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## ► To cite this version:

Gianluca Frasca, Frédéric Gueydan, Patrick Monie, Jean-Pierre Brun. Deformation mechanisms in a continental rift up to mantle exhumation. Field evidence from the western Betics, Spain. *Marine and Petroleum Geology*, 2016, 76, pp.310-328. 10.1016/j.marpetgeo.2016.04.020 . insu-01310895

**HAL Id: insu-01310895**

**<https://insu.hal.science/insu-01310895v1>**

Submitted on 3 May 2016

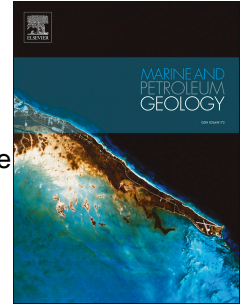
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# Accepted Manuscript

Deformation mechanisms in a continental rift up to mantle exhumation. Field evidence from the western Betics, Spain

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PII: S0264-8172(16)30115-5

DOI: [10.1016/j.marpetgeo.2016.04.020](https://doi.org/10.1016/j.marpetgeo.2016.04.020)

Reference: JMPG 2538

To appear in: *Marine and Petroleum Geology*

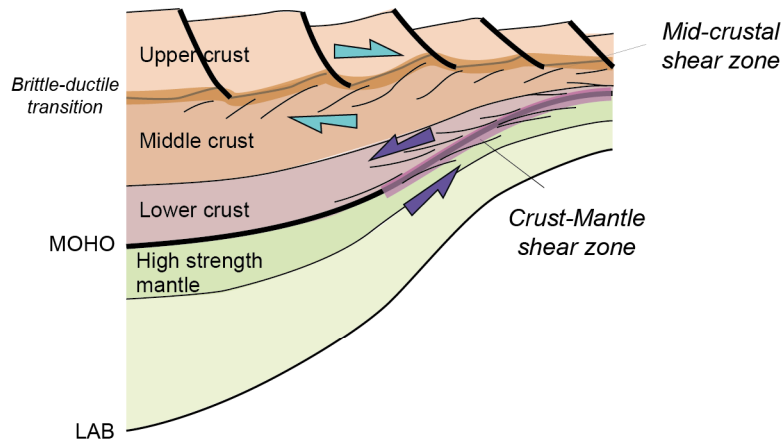
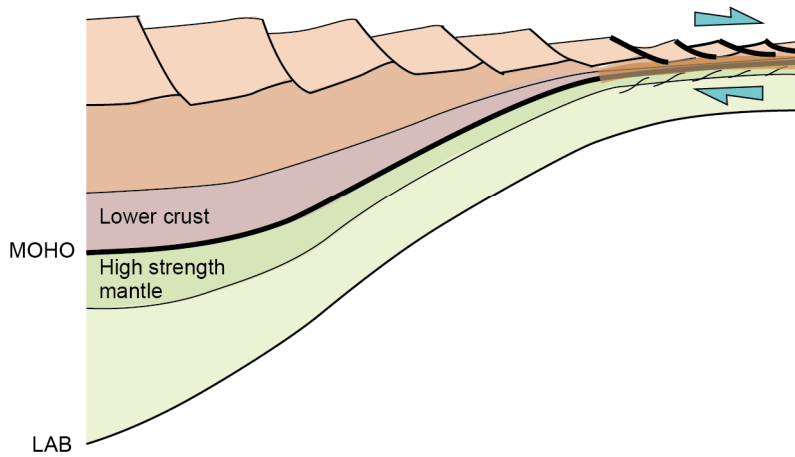
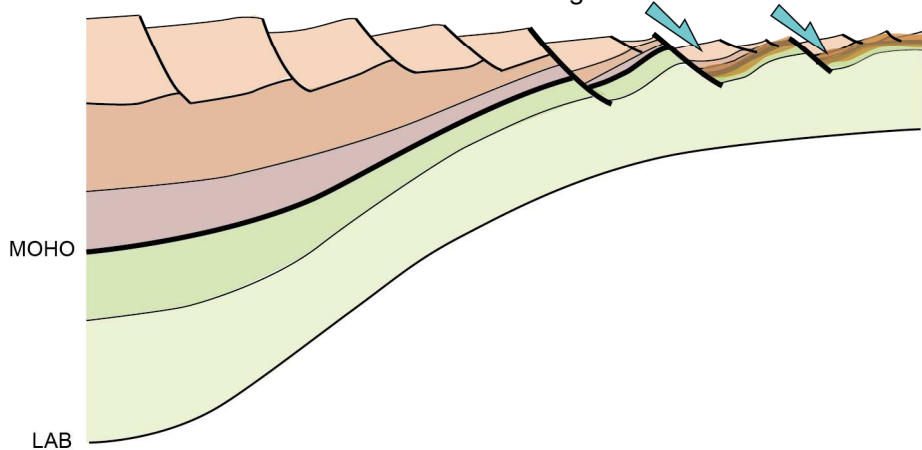
Received Date: 5 June 2015

Revised Date: 13 April 2016

Accepted Date: 20 April 2016

Please cite this article as: Frasca, G., Gueydan, F., Brun, J.-P., Monié, P., Deformation mechanisms in a continental rift up to mantle exhumation. Field evidence from the western Betics, Spain, *Marine and Petroleum Geology* (2016), doi: 10.1016/j.marpetgeo.2016.04.020.

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**33-25 Ma; Early stages of lithosphere necking***Crust-mantle decoupling, crust heating by exhuming mantle***25-22 Ma; Advanced stages of lithosphere necking***Crust-mantle coupling, localisation (hyper-stretching), onset of cooling***22-20 Ma; Late stages of lithosphere necking***Mantle faulting & block tilting, cooling*

1 **Deformation mechanisms in a continental rift up to mantle exhumation.**

2 **Field evidence from the western Betics, Spain**

3

4

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6

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12

13

14

15

16 **Abstract**

17

18 The identification of the structures and deformation patterns in magma-poor continental rifted

19 margins is essential to characterize the processes of continental lithosphere necking. Brittle

20 faults, often termed mantle detachments, are believed to play an essential role in the rifting

21 processes that lead to mantle exhumation. However, ductile shear zones in the deep crust and

22 mantle are rarely identified and their mechanical role remains to be established. The western

23 Betics (Southern Spain) provides an exceptional exposure of a strongly thinned continental

24 lithosphere, formed in a supra-subduction setting during Oligocene-Lower Miocene. A full

25 section of the entire crust and the upper part of the mantle is investigated. Variations in crustal

26 thickness are used to quantify crustal stretching that may reach values larger than 2000%  
27 where the ductile crust almost disappear, defining a stage of hyper-stretching. Opposite senses  
28 of shear top-to-W and top-to-E are observed in two extensional shear zones located close to  
29 the crust-mantle boundary and along the brittle-ductile transition in the crust, respectively. At  
30 locations where the ductile crust almost disappears, concordant top-to-E-NE senses of shear  
31 are observed in both upper crust and serpentinized mantle. Late high-angle normal faults with  
32 ages of ca. 21 Ma or older ( $^{40}\text{Ar}/^{39}\text{Ar}$  on white mica) crosscut the previously hyper-stretched  
33 domain, involving both crust and mantle in tilted blocks. The western Betics exemplifies,  
34 probably better than any previous field example, the changes in deformation processes that  
35 accommodate the progressive necking of a continental lithosphere. Three successive steps can  
36 be identified: i/ a mid-crustal shear zone and a crust-mantle shear zone, acting synchronously  
37 but with opposite senses of shear, accommodate ductile crust thinning and ascent of  
38 subcontinental mantle; ii/ hyper-stretching localizes in the neck, leading to an almost  
39 disappearance of the ductile crust and bringing the upper crust in contact with the  
40 subcontinental mantle, each of them with their already acquired opposite senses of shear; and  
41 iii/ high-angle normal faulting, cutting through the Moho, with related block tilting, ends the  
42 full exhumation of the mantle in the zone of localized stretching. The presence of a high  
43 strength sub-Moho mantle is responsible for the change in sense of shear with depth. Whereas  
44 mantle exhumation in the western Betics occurred in a backarc setting, this deformation  
45 pattern controlled by a high-strength layer at the top of the lithosphere mantle makes it  
46 directly comparable to most passive margins whose formation lead to mantle exhumation.  
47 This unique field analogue has therefore a strong potential for the seismic interpretation of the  
48 so-called “hyper-extended margins”.

49

50

51 **1/ Introduction**

52

53 During the last decades, the exhumation of subcontinental lithospheric mantle in rifted  
54 margins became a commonly recognized tectonic feature. In the Iberia margin, the case-study  
55 for magma-poor continental margin (Boillot et al., 1980; Boillot et al., 1987; Beslier et al.,  
56 1990; Whitmarsh and Miles, 1995; Brun and Beslier, 1996), the crustal thickness decreases  
57 abruptly in a short horizontal distance (75 km; Whitmarsh et al., 2001; Péron-Pinvidic and  
58 Manatschal, 2009) leading to wide mantle exposures directly to the sea-floor. The sudden  
59 extreme thinning of the continental crust has been well imaged through reflection seismics  
60 (Péron-Pinvidic et al., 2007; Ranero and Perez-Gussinyé, 2010) and ascribed to polyphased  
61 and complex activity of ductile and brittle structures (Manatschal et al., 2001; Péron-Pinvidic  
62 et al., 2007).

63

64 The bulk process responsible for lithosphere thinning up to continental breakup is a  
65 “necking” instability that leads to a sharp decrease in crustal thickness and to strong  
66 decoupling between upper crust and mantle allowing the mantle to exhume in the rift center  
67 (Brun and Beslier, 1996). The final stage of lithosphere thinning, so-called “hyperextension”,  
68 is then characterized by coupled continental brittle crust and mantle. Faults penetrate into the  
69 mantle and “extensional allochthons” of upper crust lie in direct contact with the mantle  
70 (Beslier et al., 1993; Manatschal et al., 2001). This final stage of rifting is now well  
71 documented worldwide, e.g. in the South Atlantic (Contrucci et al., 2004; Moulin et al., 2005),  
72 in the Red Sea (Cochran and Karner, 2007) and in Norway (Osmundsen and Ebbing, 2008),  
73 and fossil examples can be observed directly in the field in the obducted margins in the Alps  
74 (Manatschal, 2004; Beltrando et al., 2012; Mohn et al., 2010, and references therein) and in  
75 the Pyrenees (Clerc et al., 2012; Clerc and Lagabriele, 2014). Normal faults and detachment

76 faults shape this late stage of rift (i.e. “hyperextension”) and may obliterate the earlier  
77 structures that are responsible for a large part of the lithosphere thinning. It is therefore  
78 difficult to obtain a detailed resolution of the first stages of “necking” from seismic data (for a  
79 detailed discussion see Reston, 2007). In the other hand, analogue and numerical models  
80 reveal that large ductile shear accommodates crust-mantle decoupling and hence controls the  
81 process of sub-continental mantle exhumation during lithosphere necking (Brun and Beslier,  
82 1996; Gueydan et al., 2008). More generally, many aspects of the relationships between upper  
83 crustal faulting and crust-mantle deformation during necking remain poorly understood. As  
84 an example, fault displacements observed in seismic lines always remain rather small  
85 compared to the overall crustal thinning (Reston, 2007). To reconcile such discrepancies  
86 various solutions have already been considered: i) large-scale crustal detachments (Lister et  
87 al., 1986; Manatschal et al., 2001; Froitzheim et al., 2006; Mohn et al., 2012), ii) upper crustal  
88 brittle faulting accommodated by lower crustal flow (Brun and Beslier, 1996; Ranero and  
89 Perez-Gussinyé, 2010; Gueydan and Précigout, 2014) or iii) depth-dependent mechanisms of  
90 thinning at different levels in the crust (Davis and Kuszniir, 2004; Huisman and Beaumont,  
91 2011).

92  
93 To unravel the different stages of necking, a key issue is to characterize deformation  
94 of the rifted lithosphere at different levels: subcontinental mantle, deep crust and upper crust.  
95 However, sections reaching the deepest levels of a rifted lithosphere are rarely exposed in the  
96 field. The large bodies of subcontinental mantle exposed in the westernmost Alboran region,  
97 in the western Mediterranean, provide a unique opportunity to study the brittle and ductile  
98 deformations that developed during lithosphere necking. In this region, a complete section of  
99 the stretched continental crust is associated to the exhumation of the Ronda Peridotites that is,  
100 with its three main massifs, the world largest outcrop of subcontinental mantle (Balanyá et al.,

101 1997; Argles et al., 1999). Major extensional shear zones located at the crust-mantle interface  
102 are coeval with the subcontinental mantle exhumation (Afiri et al., 2011; Précigout et al.,  
103 2013; Frets et al., 2014; Gueydan et al., 2015). The western Betics is thus a prime location to  
104 observe and measure the effects of stretching at different levels of the continental lithosphere  
105 (Tubía and Cuevas et al., 1986; Balanyá et al., 1997; Argles et al., 1999; Précigout et al.,  
106 2013).

107

108 In the present paper, we use the outstanding field example of the western Betics to  
109 study the patterns of deformation and kinematics that characterize the progressive necking of  
110 a continental lithosphere. After a review of the geological and tectonic setting of the Ronda  
111 peridotites, we describe i) the structural and geochronological data collected during this study,  
112 ii) the major ductile shear zones located in the middle crust and at the crust-mantle transition  
113 and iii) the regional gradient of thinning undergone by the crustal units. The discussion is  
114 dedicated to the variations of shear sense with depth and to the processes of lithosphere  
115 necking in the western Betics.

116

117

## 118 **2/ Geological setting**

119

120 The *Betic-Rif* belt is divided into a metamorphic Internal Zone (Alboran Domain) and  
121 a non-metamorphosed External Zone (inset of Fig. 1). The Internal-External Zone Boundary  
122 (IEZB) separates the Alboran Domain from an external thrust-and-fold belt (e.g. Crespo-  
123 Blanc and Frizon de Lamotte, 2006; Chalouan et al., 2008). Tertiary foreland basins wrap the  
124 belt to the North, West and South (Flinch, 1993; Fernández et al., 1998) (shaded in inset of  
125 Fig. 1). The External Zone represents the subducting Iberian and Maghrebian margins (pale



126 grey in inset of Fig. 1) and the Alboran Domain is a portion of the upper plate of the western  
127 Mediterranean subducting system (dark grey in inset of Fig. 1) (e.g. Garrido et al., 2011; Platt  
128 et al., 2013). A large part of the Alboran Domain stands below sea level in the Alboran Sea  
129 basin, whose development is mostly coeval with Miocene shortening in the arcuate External  
130 Zone (García-Dueñas et al., 1992; Watts et al., 1993; Balanyá et al., 1997; Comas et al., 1999;  
131 Platt et al., 2013).

132  
133 The metamorphic tectonic units of the Alboran Domain contain in the westernmost  
134 sector the Ronda Peridotites. The *exhumation* mechanism and timing of the *Ronda Peridotites*  
135 have been and are still *matter of controversy*. Among the proposed mechanisms of mantle  
136 rock exhumation are: i) mantle core complex (Doblas and Oyarzun, 1989), ii) crosscutting  
137 detachment faults during the extensional collapse of the Betic-Rif chain (Van der Wal and  
138 Vissers, 1993; Platt et al., 2003a), iii) transpressional extrusion of a mantle wedge (Tubía et  
139 al., 1997; Mazzoli and Martín-Algarra, 2011; Tubía et al., 2013), and iv) inversion during slab  
140 rollback of a thinned back-arc lithosphere (Garrido et al., 2011; Hidas et al., 2013; Précigout  
141 et al., 2013). Ages of exhumation also strongly vary: i) Paleozoic (Kornprobst, 1976; Ruiz  
142 Cruz and Sanz de Galdeano, 2014), ii) Mesozoic (Van Hinsbergen et al., 2014; Vissers et al.,  
143 1995) and iii) Oligo-Miocene (Hidas et al., 2013; Précigout et al., 2013).

144  
145 *2.1/ The Ronda – Beni Bousera subcontinental mantle*

146  
147 The *Ronda Peridotites* in southern Spain consist of three main massives, called  
148 Bermeja, Alpujata, Carratraca massives (Fig. 1). Several smallest mantle outcrops connect the  
149 three larger massives suggesting the original continuity of a single mantle sliver (Navarro-  
150 Vilá and Tubía, 1983; see Didon et al., 1973 for an alternative view). Furthermore, the Ronda

151 Peridotites in the western Betics were probably originally continuous with the Beni Bousera  
152 massif (Sánchez-Gómez et al., 2002) on the Rifan side of the Gibraltar arc (inset in Fig. 1)  
153 and then dismembered during the Miocene formation of the Gibraltar arc (Balanyá et al.,  
154 1997; Berndt et al., 2015; Chalouan et al., 2008; Frasca et al., 2015). The peridotites display a  
155 kilometer-scale petrological zoning with Grt/Sp-peridotites at the top, granular and  
156 porphyroclastic Sp-peridotites in the middle and Plag-bearing peridotites at the base (Obata,  
157 1980). The Ronda Peridotites sliver is separated from the Alboran continental crustal rocks by  
158 *two major tectonic contacts*: i) the Ronda Peridotites Thrust at the base and ii) the Crust-  
159 Mantle extensional shear zone at the top (respectively black and white colors in map and  
160 cross-section of Fig. 1) (Tubía et al., 1997; Mazzoli et al., 2013; Précigout et al., 2013;  
161 Johanesen and Platt, 2015).

162  
163 The *Ronda Peridotites Thrust* is a “hot” thrust (Tubía et al., 1997; Esteban et al., 2008)  
164 characterized by a metamorphic sole with: 1/ partial melting in the footwall metamorphic  
165 rocks with ages ranging between 22 Ma and 19 Ma (Esteban et al., 2011), and 2/ high-  
166 temperature minerals in the basal Mesozoic sediments deformed during Lower Miocene  
167 (high-temperature: Mazzoli et al., 2013; deformed Lower Miocene Nava Breccia: Mazzoli  
168 and Martín-Algarra, 2011). The Plagioclase tectonites (Fig. 1) usually mark the deformation  
169 immediately above the Ronda Peridotites Thrust (Hidas et al., 2013; Précigout et al., 2013).

