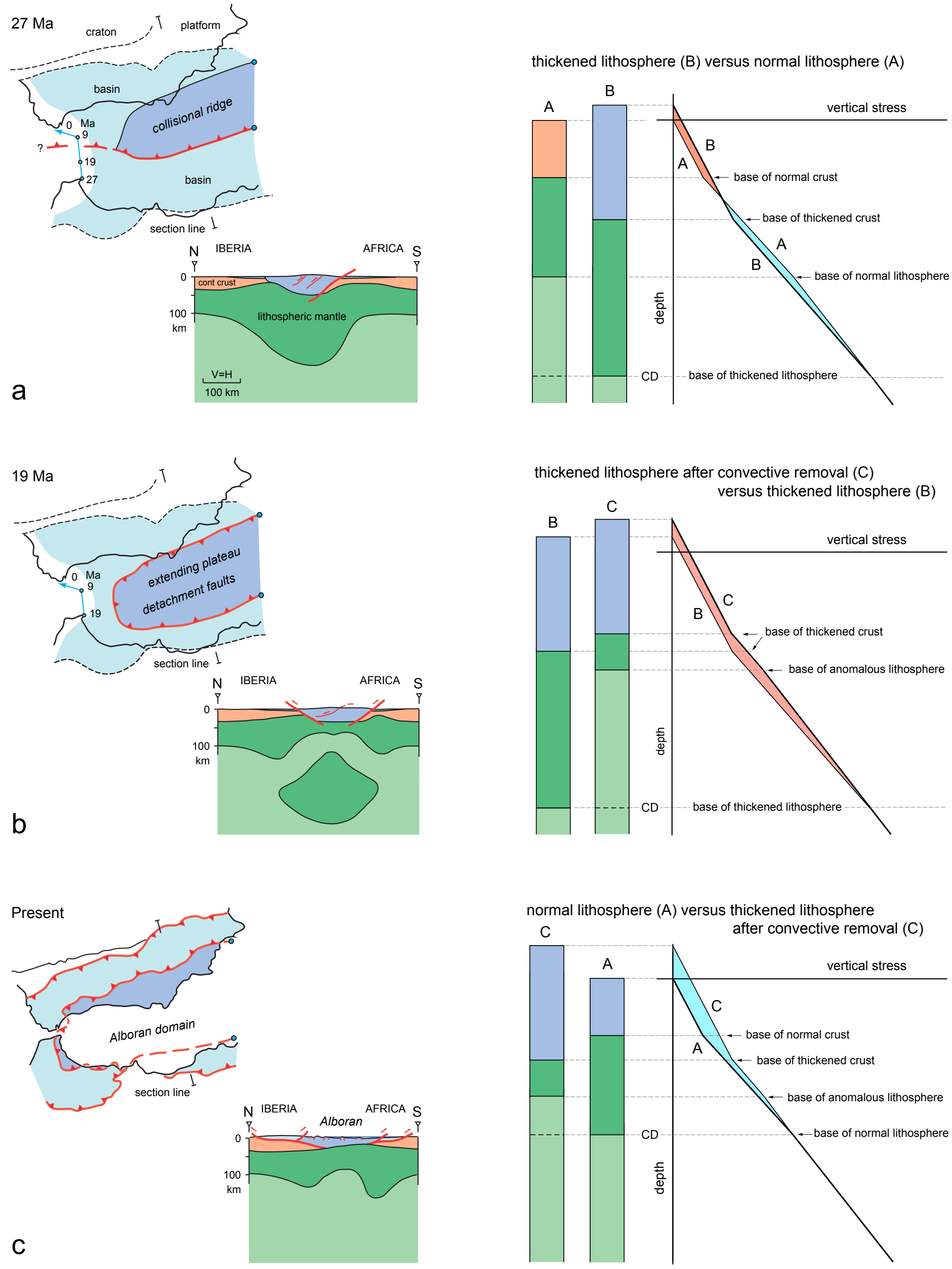

Figure 18. Tectonic evolution of the Alboran region, partly modified from Platt & Vissers (1989). Maps and lithosphere-scale cross sections illustrate three stages. Coastlines of SW Spain and NW Africa shown for reference. Plate motion vectors for Africa relative to Europe since 27 Ma after Dewey el al. (1989). Diagrams to the right illustrate pertinent changes in gravitational potential energy of the Alboran lithosphere; CD denotes compensation depth for isostatic equilibrium. (a) At Late Oligocene times (27 Ma), a collisional ridge with a thickened lithospheric root had formed in response to Paleogene convergence. Basins were probably underlain by thin continental crust. Diagram to the right illustrates how vertical stress increases with depth for two lithosphere columns, one for normal and one for thickened lithosphere. Lithosphere thickening leads to a net decrease in potential energy as explained in text. (b) Paleogeography and lithosphere section envisaged for the Burdigalian (Early Miocene, ~19 Ma). Convective removal of lithospheric root has caused uplift, increase in potential energy of collisional ridge, and extension. Extension is accomodated by crustal shortening around margins. Colums and graph to the right illustrate marked increase in gravitational potential energy due to loss of overthickend lithosperic root. (c) Present-day geometry of the Alboran system. Extending Alboran domain has emplaced onto surrounding continental margins, its center subsides as lithosphere thickens by cooling and continued slow convergence. Diagram to the right shows that while extension driven by excess potential energy proceeds, cooling leads to growth of the mantle lithosphere, with the tendency to return to a normal lithosphere with associated lower potential energy. The system looses potential energy, hence the area between the two graphs is shown in blue.

stacking of tectonic units in the Alboran region to create a region of thickened continental crust and lithosphere. Early thrusting directions have been documented towards the NW in the Nevado-Filabride Complex (Bakker et al., 1989), the NE (Tubía et al., 1992) or the S (Platt et al., 1993) in the Alpujarride Complex, and the NW in the Malaguide Complex (Lonergan, 1993). In any case, early kinematic indicators have most probably been affected by large rotations about vertical axes (Allerton et al., 1993). The paleogeography