170  
171 Conversely, the *Crust-Mantle Shear Zone* shows structural and petrological evidences  
172 of an extensional nature (Tubía and Cuevas, 1986; Balanyá et al., 1997; Argles et al., 1999;  
173 Sánchez-Gómez et al., 1999; Précigout et al., 2013). Garnet-spinel mylonites mark the Crust-  
174 Mantle Shear Zone with a top-to-SW shearing in the Bermeja massif (Précigout et al., 2013),  
175 top-to-E shearing in the Alpujata massif (Tubía and Cuevas, 1986), a variable sense of shear

176 in the Carratraca area (Tubía et al., 2004), and top-to-NW sense of shear in the Beni Bousera  
177 massif (Afiri et al., 2011; Frets et al., 2014). Simultaneously, mantle rocks have recorded a  
178 continuous decompression, from the garnet stability field to spinel-peridotite facies (Garrido  
179 et al., 2011), related to ductile strain localization at the very top of the mantle units (Précigout  
180 et al., 2007; Afiri et al., 2011; Précigout et al., 2013). Gabbros and Cr-rich pyroxenites  
181 developed during mantle exhumation (Marchesi et al., 2012; Hidas et al., 2015). Nevertheless,  
182 all mantle massives are dominated by spinel lherzolites in which mafic layers are common  
183 (Garrido and Bodinier, 1999). The granular plagioclase-peridotites at the base are separated  
184 from the overlying spinel tectonites by a “recrystallization front” that marks the asthenosphere  
185 boundary (Van der Wal and Vissers, 1996; Lenoir et al., 2001; Soustelle et al., 2009) that is  
186 not exposed in the Carratraca massif (Tubía et al., 2004).

187

## 188 *2.2/ The western Alboran crust*

189

190 The western Alboran crustal envelope of the Ronda Peridotites is mainly made of  
191 Paleozoic rocks divided in two tectonic units, Alpujarride and Malaguide, which form an  
192 Alpine nappe stack most probably related to subduction during the Eocene (e.g. Vergés and  
193 Fernández, 2012; Platt et al., 2013). This Alpine nappe stack represents the Alboran crust, on  
194 top of the Ronda peridotites, during the Oligo-Miocene evolution (Torres-Roldán, 1979;  
195 Tubía et al., 1992). The intrusion of Oligocene tholeiitic dykes in both nappes indicates that  
196 nappe stacking is pre-Oligocene (Esteban et al., 2013). The Alpujarrides rocks are sometimes  
197 divided in granulites, migmatites, gneisses and schists (e.g. Chalouan et al., 2008). Hereafter,  
198 following partly previous field observations (Balanyá et al., 1997; Argles et al., 1999), we  
199 divide the entire crust –i.e. the Alpujarrides-Malaguide pair- into three lithological groups,  
200 namely lower, middle and upper crust (Log of Fig. 1).

201

202       The *lower crust* is composed by pelitic granulites, characterized by cm-size garnet  
203 porphyroclasts interlayered with minor marbles and rare mafic granulites and by migmatites  
204 without garnet that often show disrupted and folded leucosomes. A two-step evolution of the  
205 high-temperature metamorphism during the Hercynian and the Oligocene has been recently  
206 discussed in Gueydan et al. (2015). The *middle crust* is, from base to top, made of i)  
207 sillimanite gneisses, with some strongly stretched migmatitic leucosomes, ii) fibrolite  
208 gneisses, without leucosome but still with thin needles of sillimanite, and iii) andalusite  
209 gneisses, where andalusite is the only alumino-silicate represented. Quartzites, phyllites and  
210 schists with rare andalusite appear close to the transition with the upper crust. The *upper crust*  
211 is composed of the Malaguide rocks made of Paleozoic slates, carbonates, clastic rocks and of  
212 scattered Permian terrigenous sediments and Mesozoic carbonates.

213

214       Most of the lower and middle crust rocks have amphibolite facies parageneses and  
215 partial melting is concentrated in the lower crust. Note that high-pressure low-temperature  
216 metamorphism reported in the area (Argles et al., 1999) is related to the pre-rift history - i.e.  
217 the Tertiary nappe stacking of Malaguide and Alpujarride units (Balanyá et al., 1997) and  
218 probably also the Variscan tectonic history (Ruiz Cruz and Sanz de Galdeano, 2014). The  
219 lithological boundaries between lower, middle and upper crustal units roughly reflect the  
220 original 650° C and 350° C isotherms related to the high-temperature metamorphism, as  
221 derived by Raman spectrometry on carbonaceous material (Negro et al., 2006) (Log in Fig. 1).  
222 The different units are characterized by a fairly continuous temperature increase from upper  
223 to lower crustal levels (Argles et al., 1999; Negro et al., 2006). The temperature peak is  
224 corroborated also by the evidence of partial melting at the very base of the crust (Platt and  
225 Whitehouse, 1999) and by a total reset of low-T chronometers in almost the entire crustal

226 section, except in the shallowest levels (Monié et al., 1994; Platt et al., 2003a; Esteban et al.,  
227 2004, 2013). At the onset of rifting, the continental geotherm was rather hot and characterized  
228 with estimates of Moho temperatures of ~700°C (Negro et al., 2006; Gueydan et al. 2015) and  
229 ~800°C (Argles et al., 1999). Estimates of initial crustal thickness vary from 50 km (Argles et  
230 al., 1999) to 35-40 km (Gueydan et al. 2015). However, such differences do not affect the  
231 results of the present study.

232

233 In the Alpujarrides, a *regional foliation* defined by medium-pressure/high-temperature  
234 mineral assemblages showing a decrease in pressure (Balanyá et al., 1997; Argles and Platt,  
235 1999; Argles et al., 1999) developed during crustal thinning becoming parallel to the Crust-  
236 Mantle shear zone (Fig. 1, Précigout et al., 2013; Gueydan et al., 2015; Balanyá et al., 1997;  
237 Sánchez-Gómez et al., 1999).

238

### 239 *2.3/ Continental rifting in a supra-subduction setting*

240

241 The parallelism of foliations in crust and mantle and their development during  
242 decompression indicate that the western Alboran domain resulted from the extensional  
243 exhumation of a continental lithosphere section (Argles et al., 1999). However, Hercynian,  
244 Jurassic and Alpine ages (Montel et al., 2000; Sánchez-Rodríguez and Gebauer, 2000;  
245 Sánchez-Navas et al., 2014) suggest that the mantle part of the section likely underwent a  
246 rather complex evolution. The present paper does not aim at discussing this long and complex  
247 geological evolution of the mantle but only at constraining the deformation pattern that  
248 resulted from rifting and related exhumation of the lithospheric mantle. In spite of local  
249 complexities, the constant crosscutting relationships reported in the previously published  
250 maps (Tubía et al., 1997; Esteban et al., 2008; Mazzoli and Martín-Algarra, 2011; Précigout

251 et al. 2013; Frasca et al., 2015) show that the Lower-Miocene Ronda Peridotite Thrust  
252 postdates the mantle-crust extensional shear zone. Thinning of the continental lithosphere  
253 thinning therefore occurred before the Lower Miocene. Because thinning affected the Eocene  
254 Malaguide-Alpujarride nappe-stack (Vissers et al., 1995), an Oligocene-early Miocene age,  
255 supported by different types of data, can be argued for the rifting.

256

257 The exhumation and cooling of the mantle is dated by the garnet pyroxenites seated  
258 immediately below the extensional shear zone that separates the mantle from the crust (Sm-  
259 Nd age of  $21.5 \pm 1.8$  Ma on garnet and clinopyroxene; Zindler et al., 1983) (mean Lu-Hf ages  
260 of  $25 \pm 1$  Ma or  $24 \pm 3$  Ma on garnet; Blichert-Toft et al., 1999; Pearson and Nowell, 2004)  
261 and by leucosomes stretched within the foliation in partially molten lower crust (U-Th-Pb age  
262 of  $21.37 \pm 0.87$  Ma on monazite; Gueydan et al., 2015) (U-Pb age of  $22.0 \pm 0.3$  Ma on zircon;  
263 Platt et al., 2003b). Note that mantle was exhumed from diamond crystallization conditions  
264 (more than 150 km depth; Pearson et al., 1989; Davies et al., 1993) to garnet stability field  
265 (70-90 km) in a previous deformation event, most probably during Jurassic Tethyan rifting  
266 (Van der Wal and Vissers, 1993; Afiri et al., 2011; Garrido et al., 2011).

267

268 As summarized above, the foliation related to the *Crust-Mantle Shear Zone*, which  
269 affected simultaneously the crust and the upper part of the mantle, developed during  
270 decompression under high-temperature conditions. The same sense of shear is recorded in the  
271 lower part of the crustal envelope (Balanyá et al., 1997; Argles et al., 1999) and in garnet-  
272 spinel mylonitic zone at the top of the peridotites (Afiri et al., 2011; Balanyá et al., 1997;  
273 Précigout et al., 2013; Gueydan et al., 2015). Heating of the lower crust by the exhuming  
274 mantle was high enough to induce partial melting in the crustal rocks (Argles et al., 1999;  
275 Platt and Whitehouse, 1999). Mantle-related magmatic activity also attests for the hot

276 conditions of continental lithosphere during rifting. The crustal emplacement of tholeiitic  
277 andesites and diorite dyke swarms is likely related to mantle rising (Garrido et al., 2011), as  
278 indicated by their location, mostly in the western Betics (Esteban et al., 2013), and their major  
279 elements composition (Duggen et al., 2004). The dykes that are abundant in the Malaguide  
280 and its Triassic sedimentary cover (Fernández-Fernández et al., 2007) emplaced in the upper  
281 and middle crust during E-W stretching (see discussion in Esteban et al., 2013). The ages of  
282  $30.2 \pm 0.9$  Ma age ( $^{40}\text{Ar}/^{39}\text{Ar}$  on whole rock; Turner et al., 1999) and  $33.6 \pm 0.6$  Ma (laser  
283  $^{40}\text{Ar}/^{39}\text{Ar}$  on a plagioclase from a basalt; Duggen et al., 2004) were recently confirmed by a  
284  $33.1 \pm 1.5$  Ma age (U/Pb SHRIMP on zircons; Esteban et al., 2013). Younger lower Miocene  
285 ages (22-23 Ma: K/Ar age on whole rock; Torres-Roldán et al., 1986;  $17.7 \pm 0.6$  Ma  $^{40}\text{Ar}/^{39}\text{Ar}$   
286 age on whole rock; Turner et al., 1999) are likely related to the thermal cooling of dykes that  
287 is controlled by the country rock temperature –i.e. the latest part of the regional high-  
288 temperature metamorphic history.

289  
290 The westward rollback of the Alboran subducting slab is now largely accepted to have  
291 shaped the arcuate Betic-Rif belt (Balanyá et al., 2007; Garrido et al., 2011; Platt et al., 2013;  
292 Précigout et al., 2013; Johanesen et al., 2014) (Fig. 2). Several large-scale lines of evidence  
293 support this interpretation: i) the evolution in the volcanism types (Duggen et al., 2004), ii)  
294 the westward shift in the deposition of the foreland basin (Iribarren et al., 2009) and iii) recent  
295 tomography imaging of the mantle (Bonnin et al., 2014; Palomeras et al., 2014). Various  
296 scenarios have been proposed with significant differences in terms of timing, direction and  
297 amount of displacement (Royden, 1993; Lonergan and White, 1997; Gueguen et al., 1998;  
298 Wortel and Spakman, 2000; Gutscher et al., 2002; Faccenna et al., 2004; Rosenbaum and  
299 Lister, 2004; Vergés and Fernández, 2012). Whatsoever these differences, an overall  
300 westward trench displacement is more likely responsible for rifting in the Alboran upper plate

301 and consequent exhumation of the subcontinental mantle (Garrido et al., 2011; Hidas et al.,  
302 2013; Précigout et al., 2013) (Fig. 2a). The geochemical signature of magmatic intrusions in  
303 the Ronda Peridotites indicates that they were located in the subduction upper plate (Marchesi  
304 et al., 2012) (Fig. 2b).

305  
306 Rifting of the continental lithosphere and subsequent mantle exhumation thus occurred  
307 in a supra-subduction setting during Alboran slab rollback in Oligocene-Lower Miocene. The  
308 final thrust emplacement of the *Ronda peridotites* on top of the Iberian passive margin  
309 corresponds to a rift inversion that occurred within the subduction upper plate in lower  
310 Miocene (Hidas et al. 2013; Précigout et al., 2013) (Fig. 2c).

311

#### 312 *2.4/ Positioning of our study*

313

314 Our study focuses on an area of the Western Alboran that displays a complete section  
315 of thinned continental lithosphere. In the following, the deformation pattern is described at the  
316 crust-mantle contact and in the lower, middle and upper crust. Thus, the geometry and  
317 kinematics of rift-related structures are defined and the amount of crustal thinning across the  
318 exhumed section of thinned continental lithosphere is estimated. Furthermore, we argue that a  
319 relationship exists between the lowermost sedimentary basin and the extensional features  
320 described. Finally, the implications of our results for the mechanics of lithosphere necking  
321 and non-volcanic passive margin formation are discussed.

322

323 Following the estimates of Hidas et al. (2013) and Précigout et al. (2013), the amount  
324 of crustal thinning in the western Alboran Domain remained moderate bringing the mantle  
325 rocks at mid-crustal levels (10-15 km). Conversely, Argles et al. (1999) and Platt et al.



326 (2003a) suggest that crustal detachments brought mantle rocks to very shallow levels in the  
327 Carratraca area, leaving a strongly attenuated crustal envelope (only few km thick).  
328 Furthermore, the intrusion of gabbros in the Ronda Peridotites during the late stage of mantle  
329 thinning confirms that the whole continental lithosphere has been affected by an extreme  
330 stretching (Hidas et al., 2015).

331  
332 We focused our study on the Carratraca region because it displays: i) a complete  
333 lithospheric section (Soto and Gervilla, 1991; Argles et al., 1999; Tubía et al., 2004) and ii) an  
334 excision of the deepest parts of the crust (Chamón Cobos et al., 1972; Cano Medina and Ruiz  
335 Reig, 1990; Del Olmo Sanz, 1990). In addition, the structures related to mantle exhumation  
336 are in this region rather well preserved: iii) from reworking during rift inversion (absence of  
337 plagioclase tectonites related to hot thrusting) and iv) from the effects of Miocene extension  
338 related to the formation of the Alboran Sea basin (Comas et al., 1992).

339

340

### 341 **3/ Geometry, deformation and ages of lithosphere thinning**

342

#### 343 *3.1/ Structural map of the Carratraca massives*

344

345 The three main types of geological units of the study area are (Fig. 3a): i) mantle rocks  
346 (green; dark green highlights the grt-sp mylonites), ii) the crustal envelope (purple, brown and  
347 beige) and iii) the Alosaina basin that is filled with terrigenous deposits of Lower Miocene  
348 age (Sanz de Galdeano et al., 1993; López-Garrido and Sanz de Galdeano, 1999; Suades and  
349 Crespo-Blanc, 2013). Foliation trajectories (Fig. 3a) are drawn from 1406 measurements (see  
350 also stereoplots in Fig. S1 of “supplementary material”). The N-S trending cross-section (Fig.

351 3b) is perpendicular to the main foliation trend and cuts, along 22 km from the El Chenil  
352 region to the El Chorro region, the Carratraca mantle massives, the crustal envelope and the  
353 Alosaina basin. The crust-mantle boundary dips northward, while the Ronda Peridotite thrust  
354 is generally flat-lying. The two main high-angle normal faults of Cerro Tajo and La Robla dip  
355 southward and crosscut the Ronda peridotites and its crustal envelope.

356

357 In the crustal envelope, foliations dip northward at variable angles (Fig. 3). The main  
358 foliation trend is generally parallel to the condensed metamorphic isograds, from granulite to  
359 greenschist facies, easing the identification of lower, middle and upper crust. Significant  
360 jumps in P-T conditions occur along the Los Grenadillos fault zone (Fig. 3) that likely acted  
361 as an important mid-crustal shear zone (Argles et al., 1999). Approaching the upper crust, the  
362 intensity of foliation decreases, and lithological boundaries display more evidence of layer-  
363 parallel brittle shear. Low-angle normal faults are common and show breccias and gouges  
364 with subhorizontal striaes.

365

366 Foliation trajectories highlight the large-scale 3D geometry of the studied domain with  
367 broad anticlines cored by the peridotites massives and synclines defined by upper crustal  
368 levels and the Alosaina basin. The Cerro Tajo and La Robla faults are large-scale normal  
369 faults (Fig. 3) that bound the peridotite massives to the south and south-east, defining the  
370 three studied tilted blocks: Agua, Robla and Alosaina.

371

372 *3.2/ <sup>40</sup>Ar/<sup>39</sup>Ar on high angle normal faults*

373

374 The Cerro Tajo and the La Robla faults put in direct contact at map-scale mantle rocks,  
375 upper crustal rocks and Alosaina basin sediments (Fig. 3a), indicating large normal sense

376 offsets (up to 5 km; Fig. 3b). Both faults show a polyphased kinematic history (Soto and  
377 Gervilla, 1991; Argles et al., 1999; Esteban et al., 2004). A structural and kinematic analysis  
378 of the two faults has been carried out using classical brittle shear criteria (e.g. striae, mineral  
379 fibers, Riedel-type shear faults or deflection of foliations) observable on 10m-scale fault  
380 surfaces. The damage zones of both faults commonly show meter-scale offsets (Figs. 4d and  
381 4e). Both faults display similar kinematic patterns with a principal direction of stretching top-  
382 to-S or E, and a principal direction of shortening top-to-N-NW.

383  
384 The La Robla fault has been interpreted as a recent normal fault (Insua-Arevalo et al.,  
385 2012). Instead, the Cerro Tajo fault has been described with movements top-to-N (Argles et  
386 al., 1999; Esteban et al., 2004) or top-to-S with a dextral strike-slip component (Soto and  
387 Gervilla, 1991; Crespo-Blanc and Campos, 2001), during Burdigalian to Serravallian.  
388 Considering the large normal offset of these faults and their obvious control on the bulk  
389 structure of the area, we collected a sample of a hydrothermal tectonic breccia (Fig. 4b) along  
390 the Cerro Tajo fault (Fig. 4a) in order to date displacement along these high-angle normal  
391 faults. Millimetric white micas are pervasive in the very fine-grained matrix of the tectonic  
392 breccia (white arrow in Fig. 4c). White micas that are not observed in the protolith of the  
393 breccias define pseudomorphs after garnet (on the left in Fig. 4c) clearly indicating a neo-  
394 formation of the micas during shearing in presence of fluids (see “supplementary material”  
395 and Fig. S3 for further petrographic analysis).  $^{40}\text{Ar}/^{39}\text{Ar}$  step heating method on the white  
396 micas extracted from the matrix of the tectonic breccia give an age of  $21 \pm 0.3$  Ma (see age  
397 spectrum in Fig. 4f). Details on the method, a table summarizing the data and a complete set  
398 of isotopic results are given in the “supplementary material”.

399

400 The age of these neo-formed white micas from the tectonic breccia clearly indicates  
401 that at least part of the fault activity is older than what was previously reported. It is rather  
402 difficult to directly attribute this age to the extensional or compressional features observed in  
403 the field, since the micas were separated from fault rocks. However, the age gives a minimum  
404 boundary for the activity of the major fault zone that coincides with the  $^{40}\text{Ar}/^{39}\text{Ar}$  age of the  
405 micas composing the regional foliation related to the extension, as described above (Monié et  
406 al., 1994). On this basis and taking into account previously published data, we consider that a  
407 first extensional stage, most probably at ca. 21 Ma or older, with a top-to-S sense of shear in  
408 present-day coordinates (see also Soto and Gervilla, 1991) was then followed by a  
409 contractional reactivation with a strong component of dextral strike-slip shear during late  
410 Burdigalian (Esteban et al., 2004; Frasca et al., 2015).

411

### 412 *3.3/ The Alosaina basin*

413

414 The *Alosaina Basin* (here name as such for sake of simplicity) (Figs. 1 and 3) is a  
415 terrigenous Aquitanian to Langhian basin that includes three main groups (Serrano et al.,  
416 2007; Suades and Crespo-Blanc, 2013 and references therein). At the base, the Ciudad-  
417 Granada group is Aquitanian (22-20 Ma; dark yellow in Fig. 3). In the middle, La Viñuela  
418 group is Burdigalian (20-18 Ma) and at the top the “Neonumidian” olistostrome-type deposits  
419 are Burdigalian - Langhian (around 18-15 Ma) (Bourgeois, 1978; Martín-Algarra, 1987). A  
420 first important feature of the basin is an upward deepening trend associated to a change in  
421 sedimentation-type around 20 Ma when clasts of metamorphic rocks of the Alboran Domain  
422 and locally peridotites started to be deposited in the basin (Aguado et al., 1990). A second  
423 important character of the basin is the important amount of olistostromic deposits that have

424 been associated to thrusting in previous studies (e.g. Suades and Crespo-Blanc, 2013; Frasca  
425 et al., 2015).

426  
427 Both high-angle normal faults of Cerro Tajo and La Robla (Fig. 3) are likely  
428 contemporaneous to sediment deposition in the Alosaina basin because, first, they display  
429 dip-parallel offsets up to 5 Km and, second, the distribution of sediments is strongly linked to  
430 the location of faults. The base of the transgressive cover of the Alboran Domain (dark yellow  
431 in Fig. 3a) crops out completely only near Alosaina (Bourgeois et al., 1972a, 1972b) while  
432 several smaller outcrops are scattered along the unconformity with the Alboran metamorphic  
433 rocks (Peyre, 1974; Sanz de Galdeano et al., 1993; Serrano et al., 2007; Alcalá et al., 2014).  
434 The base of the transgressive units is lower Aquitanian, similar to the activity of the La Robla  
435 fault inferred from  $^{40}\text{Ar}/^{39}\text{Ar}$  dating of fault gouges (Fig. 4). These coeval onsets of  
436 sedimentation in the Alosaina Basin and high-angle normal faulting in La Robla and Cerro  
437 Tajo provide an important constraint for the tectonic calendar of the area.

438

439

#### 440 **4/ Strain and kinematics in crust and mantle**

441

442 A regional foliation coeval with medium-pressure/high-temperature metamorphism  
443 developed during continental rifting and related crustal thinning whose mean direction is  
444 almost parallel to the condensed metamorphic isograds in both ductile crust and mantle. Two  
445 types of shear indicators are observed: ductile in mantle and lower/middle crust and brittle-  
446 ductile in middle/upper crust. Figure 5 shows outcrop photographs of the different types of  
447 shear criteria used throughout the entire thinned section of the continental lithosphere. In  
448 Figure 6, arrows represent the mean senses of ductile shear in the lower crust and mantle

449 (violet), and in the middle crust (dark blue). Two opposite senses of ductile shear, top-to-E  
450 and top-to-W, are observed that will be discussed in detail below. Brittle/ductile shear (light  
451 blue arrows) that are observed in the upper crust and top middle crust can be related either to  
452 rifting/thinning or to local later reactivation. The criteria adopted for data selection are also  
453 discussed below.

454

#### 455 *4.1/ Kinematics of ductile deformation*

456

457 *a) In the Crust-Mantle shear zone.* In the garnet-mylonitic peridotites of the Agua and  
458 Robla blocks, C'-type shear bands often associated with strongly stretched pyroxenite layers  
459 give a sense of shear dominantly top-to-W (Fig. 5b and Fig. 6). Stretching lineations are  
460 mainly defined by stretched orthopyroxene crystals. Pyroxenites layers that are parallel to the  
461 crust-mantle boundary (Fig. 6) are deflected in the vicinity of the Cerro Tajo fault and close to  
462 El Chenil, defining a slight arcuate trend. This is responsible for the dispersion of foliation  
463 poles in the stereoplots (Fig. S1). In the lower crust, the main foliation is defined by biotite  
464 and sillimanite and locally by strongly stretched leucosome lenses. C'-type shear bands with  
465 melt injections are abundant. Sigmoidal melt pressure shadows around garnet porphyroclasts  
466 and deflection of leucosomes against C'-type shear bands indicate a sense of shear  
467 dominantly top-to-W (Fig. 5c). Stretching lineations that are commonly defined by quartz  
468 rods and elongated K-feldspar in granulites and migmatites are subhorizontal and trend  
469 mainly E-W, almost parallel to lineations in the garnet-mylonitic peridotites. In summary, in  
470 the Agua and Robla blocks, the crust-mantle boundary displays a top-to-W sense of shear at  
471 the base of the crust and at the top of the mantle (Fig. 6). Few top-to-E senses of shear are  
472 observed but only in scattered and discontinuous brittle/ductile low-angle shear zones (Fig. 6).

473

474           *b) In the middle crust*, two senses of ductile shear are observed: top-to-W in the lower  
475 part of the middle crust and top-to-NE in the upper part of the middle crust. ENE to NE-  
476 trending stretching lineations are defined by sillimanite needles in the upper part of the middle  
477 crust and by still few E-W trending quartz-sillimanite rods in the lower part. Sheath folds are  
478 common (Fig. 5f) that could partly explain the scattering in the lineation trend observed. They  
479 indicate high to extremely high values of the principal direction of stretching ( $\lambda_1 > 7.0$ ; i.e.  
480 stretching  $>600\%$ ) in the middle crust rocks. Deflection of foliations along C'-type shear  
481 bands, asymmetric boudinage of competent layers and sigma or delta tails around  
482 porphyroclasts indicate a top-to-NE sense of shear (Fig. 5e) and, less frequently, a top-to-W  
483 or SW sense of shear in the lower part of the middle crust (Fig. 5d). No superposition of the  
484 top-to-W and top-to-NE senses of shear has been identified leading Argles et al. (1999) to  
485 propose that deformation was co-axial. These two opposite senses of shear, their respective  
486 location in the lower and upper parts of the middle crust and the absence of crosscutting  
487 relationships suggest their coeval development. Moreover, the top-to-W sense of shear in the  
488 lower part of the middle crust is in continuity with a similar sense of shear in the lower crust  
489 and mantle. In the other hand, the top-to-NE sense of shear in the upper part of the middle  
490 crust is in continuity with the sense of shear at the transition between brittle and ductile crust  
491 (Fig. 6).

492

#### 493 *4.2/ Kinematics of brittle/ductile deformation at whole crustal scale*

494

495           In the transition between middle and upper crust, shear criteria combine mixed ductile  
496 and brittle features. Asymmetric *boudinage* and C'-type shear bands in andalusite-bearing  
497 quartz veins always show a top-to-E or NE sense of brittle/ductile shearing (Fig. 6g). The  
498 same features are observed in both Agua and La Robla blocks and this geographic distribution

499 indicate that the observed senses of shear are related to a single deformation event and not to  
500 late local reactivations. Moreover, top-to-NE sense of shear in these ductile-brittle shear  
501 indicators is related to veins that contain metamorphic minerals (Fig. 6g), indicating that the  
502 development of C'-type shear bands and the overall crustal thinning are related to the regional  
503 high-temperature event responsible for the main foliation.

504

505 In the upper crust, the metamorphic grade becomes very low and the absence of  
506 metamorphic mineral renders almost impossible the separation of shear sense indicators  
507 related to crustal thinning from those resulting from late reactivation. Widespread low-angle  
508 normal faults are interpreted as associated either to crustal thinning during lower Miocene  
509 (Argles et al., 1999) or to middle Miocene extensional reactivation of the Alboran domain  
510 during the formation of the Alboran basin (García-Dueñas et al., 1992).

511

512 Top-to-E-NE brittle-ductile shear indicators are also observed throughout the entire  
513 crustal section (light blue arrows, Fig. 6) and suggest that a late top-to-E sense of brittle shear  
514 affected the entire thinned continental crust with reactivation of the previous lithological  
515 boundaries (see also Argles et al., 1999).

516

517 *4.3/ A gradient of ductile crust thinning: hyper-stretching?*

518

519 Important variations in crustal thickness can be inferred from the geological map and  
520 cross-sections (Fig. 3). The Cerro Tajo and La Robla high-angle normal faults divide the  
521 thinned continental lithosphere in three tectonic blocks (Agua, Robla and Alozaina; Fig. 6) in  
522 which crustal thickness strongly differ. Map-scale crosscutting relationships indicate that  
523 variations in crustal thickness, especially in the ductile crust, were acquired before high-angle



524 normal faulting, during the Oligocene-Lower Miocene rifting. These map-scale relationships  
525 also exclude that variations in crustal thickness could be attributed to later tectonic events and  
526 in particular to thrusting.

527  
528 The variations in crustal thickness can be partly related to erosion, but only in the  
529 upper crust. A strong thickness decrease is observed in the lower crust between La Agua and  
530 La Robla blocks where thickness values are 560 m and 370 m, respectively). In the Alozaina  
531 block the lower crust is entirely absent and the middle crust preserved only in scattered lenses  
532 of few tens meters thick and locally found in direct contact with the serpentinized mantle. The  
533 middle crust shows a comparable thickness variation, from 1510 m to 950 m in average from  
534 the Agua to La Robla blocks. The total thickness variation of the ductile crust, from 1970 m  
535 to 70 m from Agua to Alozaina blocks, corresponds to a layer-perpendicular finite strain  
536  $\lambda_v=0.04$  (i.e. 96% shortening). In plane strain, this would imply a layer-parallel finite strain  
537  $\lambda_h=25.0$  (i.e. a bulk stretching amount of 2400%), what can be called “hyper-stretching”. Such  
538 amount of thinning-stretching is entirely related to ductile strain and cannot be simply related  
539 to high-angle normal faulting. The strong gradient of layer-perpendicular shortening observed  
540 cannot result only from simple shear and is necessarily a combination of layer-parallel shear  
541 and layer-perpendicular shortening (i.e. combination of pure shear and simple shear).

542  
543 *4.4/ Kinematics of brittle/ductile deformation in the hyper-stretched continental lithosphere*

544  
545 To NE of El Chenil in the Alozaina block, where the crustal section reaches its  
546 minimum thickness, a low-angle normal fault (LANF) with top to the E-NE sense of shear  
547 (Fig. 6) juxtaposes the upper crustal rocks (Malaguide schists at the top) and serpentinized  
548 mantle rocks (Fig. 8a; for detailed map and measurements see Fig. S2). Lower and middle

549 crustal rocks are strongly stretched and reduced to meter-scale lenses (Fig. 7). In addition,  
550 mylonitic peridotites are not preserved and upper crustal rocks are in direct contact with the  
551 spinel-tectonites through the LANF.

552  
553 The fault zone whose thickness changes along strike from around 10 to 30 m,  
554 maintains a characteristic structural zoning (Figs. 8a and 8b). The LANF footwall is marked  
555 by a serpentinite zone of 5 to 30 meters thick with a variable degree of brecciation and several  
556 outcrops of ophicalcites (Fig. 8c), suggesting the presence of hydrothermal fluids during the  
557 LANF activity. In the hanging-wall, the overlying crustal rocks (Fig. 8d) are more pervasively  
558 deformed and, in the core, characterized by a gouge with clasts of quartz-veins and gneisses  
559 and locally blocks of ophicalcites. In the upper crust close to the fault zone, C'-type shear  
560 bands indicate a top-to-E-NE sense of shear (Fig. 8e and stereoplots in Fig. S2), as in the  
561 middle and upper crust described in sub-section 4.2.

562  
563 The LANF is cut during the Aquitanian by the high-angle normal fault of La Robla  
564 (Figs. 5, 8a and 8b), supporting a rift-related origin. Deposition of breccias and sandstones is  
565 locally controlled by C'-type faults in the LANF hanging-wall (Figs. 8a and 8b). The  
566 sedimentary rocks deposited at the base of the Alosaina Basin (Ciudad Granada formation,  
567 Aquitanian in age) are breccias made of 1-10 cm-large angular clasts, in a sandy yellowish to  
568 brownish matrix (Fig. 8f) grading upward into quartz-rich sandstones.

569  
570 In summary, stretching reached its maximum value in the Alosaina block, leading to  
571 the complete omission of the ductile crust. The upper crustal rocks were carried on top of the  
572 serpentinitized mantle through a LANF with a top-to-E-NE sense of shear, like in the upper  
573 crust but opposite to the lower crust/upper mantle.

574

575

576 **5/ Discussion**

577

578 *5.1/ Change in the sense of shear with depth*

579

580 The ductile deformation that is responsible for the development of foliation and  
581 stretching lineations at regional scale is associated to a medium-pressure/high-temperature  
582 metamorphism characterized by andalusite and sillimanite in the middle crust and reaching  
583 partial melting in the lowermost crust in lower Aquitanian. This extensional deformation i)  
584 lead to extreme ductile crust thinning as a result of so-called hyper-stretching in an EW  
585 direction and ii) is characterized by a change in the sense of shear with depth, top-to-W in the  
586 mantle, lower crust and lower part of the middle crust and top-to-NE in the upper part of the  
587 middle crust and in the upper crust.

588

589 Top-to-E-NE brittle-ductile shear affected the entire thinned crustal section. The  
590 superposition of brittle-ductile deformation on ductile fabrics indicates that the originally  
591 ductile crust cooled during thinning and became part of the brittle crust during the extensional  
592 process. Consistently, in the hyper-stretched portion of the continental lithosphere, where the  
593 ductile crust thickness reaches its lowest value (less than few meters), a concordant top to NE  
594 shearing is observed in both strongly thinned crust and serpentized mantle.

595

596 The above conclusions can be summarized graphically with their rheological  
597 implications. At the onset of extension (Fig. 9a), the vertical profile of displacement is bell-  
598 shaped defining: i) a top-to-E sense of shear in the upper crust and top middle crust and ii) a

599 top-to-W sense of shear in the lower crust and the sub-continental mantle. In rheological  
600 terms, this change in shear sense with depth should be controlled by two strength peaks i) the  
601 brittle-ductile transition in the crust and ii) a high strength layer in the sub-Moho mantle that  
602 are separated by a weak decoupling layer (i.e. the ductile crust). Upper crust and mantle are  
603 thus mechanically decoupled and their respective deformation is accommodated by  
604 horizontal flow in the ductile crust. Two main shear zones, the “midcrustal shear zone” and  
605 the “crust-mantle shear zone” control the horizontal displacement with opposite sense of shear.  
606 After a significant amount of extension and thinning (Fig. 9b) the two shear zones  
607 progressively merge in a single one and the upper crust and mantle becomes mechanically  
608 coupled, as exemplified by the same top to NE shearing in both crust and mantle in the hyper-  
609 stretched portion of the lithosphere.

610

### 611 *5.2/ Some remarks about late high-angle normal faulting*

612

613 The Cerro Tajo and La Robla faults (Fig. 3) are primarily normal faults which still  
614 display large dip-parallel offsets, up to 5 Km. During thrusting in late Burdigalian, these  
615 faults have undergone minor contractional reactivation with a component of dextral shear  
616 (Soto and Gervilla, 1991; Frasca et al., 2015) that did not significantly affected their large  
617 normal offset. The two faults that cut trough the crust and mantle controlled block tilting in  
618 the Sierra de Agua and Sierra de la Robla. Whereas their prolongation in the Alosaina basin  
619 cannot be precisely mapped, few outcrops inside the basin indicate that the sediments are  
620 tilted in their hanging-wall (Fig. S2). The Ciudad-Granada group (Serrano et al., 2007) is the  
621 first sedimentary formation deposited in the basin (middle and late Aquitanian, 22-20 Ma). It  
622 is worth noticing that the  $^{40}\text{Ar}/^{39}\text{Ar}$  age that we obtained for the La Robla fault gouge is also  
623 Aquitanian -  $21 \pm 0.3$  Ma - i.e. synchronous with the early sediment deposition in the

624 Alosaina basin. The sediments are distributed in areas where the ductile crust is missing and  
625 their deposition is locally controlled by top-to-ENE C'-type shear bands at the base of the  
626 upper crust. The above observations and data indicate that the base of the Alosaina basin is a  
627 syn-rift deposit. The Alosaina and Malaga outcrops of the Ciudad Granada group were  
628 deposited at depths of less than 200 m, in a shelf environment (Serrano et al., 2007). The  
629 shallow environment of the deposition is especially interesting because syn-rift sediments are  
630 rarely drilled in modern passive margins (Wilson et al., 2001; Péron-Pinvidic et al., 2007).

631  
632 The two faults trend at small angle to the direction of the stretching lineations in the  
633 Agua and La Robla blocks. Consequently, these normal faults cannot have developed in the  
634 kinematic continuity of the ductile deformation recorded in the tilted blocks. In fact, available  
635 paleomagnetic data show that the studied area has undergone dextral rotations up to 60°  
636 (Feinberg et al., 1996; Villasante-Marcos et al., 2003) in relation with the formation of the  
637 Gibraltar arc during slab rollback (Fig.2). Consequently, the rift that lead to hyper-stretching  
638 and mantle exhumation has undergone large rotations during the late stages of its  
639 development and during later inversion/thrusting (Platt et al., 1995; Frasca et al., 2015).  
640 Whereas it is beyond the scope of the present paper to elaborate on these large-scale tectonic  
641 aspects, it is especially interesting to note that these high-angle normal faults, whose  
642 development is rather late in the rift history, cut through a brittle mantle layer of more than 5  
643 km thick. This demonstrates that the extreme thinning of the continental crust in the area had  
644 already permitted a fast cooling of the underlying mantle.

645

646 *5.3/ Process of lithosphere necking in the western Betics*

647

648 The geological observations and the measurements carried out in the Carratraca massif  
649 of the Ronda peridotites and its crustal envelope bring new light on the process of lithosphere  
650 necking. Whereas the evolution of necking is progressive, it is summarized for convenience in  
651 three snapshots, called early, advanced and late stages (Fig. 10).