and gross lithosphere structure of the Alboran region are illustrated in Fig. 18a. The stress-depth graphs to the right in Fig. 18a serve to illustrate the physics of lithosphere thickening. As outlined by Molnar and Lyon-Caen (1988), the gravitional potential energy of a lithosphere column equals the vertically integrated vertical stress, i.e., the area below the vertical stress - depth curve for that column. It follows that for a normal and a thickened lithosphere the difference of these areas below the two graphs represents the difference in potential energy of the two columns. In the case of lithosphere thickening, the effect of an increase in potential energy of the crust (indicated in red in Fig. 18) is commonly outweight by a larger decrease (indicated in blue) due to thickening of the underlying lithospheric mantle (Fig. 18a), such that the net potential energy of the thickened lithosphere will be less that that of a normal lithosphere (see also Platt and England, 1994). Fig. 19 illustrates the evolution of the Alboran Domain and surrounding chains for two different hypotheses with regard to the underlying upper mantle processes, i.e., homogeneous thickening followed by convective removal of the thermal boundary layer (Fig. 19a), and subduction of a relatively small oceanic domain followed by slab detachment (Fig. 19b).

Some form of detachment or convective removal of lithospheric mantle occurred in latest Oligocene or earliest Miocene time (Figs 18b, 19a and b, panels 2). Removal of part of the lithospheric mantle beneath the collisional ridge had several direct and indirect consequences.

(1) There should have been an immediate increase in surface elevation (England & Houseman, 1989), possibly reflected in the increased clastic sedimentation during the Oligocene to early Miocene in the flysch basins on the north African margin and in the Gibraltar area (Wildi, 1983; Olivier, 1984).

(2) The increase in potential energy of the system (Fig. 18b) resulting from the increase in surface elevation caused the Alboran domain to extend, exhuming high-pressure crustal and mantle rocks along large-scale extensional detachments, and forming extensional basins at the surface. Panels 3 in Fig. 19a and b illustrate this extensional collapse for the two scenarios. The extension certainly included a N-S component, as indicated by kinematic data from the extensional basins and the predominance of E-W trending horst and