652  
653 *The early stages of necking* (Fig. 10a), which are dated between 33 and 25 Ma by the  
654 intrusion of the Malaga tholeiites (Turner et al., 1999; Esteban et al., 2013), are characterized  
655 by the existence of a high strength subcontinental mantle. A common sense of shear top-to-W  
656 in lower crust and uppermost mantle is opposite to the sense of shear top-to-E-NE observed in  
657 the middle crust (Fig. 9a). This inversion in the sense of shear with depth strongly constrains  
658 the mechanical behavior of the rifted lithosphere. It shows that the ductile middle and lower  
659 crust played the role of a decoupling layer (i.e. *décollement*) between the brittle upper crust  
660 and the high strength subcontinental mantle. In rheological terms, it indicates that the mantle  
661 strength was higher than the strength of the overlying ductile crust and underlying deeper  
662 lithospheric mantle. In addition, strain weakening processes lead to strain localization both at  
663 the brittle-ductile transition in the crust and in the uppermost part of the subcontinental mantle  
664 (Fig 9a; see discussion in Gueydan et al. 2014). The formation of these two shear zones  
665 provides a simple explanation for the observed inversion of shear sense at the boundary  
666 between middle and lower crust (Fig 9a and 10a). Analogue and numerical models show that  
667 lithosphere necking is accommodated by intense strain localization in the subcontinental  
668 mantle (Gueydan and Précigout, 2014) and crust-mantle decoupling (crustal *décollement*)  
669 (Brun and Beslier, 1996; Nagel and Buck, 2004; Gueydan et al., 2008). Moreover, modelling  
670 also exemplifies that the necking process corresponds, at lithosphere scale, to a pure shear-  
671 type deformation accommodated by conjugate shear zones with opposite senses of shear,  
672 respectively at the base of the brittle crust and in the subcontinental mantle (Brun and Beslier,

673 1996; Gueydan et al., 2008). The strong mylonitization and thinning of the high-strength  
674 mantle (Précigout et al., 2007) allows the deeper and weaker mantle to come into contact with  
675 the crust, hence implying a strong heating of the lower crustal levels (Gueydan et al., 2015).  
676 This is in agreement with partial melting observed in the lower part of the crust (Platt and  
677 Whitehouse, 1999; Whitehouse and Platt, 2003). Such a high geotherm (Negro et al., 2006) is  
678 likely related to the supra-subduction setting of rifting in the western Alboran.

679  
680 *The advanced stages of necking* (Fig. 10b), which are dated between 25 and 22 Ma by  
681 crystallization and cooling ages within the high-temperature foliation (Esteban et al. 2013; see  
682 data compilation in section 2), are characterized by a strong thinning of the ductile crust  
683 observed from the Agua to Alozaina blocks. The extremely large gradient of stretching  
684 (2400%) can be qualified of “hyper-stretching”. The drastic decrease of the crustal thickness  
685 triggers cooling of the attenuated crust and upper part of the lithospheric mantle.  
686 Consequently, the initial strength profile with two peaks (Fig. 9a) becomes single peak in the  
687 localized zone of stretching (Fig. 9b). In other words, the crust and mantle becomes  
688 mechanically coupled and the two major shear zones (i.e. mid-crustal and crust-mantle shear  
689 zones) merge into a single, with top to NE shearing between upper crust and serpentinized  
690 mantle. Four-layers analogue models of lithosphere extension have consistently shown that  
691 the boudinage of the high strength mantle was able to bring into contact the upper part of the  
692 crust and the ductile lithospheric mantle with their already acquired opposite senses of shear  
693 (Brun and Beslier, 1996).

694  
695 *The late stages of necking* (Fig. 10c), which is dated between 22 and 20 Ma (our  
696  $^{40}\text{Ar}/^{39}\text{Ar}$  minimum age on the Cerro Tajo fault and Alozaina basin), are characterized by the  
697 development of steeply dipping normal faults and related block tilting within this domain of

698 localized stretching. Cooling thus occurs rapidly in the thinned crust and the underlying  
699 mantle, and induces a downward migration of the brittle-ductile transition in the lithospheric  
700 mantle. In the western Betics, this event of late brittle deformation is indeed superposed to  
701 previous ductile fabrics and is characterized by the formation of tilted blocks of Agua, La  
702 Robla and Alozaina, whose formation has been interrupted by thrusting and rift inversion  
703 around 20 Ma (Frasca et al., 2015). The syn-rift deposits are coeval with this late stage and  
704 mark the onset of subsidence and cooling of the stretched portion of the lithosphere.

705  
706 Whereas mantle exhumation in the western Betics occurred in a backarc setting, the  
707 observed deformation pattern is strongly controlled by the presence of a high-strength layer at  
708 the top of the lithosphere mantle. This makes it directly comparable to most passive margin  
709 formation that have lead to mantle exhumation and therefore it provides a useful field  
710 analogue for the seismic interpretation of the so-called “hyper-extended margins”.

711  
712 *5.4/ Field evidence versus models*

713  
714 In summary, the western Betics, probably better than any previous field study,  
715 exemplifies the changes in deformation processes that accommodate the progressive necking  
716 of a continental lithosphere up to mantle exhumation. The conceptual model shown in figure  
717 10 summarizes our field observations and measurements in three steps that directly result  
718 from the three major thermo-mechanical changes that occur during necking of a continental  
719 lithosphere: i) the reversal of shear sense with depth is controlled by the initial two-peak  
720 strength profile of the lithosphere, ii) extremely strong stretching in the central zone of the  
721 lithosphere neck results from the intense strain localization in the subcontinental mantle, and



722 iii) the late steep normal faults that crosscut both crust and mantle result from fast cooling in  
723 the zone of hyper-stretching.

724

725 This progressive deformation pattern that characterizes mantle exhumation in the  
726 western Betics (Figs. 11e and f) can be compared with end-member mechanical models in  
727 which the bulk deformation is either i) rather symmetrical as a result of an efficient  
728 decoupling between the upper crust and lithospheric mantle along a crustal décollement (Figs.  
729 11a and b, Brun and Beslier, 1996) or ii) starting symmetrical and becoming strongly  
730 asymmetrical, controlled by one or more lithosphere-scale detachment(s) (Figs. 11c and d;  
731 Mohn et al., 2012). For convenience, only the first stages of lithosphere necking and the last  
732 stages of mantle exhumation are represented in Figure 11. The two models differ significantly  
733 not only in terms of bulk structural symmetry but also considering the role played by ductile  
734 deformation.

735

736 In the symmetrical necking model (Figs. 11a and b, Brun and Beslier, 1996) the  
737 middle ductile crust (décollement) is sheared top to the rift axis. In the lithospheric mantle,  
738 ductile shear is localized below the high strength mantle layer with a sense of shear top away  
739 from the rift axis. During increasing extension, the thinning and rupture of the sub-Moho  
740 high-strength layer brings into contact the crust and mantle shear zones with their opposite  
741 senses of shear. The exhumed mantle only displays the sense of shear top away from rift  
742 center. In this process the same pattern of ductile shear affects the two rift margins even if the  
743 continental breakup does not occur at the center of the rift zone. In other words, a difference  
744 in bulk margin shape (i.e. width and thickness gradient) does not necessarily imply a strong  
745 internal asymmetry in terms of structure and kinematics.

746

747 In the asymmetrical hyperextension model (Figs. 11c and d, Mohn et al., 2012), even  
748 if rifting starts rather symmetrical, a strong asymmetry characterizes the final structure and  
749 deformation pattern. This model implies an excision i) of the middle crust (Mohn et al., 2012)  
750 or ii) of the high strength lithospheric mantle (Whitmarsh et al., 2001) or iii) of both  
751 middle crust and lithospheric mantle (Lavier and Manatschal, 2006). From a kinematic point  
752 of view, this model is characterized by a single sense of shear top-to-detachment hanging wall  
753 from the upper crust to the ductile mantle (Lavier and Manatschal, 2006; Whitmarsh et al.,  
754 2001). Finally and more importantly, in terms of model prediction, the two conjugate margins  
755 in this model, correspond to the hanging-wall and footwall of a lithosphere-scale detachment,  
756 respectively. Consequently, they must have strikingly different lithosphere-scale structures  
757 and deformation patterns.

758

759 The pattern of deformation associated to mantle exhumation in the western Betics  
760 (Figs. 11 e and f) has three important outcomes. First, it provides, for the first time, field  
761 evidence of a vertical reversal of shear sense as predicted by the laboratory experiments of  
762 Brun and Beslier (1996) (Fig. 11a). Second, it shows that crustal layers first underwent a  
763 strong thinning, by simultaneous faulting in the upper crust and ductile shear in the middle  
764 crust, during which extremely high values of stretching were reached prior to full mantle  
765 exhumation. Third, full mantle exhumation was accommodated by newly formed high-angle  
766 normal faults, cutting through the strongly thinned middle-lower crust and the underlying  
767 brittle mantle, and block tilting. One can expect that, if extension would not have been  
768 interrupted by rift inversion, the normal faulting and block tilting would have continued in a  
769 core complex exhumation mode of the underlying ductile mantle (Fig. 11f). Mantle  
770 detachment faults are thus more likely very late structures of the necking process.

771

772

773 **6/ Conclusions**

774

775 Our study that links field geology and geochronology leads to the following conclusions:

776

777 1/ The western Betics presents an exceptional exposure of a hyper-stretched continental  
778 lithosphere section, which includes the world largest subcontinental mantle massif (Ronda  
779 Peridotites), exhumed during the Oligocene-Lower Miocene in a back-arc tectonic setting.

780

781 2/ Variations in crustal thickness in the lithosphere section indicate amounts of stretching that  
782 may locally reach values as high as 2400%, defining a stage of hyper-stretching.

783

784 3/ Ductile deformation associated to lithosphere thinning is marked by opposite senses of  
785 shear in the lower crust/subcontinental mantle and in the upper/middle crust, highlighting a  
786 mechanical decoupling between the upper crust and the localizing subcontinental mantle. The  
787 presence of a high strength sub-Moho mantle is responsible for this change in sense of shear  
788 with depth. Whereas mantle exhumation in the western Betics occurred in a backarc setting,  
789 this deformation pattern, controlled by a high-strength layer at the top of the lithosphere  
790 mantle, makes it directly comparable to most passive margin formation that have lead to  
791 mantle exhumation. Therefore, this unique field example has a strong potential for the seismic  
792 interpretation of the so-called “hyper-extended margins”.

793

794 4/ In the course of extension, the ductile crust almost disappeared in the hyper-stretched  
795 segment (i.e. lithosphere neck) and the upper crust became mechanically coupled to the  
796 underlying serpentized mantle.

797

798 5/ On the above bases, three main steps summarize the lithosphere necking process:

799 i/ a mid-crustal shear zone and a crust-mantle shear zone that acted synchronously but  
800 with opposite senses of shear, accommodated ductile crust thinning and ascent of the  
801 subcontinental mantle,

802 ii/ hyper-stretching localized in the lithosphere neck lead to an almost disappearance  
803 of the ductile crust and brought the upper crust in contact with the subcontinental mantle, each  
804 of them with their already acquired opposite senses of shear;

805 iii/ high-angle normal faulting cutting through the Moho and related block tilting  
806 achieved the full exhumation of mantle in the domain of localized stretching.

807

## 808 **Aknowledgement**

809

810 This work was funded by the European Union FP7 Marie Curie ITN “TOPOMOD”,  
811 contract 264517. Thanks to Alexandre Pichat and Hugo Humbert for help in the field.

812 Constructive reviews by Tony Doré and Geoffroy Mohn helped improve the manuscript.

813

814

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816

817 **REFERENCES**

818

819 Afiri, A., Gueydan, F., Pitra, P., Essai, A., and Précigout, J. (2011). Oligo-miocene  
820 exhumation of the Beni-Bousera peridotite through a lithosphere-scale extensional shear zone.  
821 *Geodinamica Acta*, 24(1):49-60.

822

823 Aguado, R., Feinberg, H., Durand-Delga, M., Martín-Algarra, A., Esteras, M., and Didon, J.  
824 (1990). Nuevos datos sobre la edad de las formaciones miocenas transgresivas sobre las  
825 Zonas Internas Béticas : la Formación de San Pedro de Alcantara (Provincia de Málaga).  
826 *Revista de la Sociedad Geológica de España*, 3(1-2):79-85.

827

828 Alcalá, F. J., F. Guerrero, F. M. Martín-Martín, M., Raffaelli, G., and Serrano F. (2014).  
829 Geodynamic implications derived from Numidian-like distal turbidites deposited along the  
830 Internal–External Domain Boundary of the Betic Cordillera (S Spain). *Terra Nova*, 25:119-  
831 129.

832

833 Argles, T. and Platt, J. (1999). Stepped fibres in sillimanite-bearing veins: valid shear-sense  
834 indicators in high grade rocks?. *Journal of Structural Geology*, 21(2):153-159.

835

836 Argles, T., Platt, J., and Waters, D. (1999). Attenuation and excision of a crustal section  
837 during extensional exhumation: the Carratraca massif, Betic Cordillera, southern Spain.  
838 *Journal of the Geological Society*, 156(1):149-162.

839

840 Balanyá, J. C., Crespo-Blanc, A., Díaz-Azpiroz, M., Exósito, I., and Luján, M. (2007).  
841 Structural trend line pattern and strain partitioning around the Gibraltar arc accretionary

- 842 wedge: Insights as to the mode of orogenic arc building. *Tectonics*, 26(2):TC2005
- 843
- 844 Balanyá, J. C., García-Dueñas, V., Azañón, J. M., and Sánchez-Gómez, M. (1997).
- 845 Alternating contractional and extensional events in the Alpujarride nappes of the Alboran
- 846 domain (Betics, Gibraltar arc). *Tectonics*, 16(2):226-238.
- 847
- 848 Beltrando, M., Frasca, G., Compagnoni R. and Vitale Brovarone A. (2012). The Valaisan
- 849 controversy revisited: Multi-stage folding of Mesozoic hyper-extended margin in the Petit St.
- 850 Bernard Pass area. *Tectonophysics*. 579:17-36.
- 851
- 852 Berndt, T., Ruiz-Martínez, V. C., Chalouan, A. (2015). New constraints on the evolution of
- 853 the Gibraltar Arc from palaeomagnetic data of the Ceuta and Beni Bousera peridotites (Rif,
- 854 northern Africa). *Journal of Geodynamics*, 84:19-39.
- 855
- 856 Beslier, M.-O., Ask, M., and Boillot, G. (1993). Ocean-continent boundary in the Iberia
- 857 abyssal plain from multichannel seismic data. *Tectonophysics*, 218(4):383-393.
- 858
- 859 Beslier, M.-O., Girardeau, J., and Boillot, G. (1990). Kinematics of peridotite emplacement
- 860 during north Atlantic continental rifting, Galicia, northwestern Spain. *Tectonophysics*, 184(3):
- 861 321-343.
- 862
- 863 Blichert-Toft, J., Albarède, F., and Kornprobst, J. (1999). Lu-Hf isotope systematics of garnet
- 864 pyroxenites from Beni Bousera, Morocco: Implications for basalt origin: *Science*, 283:1303-
- 865 1306.
- 866

- 867 Boillot, G., Grimaud, S., Mauffret, A., Mougnot, D. Kornprobst, J., Mergoil, D. J., and  
868 Torrent, G. (1980). Ocean continent boundary of the Iberian margin: A serpentinite diapir  
869 west of the Galicia Bank. *Earth And Planetary Science Letters*, 48:23-34.
- 870
- 871 Boillot, G., Recq, M., Winterer, E.L., Meyer, A.W., Applegate, J., Baltuck, M., Bergen, J.A.,  
872 Comas, M.C., Davies, T.A., Dunham, K., Evans, C.A., Girardeau, J., Goldberg, G., Haggerty,  
873 J., Jansa, L.F., Johnson, J.A., Kasahara, J., Loreau, J.P., Luna-Sierra, E., Mollade, M., Ogg, J.,  
874 Sarti, M., Thurow, J., and Williamson M. (1987). Tectonic Denudation Of The Upper Mantle  
875 Along Passive Margin: A Model Based On Drilling Results (Odp Leg 103, Western Galicia  
876 Margin, Spain). *Tectonophysics*, 132:335-342.
- 877
- 878 Bonnin, M., Nolet, G., Villaseñor, A., Gallart, J., and Thomas, C. (2014). Multiple-frequency  
879 tomography of the upper mantle beneath the African/Iberian collision zone. *Geophysical*  
880 *Journal International*, 198(3):1458-1473.
- 881
- 882 Bourgois, J. (1978). La transversale de Ronda (Cordillères Bétiques, Espagne). Données  
883 géologiques pour un modèle d'évolution de l'Arc de Gibraltar. *Annales Scientifiques de*  
884 *l'Université de Besançon (France)*, 30:1-445.
- 885
- 886 Bourgois, J., Chauve, P., Lorenz, C., Monnot, J., Peyre, Y., Rigo, E. and Rivière, M. (1972a).  
887 La formation d'Alozaina. Série d'âge oligocène et aquitainien transgressive sur le Bétique de  
888 Malaga. *C. R. Acad. Sci. Paris*, 275(D):531-534.
- 889
- 890 Bourgois, J., Chauve, P., Magne, J., Monnot, J., Peyre, Y., Rigo, E. and Rivière, M. (1972b).  
891 La formation de Las Millanas. Série burdigalienne transgressive, sur les zones internes des

- 892 cordillères bétiques occidentales (région d'Alozaina-Tolox, province de Malaga, Espagne). C.  
893 R. Acad. Sci. Paris, 275(D):169-172.
- 894
- 895 Brun, J.P. and Beslier, M.O. (1996). Mantle exhumation at passive margins. Earth and  
896 Planetary Science Letters 142:161-173.
- 897
- 898 Cano Medina, F. and Ruiz Reig, P. (1990). Sheet Ardales, 1051. Geological map scale  
899 1:50000, Instituto Geológico y Minero de España, Madrid.
- 900
- 901 Célérier, B. (2013). FSA: Fault & Stress Analysis software, version 35.1, [http://www.pages-](http://www.pages-perso-bernard-celerier.univ-montp2.fr/software/dcmt/fsa/fsa.html)  
902 [perso-bernard-celerier.univ-montp2.fr/software/dcmt/fsa/fsa.html](http://www.pages-perso-bernard-celerier.univ-montp2.fr/software/dcmt/fsa/fsa.html).
- 903
- 904 Chalouan, A., Michard, A., El Kadiri, K., Frizon de Lamotte, D., Negro, F., Soto, J., and  
905 Saddiqi, O. (2008). The Rif belt. In Michard, A., Saddiqi, O., Chalouan, A., de Lamotte, D.F.,  
906 editors, Continental Evolution: The Geology of Morocco: Structure, Stratigraphy, and  
907 Tectonics of the Africa–Atlantic–Mediterranean Triple Junction. Springer: 203-302.
- 908
- 909 Chamón Cobos, C., Quinquer Agut, R., Crespo, V., Aguilar, M., and Reyes, J.L. (1972).  
910 Sheet Alora, 1052. Geological map scale 1:50000, Instituto Geológico y Minero de España,  
911 Madrid.
- 912
- 913 Clerc, C., and Lagabrielle, Y. (2014). Thermal control on the modes of crustal thinning  
914 leading to mantle exhumation: Insights from the Cretaceous Pyrenean hot paleomargins.  
915 Tectonics, 33(7):1340-1359.
- 916



- 917 Clerc, C., Lagabrielle, Y., Neumaier, M., Reynaud, J.Y., de Saint Blanquat, M. (2012).  
918 Exhumation of subcontinental mantle rocks: evidence from ultramafic-bearing clastic deposits  
919 nearby the Lherz peridotite body, French Pyrenees. *Bulletin de la Société géologique de*  
920 *France* 183(5):443-459.
- 921
- 922 Cochran, J.R., and Karner, G.D. (2007). Constraints on the deformation and rupturing of  
923 continental lithosphere of the Red Sea: The transition from rifting to drifting. In Karner, G. D.,  
924 Manatschal, G., Pinheiro, L. M., editors, *Imaging, Mapping and Modelling Continental*  
925 *Lithosphere Extension and Breakup*. Geological Society, London, Special Publications,  
926 282:265-289.
- 927
- 928 Comas, M. C., García-Dueñas, V., and Jurado, M. (1992). Neogene tectonic evolution of the  
929 Alboran Sea from MCS data. *Geo-Marine Letters*, 12(2-3):157-164.
- 930
- 931 Comas, M. C., Platt, J. P., Soto, J. I., and Watts, A. B. (1999). The origin and tectonic history  
932 of the Alboran basin: Insights from LEG 161 results. In Zahn, R., Comas, M. C. and Klaus, A.,  
933 editors, *Proceedings of the Ocean Drilling Program, Scientific Results*:555-580.
- 934
- 935 Contrucci, I., Matias, L., Moulin, M., Geli, L., Klingelhofer, F., Nouze, H., Aslanian, D.,  
936 Olivet, J.L., Rehault, J.P., and Sibuet, J.C. (2004). Deep structure of the West African  
937 continental margin (Congo, Zaire, Angola), between 5°S and 8°S, from reflection/refraction  
938 seismics and gravity data. *Geophysical Journal International*, 158:529-553.
- 939
- 940 Crespo-Blanc, A. and Campos, J. (2001). Structure and kinematics of the south iberian  
941 paleomargin and its relationship with the Flysch Trough Units: extensional tectonics within

942 the Gibraltar arc fold-and-thrust belt (Western Betics). *Journal of Structural Geology*,  
943 23(10):1615-1630.

944

945 Crespo-Blanc, A. and Frizon de Lamotte, D. (2006). Structural evolution of the external zones  
946 derived from the Flysch Trough and the south iberian and maghrebien paleomargins around  
947 the Gibraltar arc: a comparative study. *Bulletin de la Societé Géologique de France*,  
948 177(5):267-282.

949

950 Cruz San Julián, J. (1990). Sheet Teba, 1037. Geological map scale 1:50000, Instituto  
951 Geológico y Minero de España, Madrid.

952

953 Davies, G. R., Nixon, P. H., Pearson, D. G., and Obata, M. (1993). Tectonic implications of  
954 graphitized diamonds from the ronda, peridotite massif, southern Spain. *Geology*, 21(5):471-  
955 474.

956

957 Davis, M. and Kusznir, N. (2004). Depth-dependent lithospheric stretching at rifted  
958 continental margins. *Proceedings of NSF Rifted Margins Theoretical Institute*, 136:1-92.

959

960 Del Olmo Sanz, A., Moreno Serrano, F., Campos Fernández, J., Estévez, A., García-Dueñas,  
961 V., García-Rossell, L., Martín-Algarra, A., Orozco Fernández, M., and Sanz de Galdeano, C.  
962 (1990). Sheet Ronda, 1051. Geological map scale 1:50000, Instituto Geológico y Minero de  
963 España, Madrid.

964

965 Didon, J., Durand-Delga, M., and Kornprobst, J. (1973). Homologies géologiques entre les  
966 deux rives du détroit de Gibraltar. *Bulletin de la Societé Géologique de France*, 15(2):77-105.

- 967
- 968 Doblas, M. and Oyarzun, R. (1989). Neogene extensional collapse in the Western  
969 Mediterranean (Betic-Rif alpine orogenic belt): Implications for the genesis of the Gibraltar  
970 arc and magmatic activity. *Geology*, 17(5):430-433.
- 971
- 972 Duggen, S., Hoernle, K., van den Bogaard, P., and Harris, C. (2004). Magmatic evolution of  
973 the Alboran region: The role of subduction in forming the Western Mediterranean and  
974 causing the Messinian salinity crisis. *Earth and Planetary Science Letters*, 218(1-2):91-108.
- 975
- 976 Esteban, J. J., Cuevas, J., Vegas, N., and Tubía, J. M. (2008). Deformation and kinematics in  
977 a melt-bearing shear zone from the Western Betic Cordilleras (southern Spain). *Journal of*  
978 *Structural Geology*, 30(3):380-393.
- 979
- 980 Esteban, J. J., Sánchez-Rodríguez, L., Seward, D., Cuevas, J., and Tubía, J. M. (2004). The  
981 late thermal history of the Ronda area, southern Spain. *Tectonophysics*, 389(1-2):81-92.
- 982
- 983 Esteban, J.J., Cuevas, J., Tubía, J., Sergeev, S., and Larionov, A. (2011). A revised  
984 Aquitanian age for the emplacement of the Ronda Peridotites (Betic Cordilleras, southern  
985 Spain). *Geol. Mag.*, 148(1):183-187.
- 986
- 987 Esteban, J. J., Tubía, J. M., Cuevas, J., Seward, D., Larionov, A., Sergeev, S., and Navarro-  
988 Vilá, F. (2013). Insights into extensional events in the Betic Cordilleras, southern Spain: New  
989 fission-track and U-Pb SHRIMP analyses. *Tectonophysics*, 603:179-188.
- 990
- 991 Faccenna, C., Piromallo, C., Crespo Blanc, A., Jolivet, L., and Rossetti, F. (2004). Lateral

- 992 slab deformation and the origin of the arcs of the western Mediterranean. *Tectonics*,  
993 23:TC1012.
- 994
- 995 Feinberg, H., Saddiqi, O., and Michard, A. (1996). New constraints on the bending of the  
996 Gibraltar arc from paleomagnetism of the Ronda peridotite (Betic Cordilleras, Spain). In  
997 Morris A., T. D., editor, *Paleomagnetism and Tectonics of the Mediterranean Region*, volume  
998 105 of *Geol. Soc. Lond. Spec. Pubs*, The Geological Society, London:43-52.
- 999
- 1000 Fernández-Fernández, E.M., Jabaloy-Sánchez, A., Nieto, F., González-Lodeiro, F. (2007).  
1001 Structure of the Maláguide Complex near Vélez Rubio (Eastern Betic Cordillera, SE Spain).  
1002 *Tectonics*, 26:TC4008.
- 1003
- 1004 Fernández, M., Berástegui, X., Puig, C., García-Castellanos, D., Jurado, M. J., Torné, M., and  
1005 Banks, C. (1998). Geophysical and geological constraints on the evolution of the  
1006 Guadalquivir foreland basin, Spain. *Geological Society, London, Special Publications*,  
1007 134(1):29-48.
- 1008
- 1009 Flinch J.F. (1993). *Tectonic evolution of the Gibraltar Arc*. PhD thesis, Rice University,  
1010 Houston, Texas.
- 1011
- 1012 Frasca, G., Gueydan, F., and Brun, J. P. (2015). Structural record of Lower Miocene  
1013 westward Alboran Domain motion in the Western Betics (southern Spain). *Tectonophysics*,  
1014 657:1-20.
- 1015
- 1016 Frets, E. C., Tommasi, A., Garrido, C. J., Vauchez, A., Mainprice, D., Targuisti, K., and Amri,

- 1017 I. (2014). The Beni-Bousera peridotite (Rif belt, Morocco): an oblique-slip low-angle shear  
1018 zone thinning the subcontinental mantle lithosphere. *Journal of Petrology*, 55(2):283-313.  
1019
- 1020 Frizon de Lamotte, D. F., Leturmy, P., Missenard, Y., Khomsi, S., Ruiz, G., Saddiqi, O.,  
1021 Guillocheau, F., and Michard, A. (2009). Mesozoic and cenozoic vertical movements in the  
1022 atlas system (Algeria, Morocco, Tunisia): An overview. *Tectonophysics*, 475(1):9-28.  
1023
- 1024 Froitzheim, N., Pleuger, J., and Nagel, T.J. (2006). Extraction faults. *Journal of Structural*  
1025 *Geology*, 28:1388-1395.  
1026
- 1027 García-Dueñas, V., Balanyá, J. C., and Martínez-Martínez, J. (1992). Miocene extensional  
1028 detachments in the outcropping basement of the northern Alboran basin (Betics) and their  
1029 tectonic implications. *Geo-Marine Letters*, 12:88-95.  
1030
- 1031 Garrido, C. J. and Bodinier, J.-L. (1999). Diversity of mafic rocks in the Ronda Peridotite:  
1032 Evidence for pervasive melt–rock reaction during heating of subcontinental lithosphere by  
1033 upwelling asthenosphere. *Journal of Petrology*, 40(5): 729-754.  
1034
- 1035 Garrido, C. J., Gueydan, F., Booth-Rea, G., Précigout, J., Hidas, K., Padrón-Navarta, J. A.,  
1036 and Marchesi, C. (2011). Garnet lherzolite and garnet-spinel mylonite in the Ronda peridotite:  
1037 Vestiges of Oligocene backarc mantle lithospheric extension in the Western Mediterranean.  
1038 *Geology*, 39(10):927-930.  
1039
- 1040 Gueguen, E., Doglioni, C., and Fernández, M. (1998). On the post-25 Ma geodynamic  
1041 evolution of the Western Mediterranean. *Tectonophysics*, 298(1-3):259-269.

- 1042
- 1043 Gueydan, F., Morency, C., and Brun, J.-P. (2008). Continental rifting as a function of  
1044 lithosphere mantle strength. *Tectonophysics*, 460(1-4):83-93.
- 1045
- 1046 Gueydan, F. and Précigout, J. (2014). Modes of continental rifting as a function of ductile  
1047 strain localization in the lithospheric mantle. *Tectonophysics*, 612-613:18-25.
- 1048
- 1049 Gueydan, F., Précigout, J., and Montési, L. G. J. (2014). Strain weakening enables continental  
1050 plate tectonics. *Tectonophysics*, 63:189-196.
- 1051
- 1052 Gueydan, F., Pitra P., Afiri, A., Poujol, M., Essaifi, A., and Paquette, J.-L. (2015). Oligo-  
1053 Miocene thinning of the Beni Bousera peridotites and their Variscan crustal host rocks,  
1054 Internal Rif, Morocco. *Tectonics*, 34:1244-1268.
- 1055
- 1056 Gutscher, M.-A., Malod, J., Rehault, J.-P., Contrucci, I., Klingelhoefer, F., Mendes-Victor, L.,  
1057 and Spakman, W. (2002). Evidence for active subduction beneath Gibraltar. *Geology*,  
1058 30:1071-1074.
- 1059
- 1060 Hidas, K., Booth-Rea, G., Garrido, C. J., Martínez-Martínez, J. M., Padrón-Navarta, J. A.,  
1061 Konc, Z., Giaconia, F., Frets, E., and Marchesi, C. (2013). Backarc basin inversion and  
1062 subcontinental mantle emplacement in the crust: kilometre-scale folding and shearing at the  
1063 base of the proto-Alborán lithospheric mantle (Betic Cordillera, southern Spain). *Journal of*  
1064 *the Geological Society*, 170(1):47-55.
- 1065
- 1066 Hidas, K., Varas-Reus, M. I., Garrido, C. J., Marchesi, C., Acosta-Vigil, A., Padrón-Navarta,

- 1067 J. A., Targuisti, K., and Konc, Z. (2015). Hyperextension of continental to oceanic-like  
1068 lithosphere: The record of late gabbros in the shallow subcontinental lithospheric mantle of  
1069 the westernmost Mediterranean. *Tectonophysics*, 650:65-79.
- 1070
- 1071 Huisman, R. S. Beaumont, C. (2011). Depth-dependent extension, two-stage breakup and  
1072 cratonic underplating at rifted margins. *Nature*, 473:74-78.
- 1073
- 1074 Insua-Arévalo, J.M., Martínez-Díaz, J.J., García-Mayordomo, J., and Martín-González, F.  
1075 (2012). Active tectonics in the Malaga Basin: evidences from morphotectonic markers  
1076 (Western Betic Cordillera, Spain). *Journal of Iberian Geology*, 38(1):175-190.
- 1077
- 1078 Iribarren, L., Vergés, J., and Fernández, M. (2009). Sediment supply from the Betic-Rif  
1079 orogen to basins through Neogene. *Tectonophysics*, 475(1):68-84.
- 1080
- 1081 Johannesen, K. E. and Platt, J. P. (2015). Rheology, microstructure, and fabric in a large scale  
1082 mantle shear zone, Ronda Peridotite, Southern Spain. *Journal of Structural Geology*, 73:1-17.
- 1083
- 1084 Johannesen, K., Platt, J. P., Kaplan, M. S., and Ianno, A. J. (2014). A revised thermal history of  
1085 the Ronda Peridotite, S. Spain: New evidence for excision during exhumation. *Earth and  
1086 Planetary Science Letters*, 393:187-199.
- 1087
- 1088 Kornprobst, J. (1976). Signification structurale des péridotites dans l'orogène Bético-Rifain:  
1089 arguments tirés de l'étude des détritiques observés dans les sédiments Paléozoïque. *Bulletin de la  
1090 Société Géologique de France*, 3:607-618.
- 1091

- 1092 Lavier, L. L. and Manatschal, G. (2006). A mechanism to thin the continental lithosphere at  
1093 magma-poor margins. *Nature*, 440(7082):324–328.
- 1094
- 1095 Lenoir, X., Garrido, C.J., Bodinier, J.L., Dautria, J.M., and Gervilla, F. (2001). The  
1096 recrystallization front of the Ronda peridotite: Evidence for melting and thermal erosion of  
1097 subcontinental lithospheric mantle beneath the Alboran basin: *Journal of Petrology*, 42:141-  
1098 158.
- 1099
- 1100 Lister, G., Etheridge, M., and Symonds, P. (1986). Detachment faulting and the evolution of  
1101 passive continental margins. *Geology*, 14(3):246–250.
- 1102
- 1103 Lonergan, L. and White, N. (1997). Origin of the Betic-Rif mountain belt. *Tectonics*, 16:504-  
1104 522.
- 1105
- 1106 López-Garrido, A. C. and Sanz de Galdeano, C. (1999). Neogene sedimentation and tectonic-  
1107 eustatic control of the Malaga basin, south Spain. *Journal of Petroleum Geology*, 22(1):81-96.
- 1108
- 1109 Manatschal, G. (2004). New models for evolution of magma-poor rifted margins based on a  
1110 review of data and concepts from West Iberia and the Alps. *International Journal of Earth*  
1111 *Sciences*, 93:432-466.
- 1112
- 1113 Manatschal, G., Froitzheim, N., Rubenach, M., and Turrin, B. (2001). The role of detachment  
1114 faulting in the formation of an ocean-continent transition: insights from the Iberia abyssal  
1115 plain. *Geological Society, London, Special Publications*, 187(1):405-428.
- 1116



- 1117 Marchesi, C., Garrido, C. J., Bosch, D., Bodinier, J.-L., Hidas, K., Padrón-Navarta, J.A., and  
1118 Gervilla, F. (2012). A late Oligocene suprasubduction setting in the westernmost  
1119 Mediterranean revealed by intrusive pyroxenite dikes in the Ronda Peridotite (southern  
1120 Spain): *The Journal of Geology*, 120(2):237-247.
- 1121
- 1122 Martín-Algarra, A. (1987). Evolución geológica alpina del contacto entre las Zonas Internas y  
1123 las Zonas Externas de la Cordillera Bética. PhD thesis, Universidad de Granada.
- 1124
- 1125 Mazzoli, S. and Martín-Algarra, A. (2011). Deformation partitioning during transpressional  
1126 emplacement of a 'mantle extrusion wedge': the Ronda peridotites, Western Betic Cordillera,  
1127 Spain. *Journal of the Geological Society of London*, 168:373-382.
- 1128
- 1129 Mazzoli, S., Martín-Algarra, A., Reddy, S., Sánchez-Vizcaíno, V. L., Fedele, L., and Noviello,  
1130 A. (2013). The evolution of the footwall to the Ronda subcontinental mantle peridotites:  
1131 insights from the Nieves Unit (Western Betic Cordillera). *Journal of the Geological Society of*  
1132 *London*, 170:385-402.
- 1133
- 1134 Mohn G., Manatschal G., Müntener O., Beltrando M., Masini E. (2010). Unravelling The  
1135 Interaction Between Tectonic And Sedimentary Processes During Lithospheric Thinning In  
1136 The Alpine Tethys Margins. *Int. J. Earth Sci.*, 99:75-101.
- 1137
- 1138 Mohn, G., Manatschal, G., Beltrando, M., Masini, E., Kuszniir, N. (2012). Necking of  
1139 continental crust in magma-poor rifted margins: evidence from the fossil Alpine Tethys  
1140 margins. *Tectonics*, 31:TC2961.
- 1141

- 1142 Monié, P., Torres-Roldán, R., and García-Casco, A. (1994). Cooling and exhumation of the  
1143 Western Betic Cordillera,  $^{40}\text{Ar}/^{39}\text{Ar}$  thermochronological constraints on a collapsed terrane.  
1144 *Tectonophysics*, 238(1-4):353-379.
- 1145
- 1146 Montel, J.-M., Kornprobst, J., and Vielzeuf, D. (2000). Preservation of old U-Th-Pb ages in  
1147 shielded monazite: example from the Beni-Bousera hercynian kinzigites (Morocco). *Journal*  
1148 *of Metamorphic Geology*, 18(3):335-342.
- 1149
- 1150 Moulin M., Aslanian D., Olivet J.L., Klingelhoefer F., Nouzé H., Rehault J.P., Unterneuh P.  
1151 (2005). Geological constraints on the evolution of the angolan margin based on reflection and  
1152 refraction seismic data (Zaiango Project). *Geophys. J. Int.*, 162:793-810.
- 1153
- 1154 Nagel, T. J. and Buck, W. R. (2004). Symmetric alternative to asymmetric rifting models.  
1155 *Geology*, 32(11):937-940.
- 1156
- 1157 Navarro-Vilá, F. and Tubía, J. (1983). Essai d'une nouvelle différenciation des nappes  
1158 Alpujarrides dans le secteur occidental des Cordillères Bétiques (Andalousie, Espagne). *C. R.*  
1159 *Acad. Sci. Paris*, 296:111-114.
- 1160
- 1161 Negro, F., Beyssac, O., Goffé, B., Saddiqi, O., and Bouybaouéne, M. L. (2006). Thermal  
1162 structure of the Alboran Domain in the Rif (northern Morocco) and the Western Betics  
1163 (southern Spain). Constraints from Raman spectroscopy of carbonaceous material. *Journal of*  
1164 *Metamorphic Geology*, 24 (4):309-327.
- 1165
- 1166 Obata, M. (1980). The Ronda peridotite: garnet-, spinel-, and plagioclase-lherzolite facies and

- 1167 the P-T trajectories of a high-temperature mantle intrusion. *Journal of Petrology*, 21(3):533-  
1168 572.
- 1169
- 1170 Osmundsen, P. T., and Ebbing, J. (2008). Styles of extension offshore mid-Norway and  
1171 implications for mechanisms of crustal thinning at passive margins. *Tectonics*, 27:TC6016.
- 1172
- 1173 Palomeras, I., Thurner, S., Levander, A., Liu, K., Villasenor, A., Carbonell, R., and Harnafi,  
1174 M. (2014). Finite-frequency Rayleigh wave tomography of the western Mediterranean:  
1175 Mapping its lithospheric structure. *Geochemistry, Geophysics, Geosystems*, 15(1):140-160.
- 1176
- 1177 Pearson, D.G., Davies, G.R., Nixon, P.H., and Milledge, H.J. (1989). Graphitized diamonds  
1178 from a peridotite massif in Morocco and implications for anomalous diamond occurrences:  
1179 *Nature*, 335:60-63.
- 1180
- 1181 Pearson, D.G. and Nowell, G.M. (2004). Re-Os and Lu-Hf isotope constraints on the origin  
1182 and age of pyroxenites from the Beni Bousera peridotite massif implications for mixed  
1183 peridotite pyroxenite mantle sources. *Journal of Petrology*, 45: 439-455.
- 1184
- 1185 Péron-Pinvidic, G. and Manatschal, G. (2009). The Final Rifting Evolution At Deep Magma-  
1186 Poor Passive Margins From Iberia-Newfoundland: A New Point Of View. *Int. J. Earth Sci.*,  
1187 98:1581-1597.
- 1188
- 1189 Péron-Pinvidic, G., Manatschal, G., Minshull, T. A. and Sawyer, D. S. (2007).  
1190 Tectonosedimentary evolution of the deep Iberia-Newfoundland margins: Evidence for a  
1191 complex breakup history. *Tectonics*, 26:1-29.