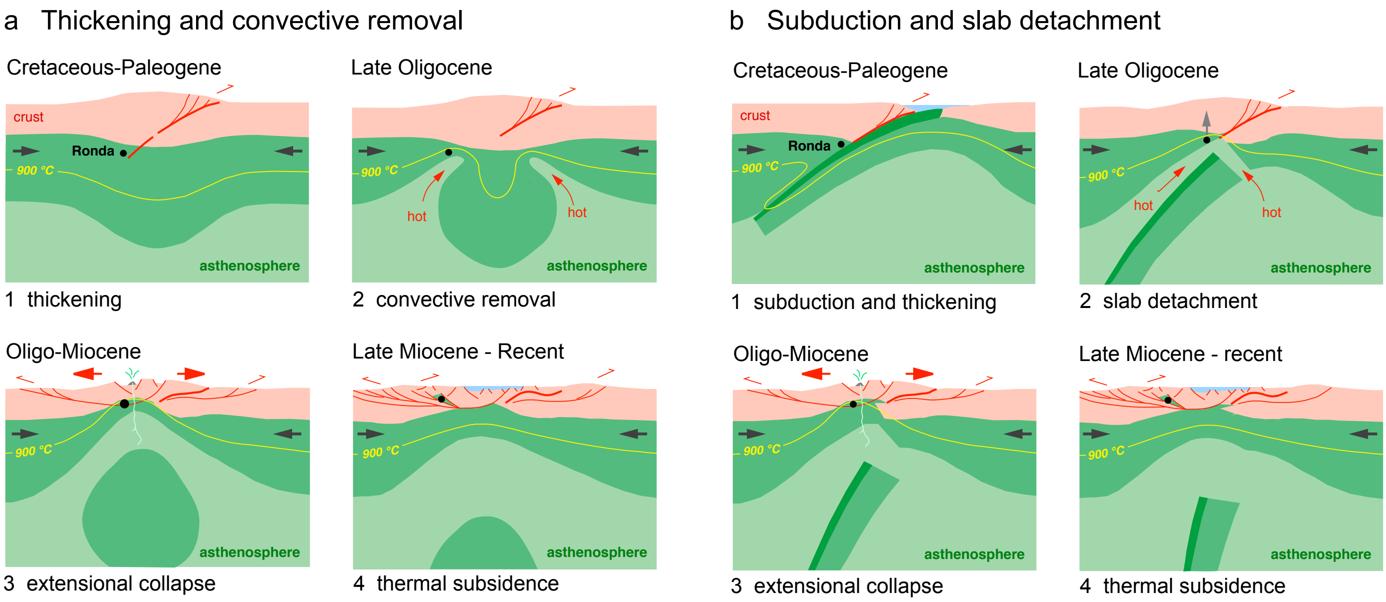

Figure 19. Diagrammatic cross sections illustrating two geodynamic scenarios for the development of a collision belt in the Alboran region subsequently affected by extensional collapse in a continuously convergent setting. (a) Scenario assuming homogeneous thickening and convective removal of overthickened lithospheric mantle: (1) Cretaceous–Paleogene lithosphere thickening is followed by (2) convective removal during the late Oligocene leading to increase in surface elevation and gravitational potential energy, driving (3) extensional collapse of the thickened crust, development of horst and graben structures in the Alboran Domain, and extensional exhumation of mantle peridotites (black dot). Upwelling of asthenosphere results in decompression melting, while the consequent sudden heating at the base of the thinning crust causes HT metamorphism, partial melting and mixed mode volcanism. (4) Cooling since the late Miocene leads to thermal subsidence while ongoing Africa Iberia convergence leads to final emplacement of thrust units. (b) Alternative scenario assuming (1) Createaceous-Paleogene subduction of a small oceanic domain, followed by (2) slab detachment and inherent ascent of asthenosheric mantle, leading to (3) extensional collapse followed by (4) cooling and thermal subsidence. Note that both scenarios can conveniently explain the exposed surface geology.

graben structures, but there was also a strong E-W component, as shown by the kinematic data from the mylonites along the crustal detachments. Sense and directions of shear changed in both space and time, presumably reflecting the complex and changing geometry of the extending crust. The spinel-tectonites and garnet-spinel mylonites in the Ronda body reflect the onset of ductile extension at upper mantle levels.

(3) An abrupt step in the geotherm was created at the base of the thinned lithosphere, which triggered partial melting in the latter (Pearce et al., 1990; Platt & England, 1994), leading to early Miocene mafic magmatism in the Malaga region, for example (Torres-Roldán et al., 1986).