- 1192
- 1193 Peyre, Y. (1974). Géologie d'Antequera et de la région Cordillères Bétiques (Espagne). PhD  
1194 thesis, University of Paris, France. 528 pp.
- 1195
- 1196 Platt, J. P., Allerton, S., Kirker, A., and Platzman, E. (1995). Origin of the western subbetic  
1197 arc (south Spain): palaeomagnetic and structural evidence. *Journal of Structural Geology*,  
1198 17(6):765–775.
- 1199
- 1200 Platt, J. P., Argles, T., Carter, A., Kelley, S., Whitehouse, M., and Lonergan, L. (2003a).  
1201 Exhumation of the Ronda peridotite and its crustal envelope: constraints from thermal  
1202 modelling of a P-T-time array. *Journal of the Geological Society*, 160(5):655-676.
- 1203
- 1204 Platt, J. P., Behr, W. M., Johanesen, K., and Williams, J. R. (2013). The Betic-Rif arc and its  
1205 orogenic hinterland: A review. *Annual Review of Earth and Planetary Sciences*, 41(1):313-  
1206 357.
- 1207
- 1208 Platt, J. P., and Whitehouse, M. (1999). Early Miocene high-temperature metamorphism and  
1209 rapid exhumation in the Betic Cordillera (Spain): evidence from U-Pb zircon ages. *Earth and  
1210 Planetary Science Letters*, 171(4):591-605.
- 1211
- 1212 Platt, J. P., Whitehouse, M., Kelley, S., Carter, A., and Hollick, L. (2003b). Simultaneous  
1213 extensional exhumation across the Alboran Basin: implications for the causes of late orogenic  
1214 extension. *Geology*, 31(3):251-254.
- 1215
- 1216 Précigout, J., Gueydan, F., Gapais, D., Garrido, C., and Essaifi, A. (2007). Strain localisation

- 1217 in the subcontinental mantle - a ductile alternative to the brittle mantle. *Tectonophysics*,  
1218 445(3-4):318-336.
- 1219
- 1220 Précigout, J., Gueydan, F., Garrido, C. J., Cogné, N., and Booth-Rea, G. (2013). Deformation  
1221 and exhumation of the Ronda peridotite (Spain). *Tectonics*, 32(4):1011-1025.
- 1222
- 1223 Ranero, C., and Pérez-Gussinyé, M. (2010). Sequential faulting explains the asymmetry and  
1224 extension discrepancy of conjugate margins: *Nature*, 468(7321): 294-299.
- 1225
- 1226 Reston, T.J. (2007). The extension discrepancy at North Atlantic non-volcanic rifted margins:  
1227 depth-dependent stretching or unrecognised faulting?: *Geology*, 35 :367-370.
- 1228
- 1229 Rosenbaum, G. and Lister, G. S. (2004). Formation of arcuate orogenic belts in the western  
1230 Mediterranean region. *Geological Society of America Special Papers*, 383:41-56.
- 1231
- 1232 Royden, L. H. (1993). Evolution of retreating subduction boundaries formed during  
1233 continental collision. *Tectonics*, 12:629-638.
- 1234
- 1235 Ruiz Cruz, M. D. and Sanz de Galdeano, C. (2014). Garnet variety and zircon ages in UHP  
1236 meta-sedimentary rocks from the Jubrique Zone (Alpujárride complex, Betic Cordillera,  
1237 Spain): evidence for a pre-alpine emplacement of the Ronda Peridotites. *International*  
1238 *Geology Review*, 56(7):845-868.
- 1239
- 1240 Sánchez-Gómez, M., Azañón, J. M., García-Dueñas, V., and Soto, J. I. (1999). Correlation  
1241 between metamorphic rocks recovered from site 976 and the Alpujárride rocks of the Western

- 1242 Betics. In Zahn, R., Comas, M. C. and Klaus, A., editors, Proceedings of the Ocean Drilling  
1243 Program, Scientific Results:307-317.
- 1244
- 1245 Sánchez-Gómez, M., Balanyá, J. C., García-Dueñas, V., and Azañón, J. M. (2002).  
1246 Intracrustal tectonic evolution of large lithosphere mantle slabs in the western end of the  
1247 Mediterranean orogen (Gibraltar arc). *Journal of the Virtual Explorer*, 8:23-34.
- 1248
- 1249 Sánchez-Navas, A., García-Casco, A. and Martín-Algarra, A. (2014). Pre-Alpine discordant  
1250 granitic dikes in the metamorphic core of the Betic Cordillera: tectonic implications. *Terra*  
1251 *Nova*, 26(6):477-486.
- 1252
- 1253 Sánchez-Rodríguez, L. and Gebauer, D. (2000). Mesozoic formation of pyroxenites and  
1254 gabbros in the Ronda area (southern Spain), followed by Early Miocene subduction  
1255 metamorphism and emplacement into the middle crust: U-Pb sensitive high-resolution ion  
1256 microprobe dating of zircon. *Tectonophysics*, 316(1-2):19-44.
- 1257
- 1258 Sanz de Galdeano, C., Serrano, F., López-Garrido, A. C. and Martín-Pérez, J. A. (1993).  
1259 Paleogeography of the Late Aquitanian-Early Burdigalian Basin in the Western Betic internal  
1260 zone. *Geobios*, 26(1):43-55.
- 1261
- 1262 Serrano, F., Guerra-Merchán, A., Kadiri, K. E., Sanz de Galdeano, C., López-Garrido, A. C.,  
1263 Martín-Martín, M., and Hlila, R. (2007). Tectono-sedimentary setting of the Oligocene-Early  
1264 Miocene deposits on the Betic-Rifian Internal Zone (Spain and Morocco). *Geobios*,  
1265 40(2):191-205.
- 1266

- 1267 Soto, J. I. and Gervilla, F. (1991). Los macizos ultramáficos de Sierra de las Aguas y Sierra  
1268 de la Robla como una ventana extensional (Béticas occidentales). *Geogaceta*, 9, 21-23.  
1269
- 1270 Soustelle, V., Tommasi, A., Bodinier, J.L., Garrido, C.J., Vauchez, A. (2009). Deformation  
1271 and reactive melt transport in the mantle lithosphere above a large-scale partial melting  
1272 domain: the Ronda Peridotite Massif, southern Spain. *J. Petrol.* 50:1235-1266.  
1273
- 1274 Suades, E. and Crespo-Blanc, A. (2013). Gravitational dismantling of the Miocene mountain  
1275 front of the Gibraltar Arc system deduced from the analysis of an olistostromic complex.  
1276 *Geologica Acta*, 11(2):215-229.  
1277
- 1278 Torres-Roldán, R. L. (1979). The tectonic subdivision of the betic zone (betic cordilleras,  
1279 south- ern Spain); its significance and one possible geotectonic scenario for the westernmost  
1280 alpine belt. *American Journal of Science*, 279(1):19–51.  
1281
- 1282 Torres-Roldán, R.L., Poli, G., Peccerillo, A. (1986). An Early Miocene arc-tholeiitic  
1283 magmatic dike event from the Alboran Sea - Evidence for precollisional subduction and back-  
1284 arc crustal extension in the westernmost Mediterranean. *Geologische Rundschau*, 75:219–234.  
1285
- 1286 Tubía, J., Cuevas, J., and Esteban, J. (2004). Tectonic evidence in the Ronda Peridotites,  
1287 Spain, for mantle diapirism related to delamination. *Geology*, 32(11):941-944.  
1288
- 1289 Tubía, J., Cuevas, J., and Esteban, J. (2013). Localization of deformation and kinematic shift  
1290 during the hot emplacement of the ronda peridotites (Betic Cordilleras, southern Spain).  
1291 *Journal of Structural Geology*, 50:148-160.

- 1292
- 1293 Tubía, J., Cuevas, J., and Ibarra, J. G. (1997). Sequential development of the metamorphic  
1294 aureole beneath the Ronda peridotites and its bearing on the tectonic evolution of the Betic  
1295 Cordillera. *Tectonophysics*, 279(1):227-252.
- 1296
- 1297 Tubía, J., Cuevas, J., Navarro-Vilá, F., Alvarez, F., and Aldaya, F. (1992). Tectonic evolution  
1298 of the Alpujárride complex (Betic Cordillera, southern Spain). *Journal of structural geology*,  
1299 14(2):193-203.
- 1300
- 1301 Tubía, J.M., Cuevas, J. (1986). High-temperature emplacement of the Los Reales peridotite  
1302 nappe (Betic Cordillera, Spain). *Journal of Structural Geology* 8:473-482.
- 1303
- 1304 Turner, S., Platt, J., George, R., Kelley, S., Pearson, D., and Nowell, G. (1999). Magmatism  
1305 associated with orogenic collapse of the betic-alboran domain, se spain. *Journal of Petrology*,  
1306 40(6):1011-1036.
- 1307
- 1308 Van der Wal, D. and Vissers, R. L. M. (1993). Uplift and emplacement of upper mantle rocks  
1309 in the Western Mediterranean. *Geology*, 21(12):1119-1122.
- 1310
- 1311 Van der Wal, D. and Vissers, R. L. M. (1996). Structural petrology of the Ronda Peridotite,  
1312 SW Spain: Deformation history. *Journal of Petrology*, 37(1):23-43.
- 1313
- 1314 Van Hinsbergen, D. J. J., Vissers, R. L. M., and Spakman, W. (2014). Origin and  
1315 consequences of Western Mediterranean subduction, rollback, and slab segmentation.  
1316 *Tectonics*, 33(4):393-419.



- 1317
- 1318 Vergés, J. and Fernández, M. (2012). Tethys-Atlantic interaction along the Iberia-Africa plate  
1319 boundary: The Betic-Rif orogenic system. *Tectonophysics*, 579(5):144-172.
- 1320
- 1321 Villasante-Marcos, V., Osete, M., Gervilla, F., and García-Dueñas, V. (2003). Palaeomagnetic  
1322 study of the Ronda peridotites (Betic Cordillera, southern Spain). *Tectonophysics*,  
1323 377(1):119-141.
- 1324
- 1325 Vissers, R. L. M., Platt, J. P., and van der Wal, D. (1995). Late orogenic extension of the  
1326 Betic Cordillera and the Alboran domain: A lithospheric view. *Tectonics*, 14:786-803.
- 1327
- 1328 Watts, A., Platt, J. P., and Buhl, P. (1993). Tectonic evolution of the Alboran sea basin. *Basin*  
1329 *Research*, 5:153-177.
- 1330
- 1331 Whitehouse, M. and Platt, J. (2003). Dating high-grade metamorphism - constraints from  
1332 rare-earth elements in zircon and garnet. *Contributions to Mineralogy and Petrology*,  
1333 145(1):61-74.
- 1334
- 1335 Whitmarsh, R. B. and Miles, P. R. (1995). Models of the development of the west Iberia rifted  
1336 continental margin at 40°30' N deduced from surface and deep-tow magnetic anomalies.  
1337 *Journal of Geophysical Research: Solid Earth*, 100(B3):3789-3806.
- 1338
- 1339 Whitmarsh, R. B., Manatschal, G., and Minshull, T. A. (2001). Evolution of magma-poor  
1340 continental margins from rifting to seafloor spreading. *Nature*, 413:150-154.
- 1341

1342 Wilson, R. C. L., Manatschal, G. and Wise S. (2001). Rifting along non-volcanic passive  
1343 margins: Stratigraphic and seismic evidence from the Mesozoic successions of the Alps and  
1344 western Iberia. In R. C. L. Wilson, R. B. Whitmarsh, B. Taylor and N. Froitzheim, editors,  
1345 Non-volcanic Rifting of Continental Margins: A Comparison of Evidence From Land and Sea.  
1346 Geol. Soc. Spec. Publ., 187: 429-452.

1347

1348 Wortel, M. J. R. and Spakman, W. (2000). Subduction and slab detachment in the  
1349 Mediterranean-Carpathian region. *Science*, 209:1910-1917.

1350

1351 Zindler, A., Staudigel, H., Hart, S.R., Endres, R., and Goldstein, S. (1983). Nd and Sm  
1352 isotopic study of a mafic layer from Ronda ultramafic complex. *Nature*, 304:226-230.

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### 1355 **Figure captions**

1356

1357 **Fig. 1. a)** Tectonic map of the Western Betics with foliation trajectories (modified after Frasca et al.  
1358 2015) showing the main geological units: Ronda peridotites with plagioclase tectonites (dark green  
1359 and pale green), lower, middle and upper crustal rocks above the peridotites (violet, dark brown and  
1360 pale brown), crustal rocks below the peridotites (pale blue), lower Miocene Alzaina basin (beige) and  
1361 Tortonian basins (pale yellow). Major tectonic contacts: Ronda Peridotites Thrust (RPT), Internal-  
1362 External Zone Boundary (IEZB), crust-mantle extensional shear zone (white line). Top left inset:  
1363 Location of the study area in the Betic-Rif belt with the location of the Ronda-Beni Bousera mantle  
1364 bodies. Bottom right inset: synthetic vertical section of the lithological and tectonic units of the  
1365 Western Betics. **b)** E-W trending cross-section; see location AA' in (a).

1366

1367 **Fig. 2. (a)** Subduction slab rollback setting of the Ronda peridotites in the Alboran domain, showing

1368 the present day geometry of the trench and its hypothetical position at 30 Ma. **(b)** Rifting at the front  
1369 of the subduction upper plate responsible for mantle exhumation. **(c)** Rift inversion and thrust  
1370 emplacement of the Ronda Peridotites on top of the Iberian margin (modified after Précigout et al.,  
1371 2013).

1372  
1373 **Fig. 3.** **(a)** Structural map of the Carratraca peridotitic massives (synthesized after Chamón Cobos et  
1374 al., 1972; Cano Medina and Ruiz Reig, 1990; Cruz San Julián, 1990; Del Olmo Sanz et al., 1990; Soto  
1375 and Gervilla, 1991; Argles et al., 1999; Tubía et al., 2004 and Frasca et al., 2015). Green star: position  
1376 of the sample (GFD7) collected for the  $^{40}\text{Ar}/^{39}\text{Ar}$  dating. LGFZ: Los Grenadillos Fault Zone. **(b)** NS  
1377 cross-section showing the geometry of the three blocks of Sierra Agua, Sierra de la Robla and  
1378 Alozaina. See location in (a).

1379  
1380 **Fig. 4.** Kinematics and age of high-angle normal faults. **(a)** Outcrop photograph of the Cerro Tajo fault  
1381 that put in contact middle crust gneisses with serpentized mantle rocks **(b)** Hand-specimen (GFD7)  
1382 of the fault breccia sampled for  $^{40}\text{Ar}/^{39}\text{Ar}$  dating (see location in Fig. 3.3a). White arrow: clast of the  
1383 gneiss protolith containing only biotite. **(c)** Thin-section of sample GFD7 showing the formation of  
1384 white mica pseudomorph after garnet. White arrow: very small white micas in the matrix (for details  
1385 see “supplementary material”, Fig. S3). **(d and e)** Stereoplots of fault surfaces (lower hemisphere  
1386 projection on a Schmidt net; FSA software by Célérier, 2013: version 35.2) for Cerro Tajo fault and  
1387 La Robla fault (e) (see location of the faults in Fig. 3.3a). Great circles: fault planes. Grey arrows:  
1388 direction of motion on fault surfaces. **(f)**  $^{40}\text{Ar}/^{39}\text{Ar}$  age spectrum for the mm-sized white micas  
1389 extracted from the sample GFD7 (see method and analytical data in “supplementary material”).

1390 **Fig. 5.** Types of shear indicators used in crust and mantle rocks **(a)** Landscape view (from point 5a in  
1391 Fig. 4) of the thinned lithosphere in the Sierra de Agua. Left inset: Summary of senses of shear at the  
1392 different lithosphere levels (not to scale; green: mantle; violet: lower crust; dark brown: middle crust;  
1393 light brown: upper crust) with location of photographs (see location in Fig. 4). **(b)** Top-to-W sense of  
1394 shear in the mantle: deflection of a pyroxenite layer in the Grt-Sp mylonitic foliation. **(c)** Top-to-W

1395 sense of shear in the lower crust: melt in veins and in pressure shadows around a garnet porphyroclasts  
1396 and C'-type shear-bands, locally enriched in melt, in molten granulites. (d) Top-to-W sense of shear in  
1397 the middle crust: sigmoidal stretched leucosome. (e) Top-to-E sense of shear in the middle crust:  
1398 sigmoidal stretched leucosome. (f) Eyed-type section of a sheath fold in the middle crust in the  
1399 sillimanite gneisses. (g) Top-to-E sense of shear in the middle/upper crust: C'-type brittle-ductile  
1400 shear bands in the andalusite schists.

1401  
1402 **Fig. 6.** Map of mean senses of shear in the Carratraca area. Colors of geological units like in Fig. 3.  
1403 Ductile shear in the mantle and lower crust (Violet arrows) and in the middle crust (Blue arrows).  
1404 Brittle/ductile shear, mainly observed in the upper crust (Pale blue) (For the whole set of lineation and  
1405 shear criteria see "supplementary material", Fig. S1).

1406  
1407 **Fig. 7.** Variations in crustal thickness. (a) Crustal thickness estimates in the Agua, La Robla and  
1408 Alozaina blocks (green: mantle; violet: lower crust; dark brown: middle crust; light brown: upper  
1409 crust) (See location in Fig. 3). (b) Estimates of the average thickness of the upper, middle and lower  
1410 crust in the three blocks, made from local cross-sections that take into account variations in mean  
1411 foliation dip.

1412  
1413 **Fig. 8.** Low-angle normal fault (LANF) in the hyper-stretched portion of the rift (El Chenil area; see  
1414 location in Fig. 6). (a) Landscape view of the El Chenil LANF. (b) Geological interpretation of the El  
1415 Chenil fault zone (for detailed map and measurements see Fig. S2). (c) Photograph of ophicalcite.  
1416 (d) Fault gouge in the fault core zone with clasts of quartz-veins (Q), breccia (B) and gneiss (G). (e)  
1417 C'-type shear bands (shown by blue arrows) in the upper crustal rocks indicating a top-to-E sense of  
1418 shear. (f) Sedimentary breccia at the base of the Alozaina basin with clasts of upper crustal rocks.

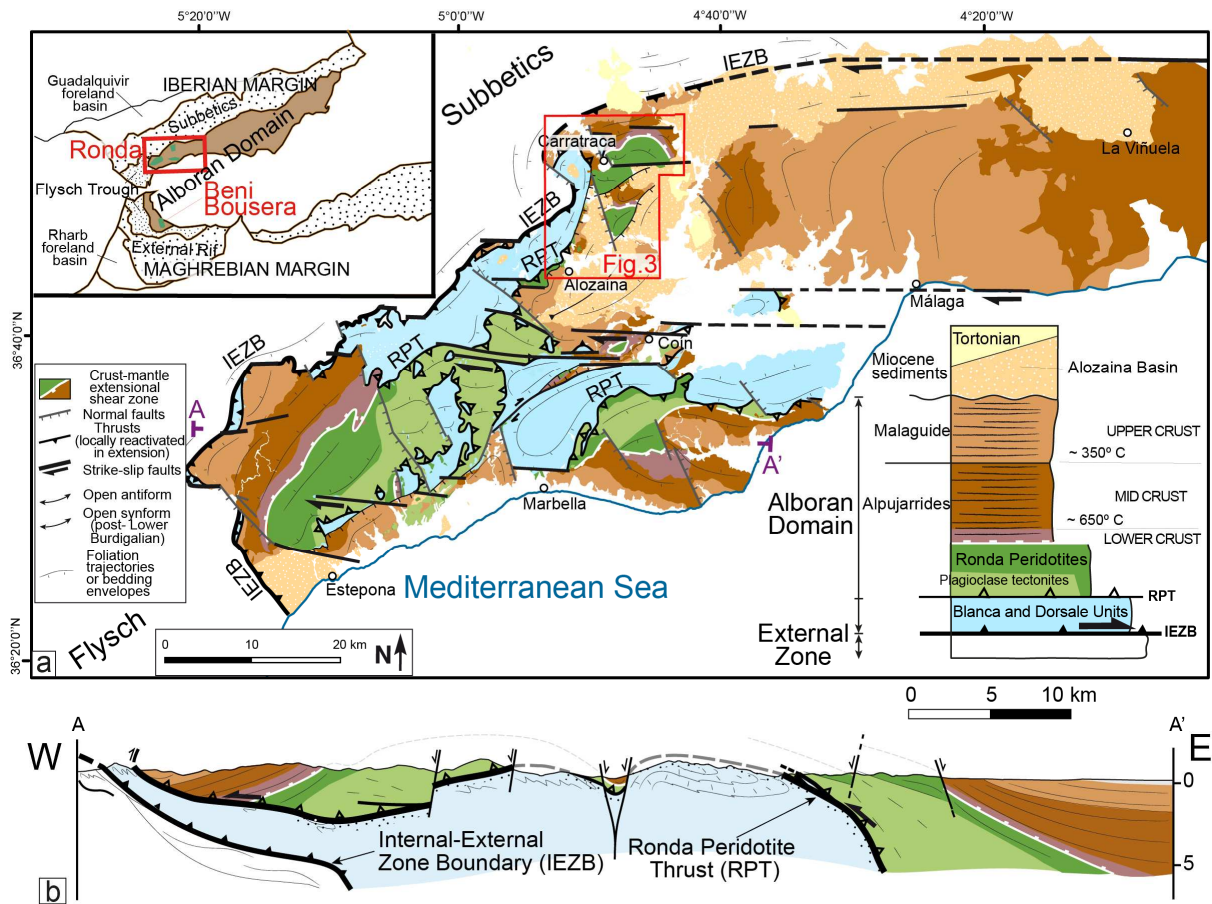
1419  
1420 **Fig. 9.** Summary of shear sense variations with depth and their implications in terms of horizontal  
1421 displacement and rheology (strength profile), (a) at the onset of extension and (b) after a strong crustal  
1422 thinning in the hyper-stretched portion of the lithosphere.

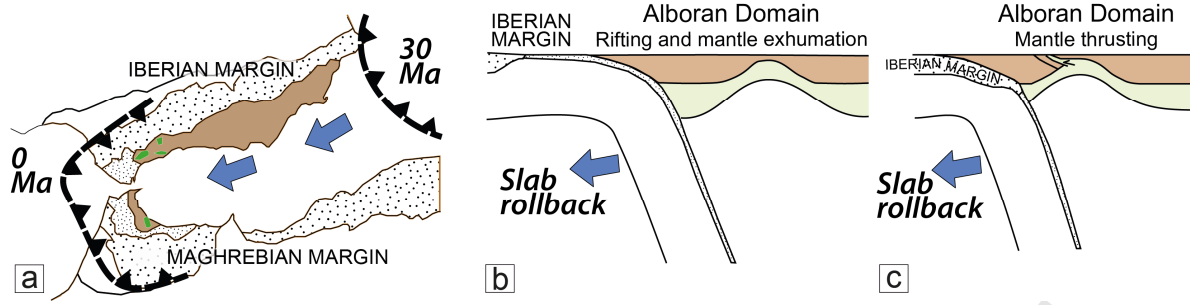
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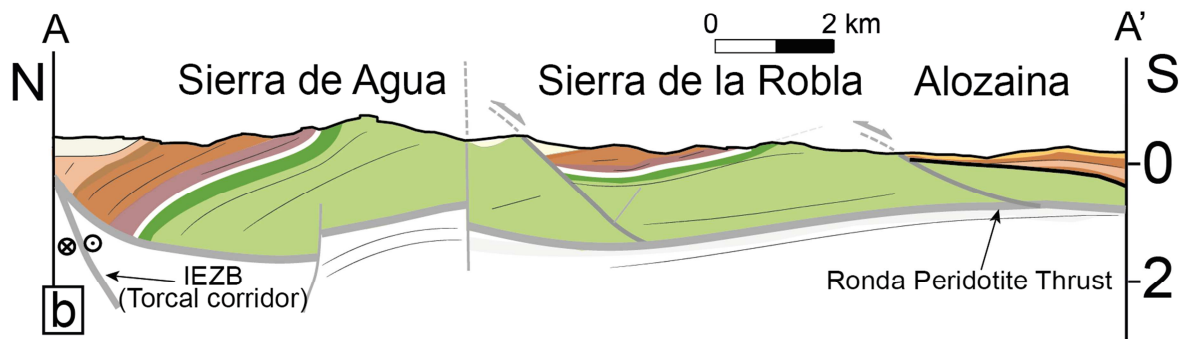
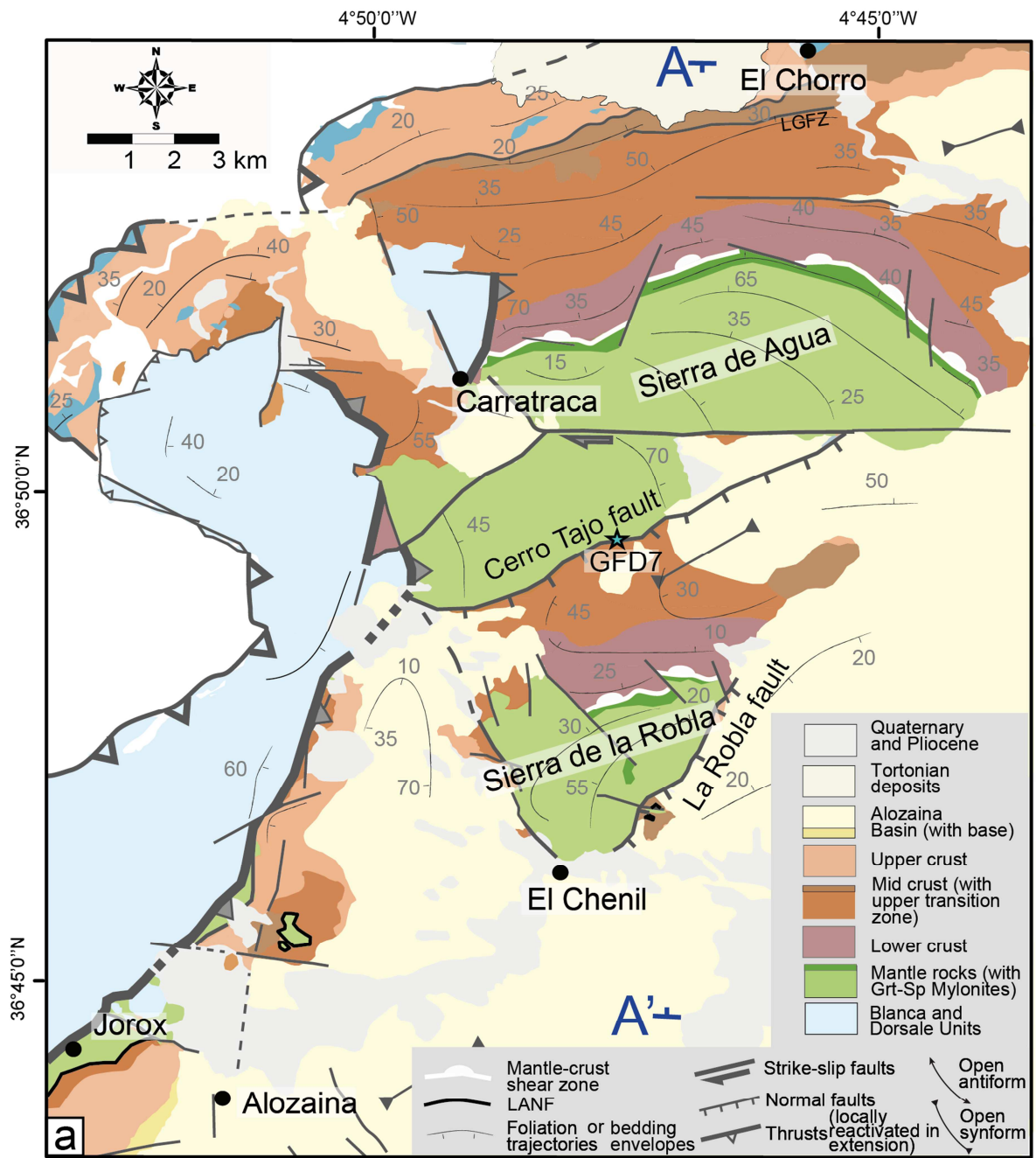
1424 **Fig. 10.** Three-stage conceptual model of lithosphere necking summarizing the progressive  
1425 deformation recorded in the crust and mantle units of the Carratraca area. (a) Early stages controlled  
1426 by the mid-crustal shear zone and the crust-mantle shear zone with opposite senses of shear. (b)  
1427 Advanced stages characterized by localization of stretching at rift center leading to an extreme  
1428 thinning of the ductile crust and omission of the lower crust. (c) Late stages of high-angle faulting  
1429 cutting through the strongly attenuated crust and the cooling mantle.

1430

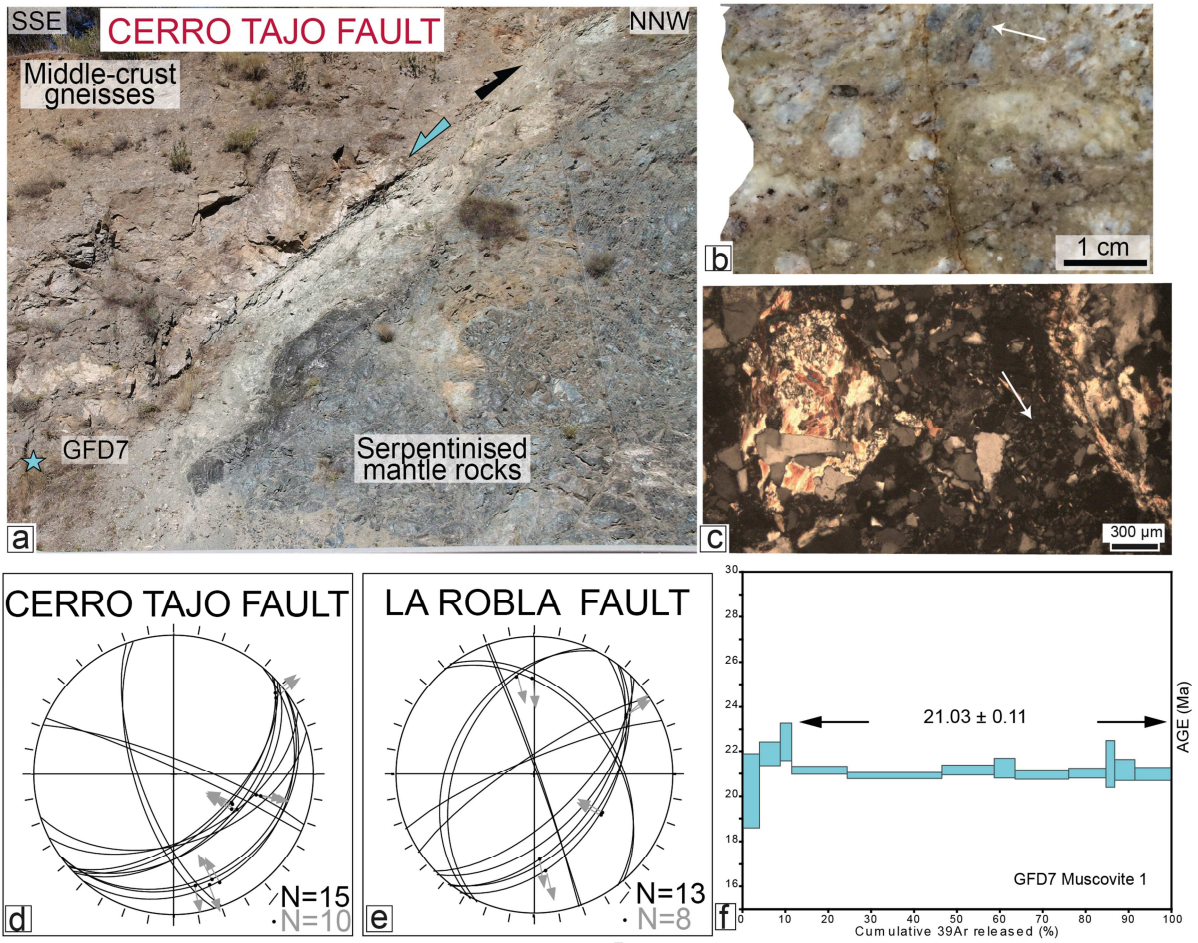
1431 **Fig. 11.** Comparison of models of lithosphere necking up to mantle exhumation, which evolve either  
1432 dominantly symmetrical (a and b) or asymmetrical (c and d), with the lithosphere-scale deformation  
1433 pattern documented in the western Betics (e and f) (as summarized in Figs. 9 and 10).