(4) The combination of convective removal of the lithospheric root and extension allowed a transient thermal pulse to affect the upper mantle and crust during and after decompression (Platt & England, 1994). In the Ronda peridotite this is reflected by decompression from the spinel-garnet transition zone into the Seiland subfacies stability field, accompanied by extensive recrystallization and melt-rock interaction to develop the granular spinel peridotites (Fig. 15). These were exhumed along low-pressure, plagioclase-bearing tectonite shear zones (i.e., major upper-mantle extensional faults accommodating tectonic denudation), which evolved into cataclastic fault zones related to the final emplacement of the peridotite bodies amidst tectonic slices of crustal origin (Tubía & Cuevas, 1986, Lundeen, 1978). In deep crustal rocks the thermal event produced the characteristic late intermediate to LP/HT metamorphism that accompanied exhumation; in shallower crustal rocks it caused local postkinematic LP/HT metamorphism.

(5) Continued extension and decompression of asthenosperic mantle resulted in partial melting; together with partial melting in the crust associated with the thermal event this resulted in scattered mafic, intermediate and silicic volcanism during the Neogene (Fig. 19a and b, panels 3).

(6) Extension of the Alboran domain caused its margins to move outwards. Because of its elongate shape this involved a component of westward motion, particularly at the western end of the region. Combined with continuing Africa/Iberia convergence this caused a pattern of outwardly directed thrusting in the external zones of the Betic and Rif chains (Platzman, 1992; Platt et al., 2003a). The margins of the Alboran domain were thrust over the surrounding basins onto the continental margins of Iberia and Africa. As a result, some regions first involved in extensional exhumation (including the Ronda and other peridotites; Fig. 12, Fig. 19a and b, panel 4) were transported into the compressional region on the margins of the Alboran domain and became involved in thrusting.

(7) Both convective removal of lithosphere and the subsequent extension elevated the thermal gradient, hence the surface heat flow of the region, so that overall cooling and subsidence followed. This resulted in much of the Alboran domain subsiding below sea-level during the Miocene and the accumulation of marine sediments. Towards the end of the Miocene much of the Betic/Alboran region was affected by a combination of strike-slip faulting, local folding and thrusting, inverting the earlier sedimentary basins. These effects may reflect the progressive cooling and decrease in potential energy of the region (Figs 18c; Fig. 19a and b, panels 4), such that the convergent motion between Africa and Europe once more became the dominant tectonic process. At the same time the presently emergent part of the internal Betic Cordillera underwent a substantial increase in surface elevation: sedimentary basins that were marine in the Tortonian are now exposed at elevations between 200 and 1000 m. The elevated region approximately corresponds to the area that has been emplaced onto the Iberian continental basement that originally underlay the External Zones (Fig. 12). This crustal thickening on the margins of the extending Alboran domain may explain the current elevation.

6. Conclusive remark

The geology of the Alboran region and its underlying geodynamics are in all likelihood not unique to the region, and many similarities can be identified with processes inferred for the Tibetan Plateau in the Alpine orogenic system, but also for the Variscan (Ménard & Molnar, 1988) and Pan-African orogenic belts (Black & Liegeois, 1993) where extensional processes can be shown to also have occurred in an essentially convergent setting. The Alboran region, however, is rather unique in that aside geological studies of the metamorphic crustal rocks and overlying basin sediments, large exposures of mantle rocks allow to directly investigate coeval processes in the pertinent upper mantle. The overall westward motion during Neogene extension of the Alboran domain indicates that the underlying processes in the upper mantle may, in fact, have occurred east and northeast of the present-day Alboran region. These earlier stages are, therefore, necessarily related to the evolution of the western Mediterranean region as a whole, and further research is needed to obtain a firm grip on that evolution, and on relationships with the adjacent parts of the Alpine belt such as the Alps. The resulting geodynamic scenarios and hypotheses should, in any case, account for the first-order characteristics of the Betic/Alboran system. Quoted differences between current hypotheses may in fact be less fundamental than they seem at first inspection. Fig. 19 illustrates how different hypotheses for the upper mantle processes in the Alboran region may be consistent with the same surface geology. As outlined by Turner et al. (1999): "Delamination, convective removal of lithosphere and slab detachment all predict significant uplift before the onset of extension, together with decompressional magmatism and transient conductive heating of the lithosphere (Platt & England, 1994). The surface geometry and evolution of the effects should differ according to the model, but the differences may not be sufficiently clearcut to be diagnostic".

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