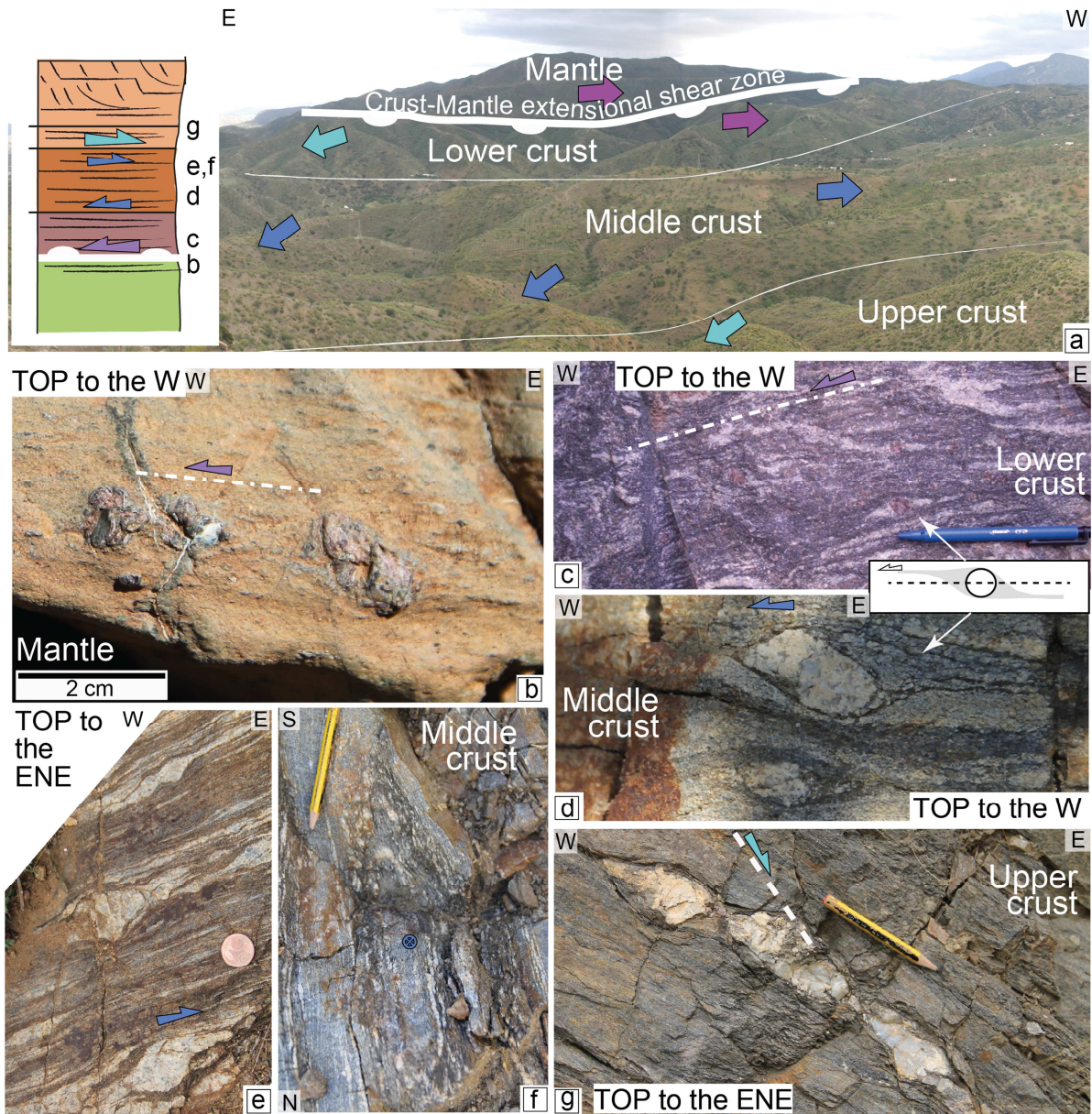


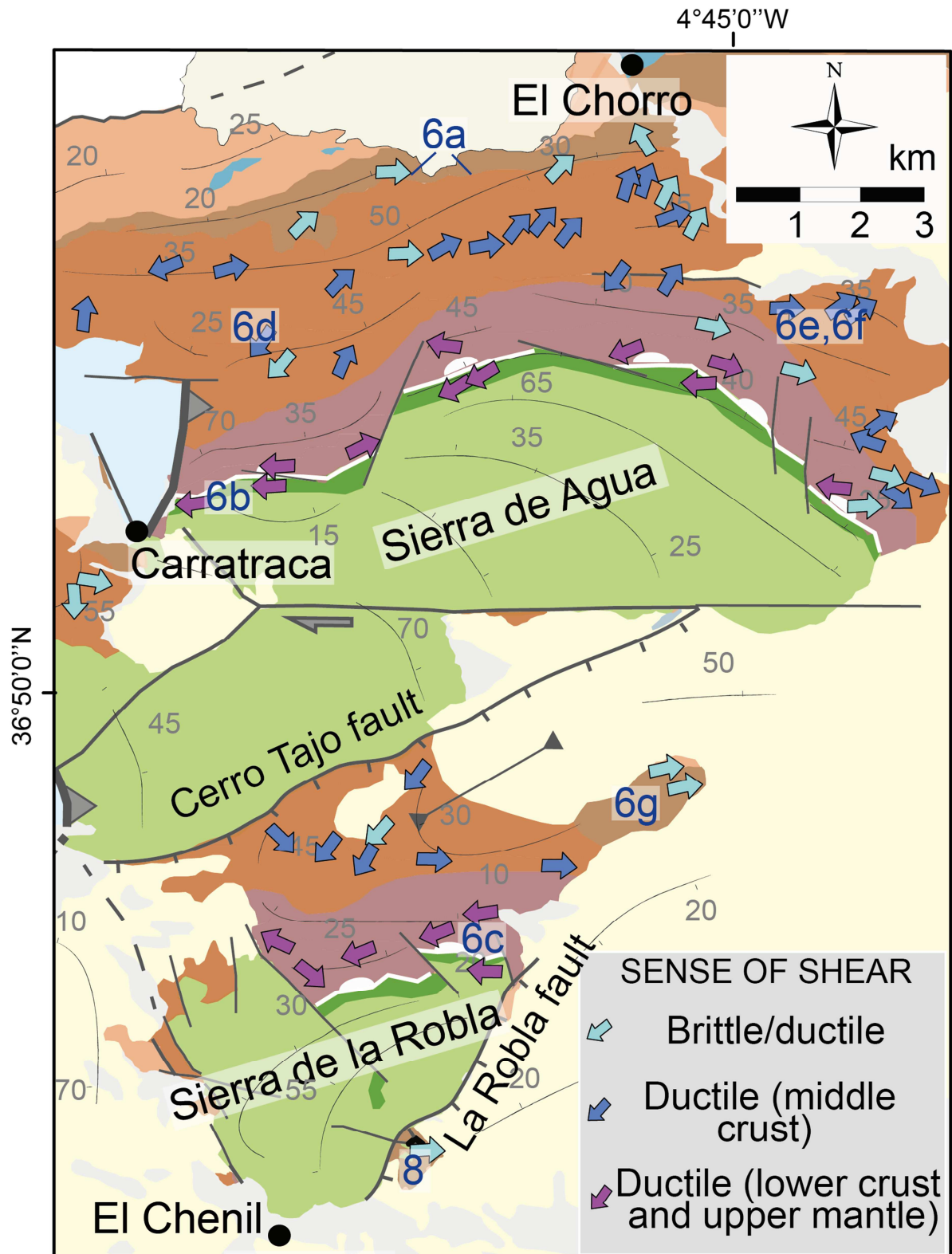


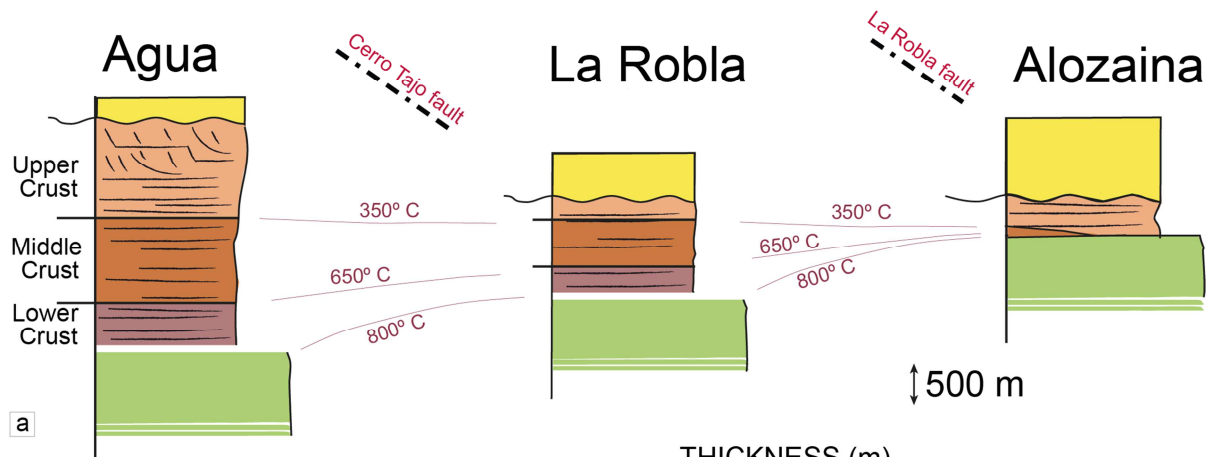










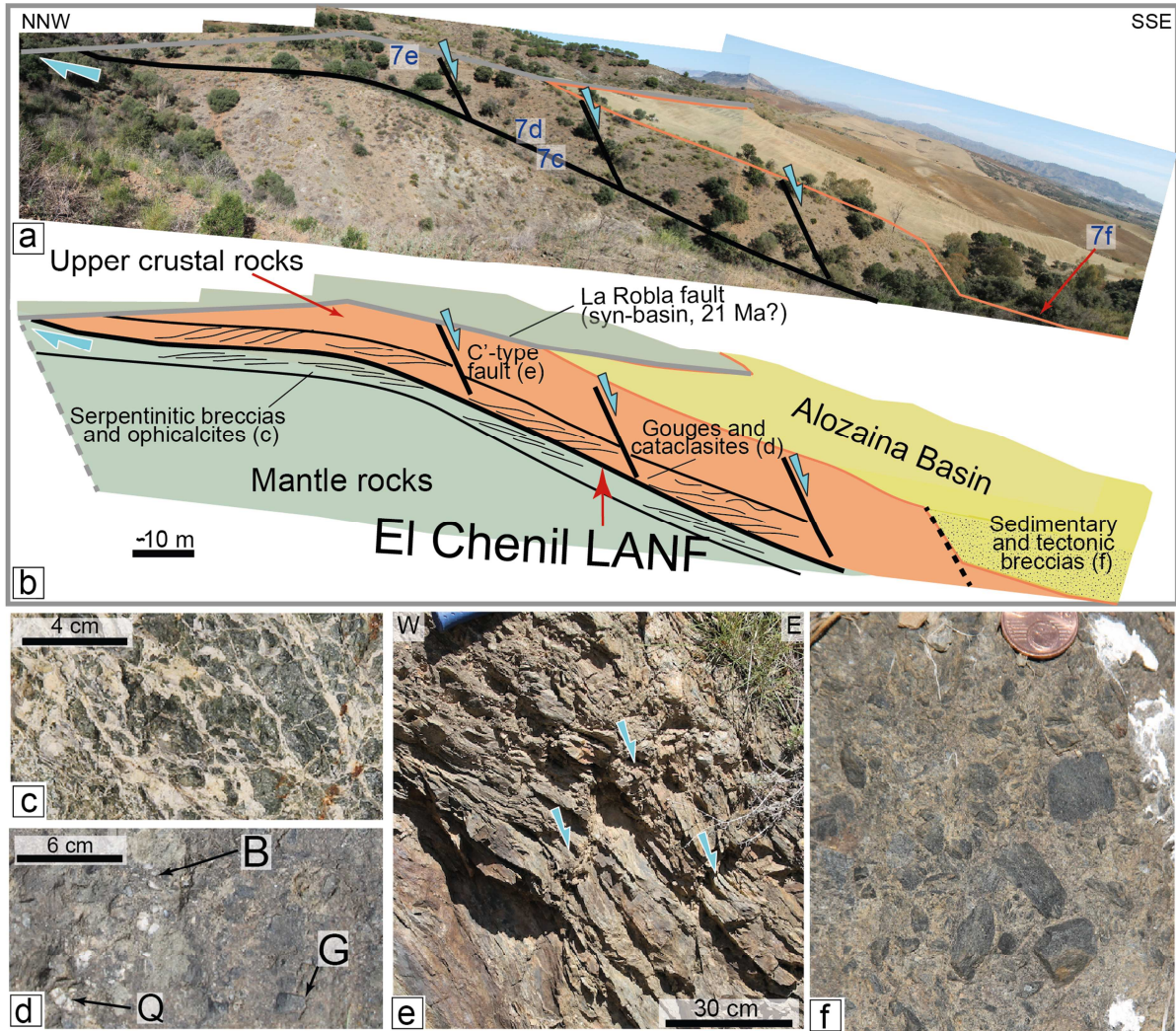


a

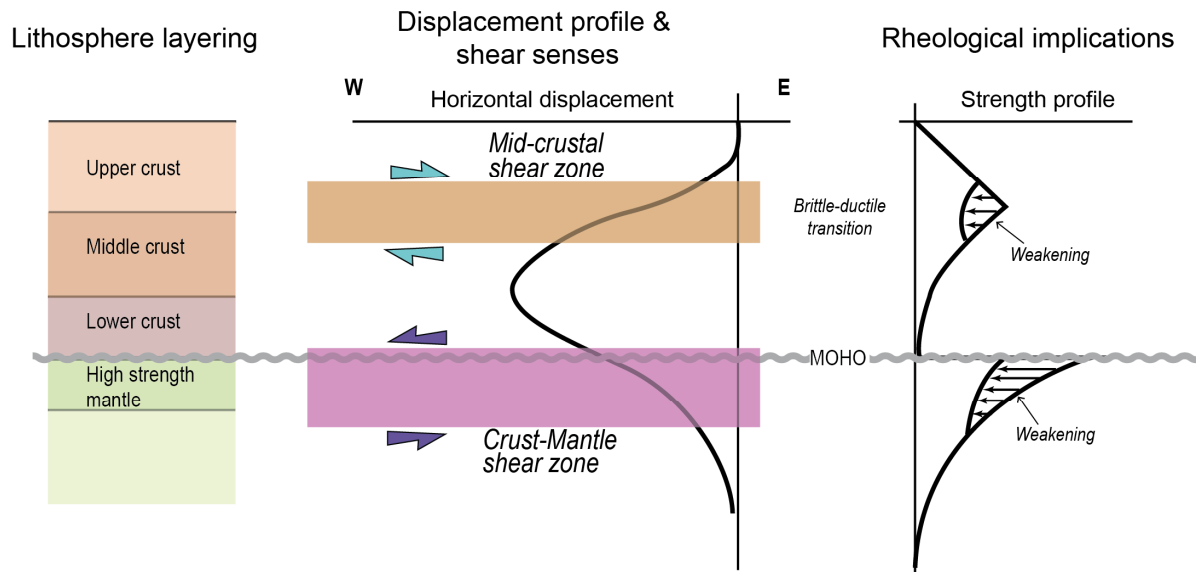
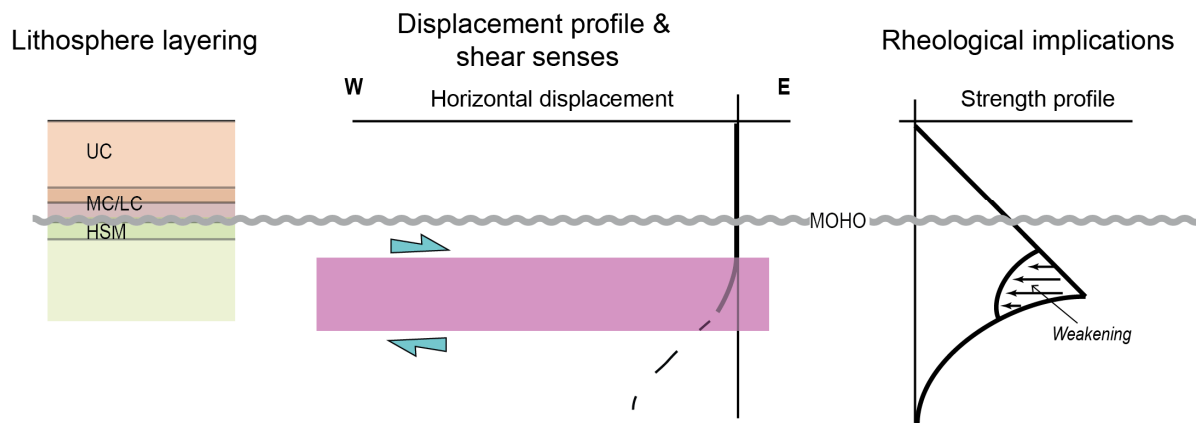
	THICKNESS (m)		
	Agua	Robla	Alozaina
Upper Crust totally or partly eroded	> 1270	0	> 480
Middle Crust	1510	950	50
Lower Crust	560	370	20
Ductile crust	1970	1070	70

b

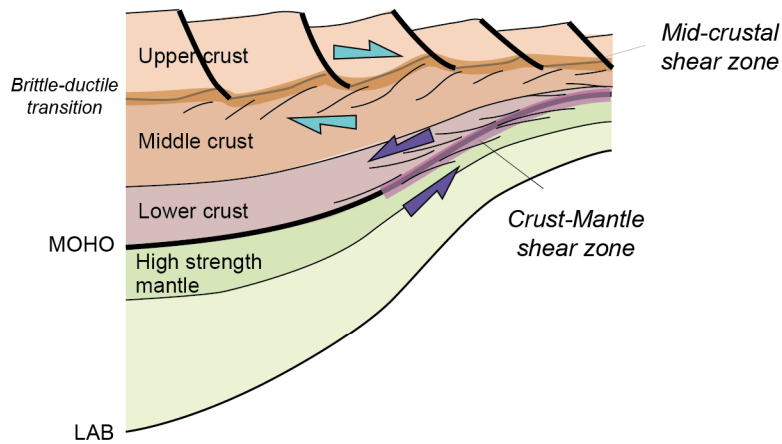
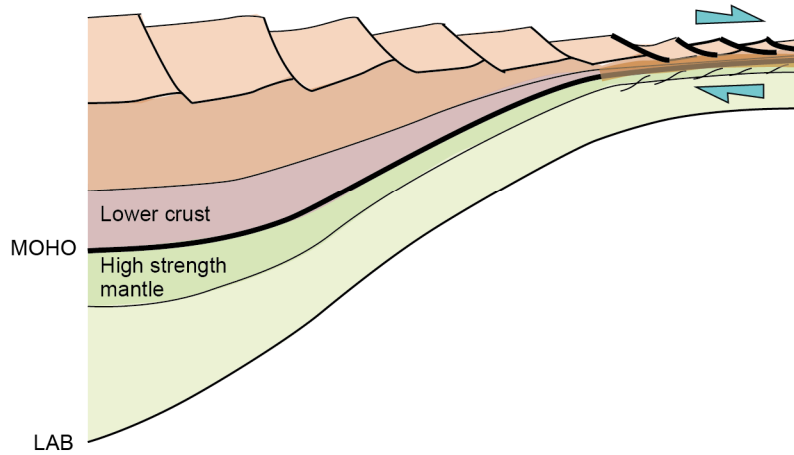
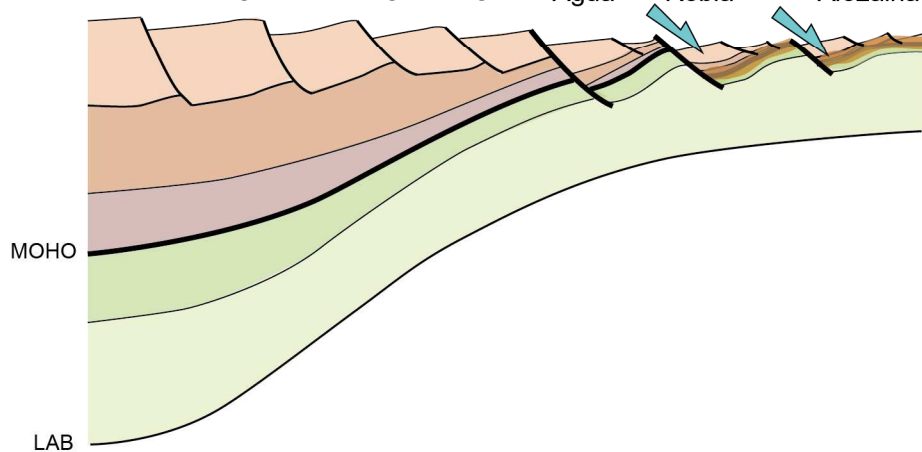
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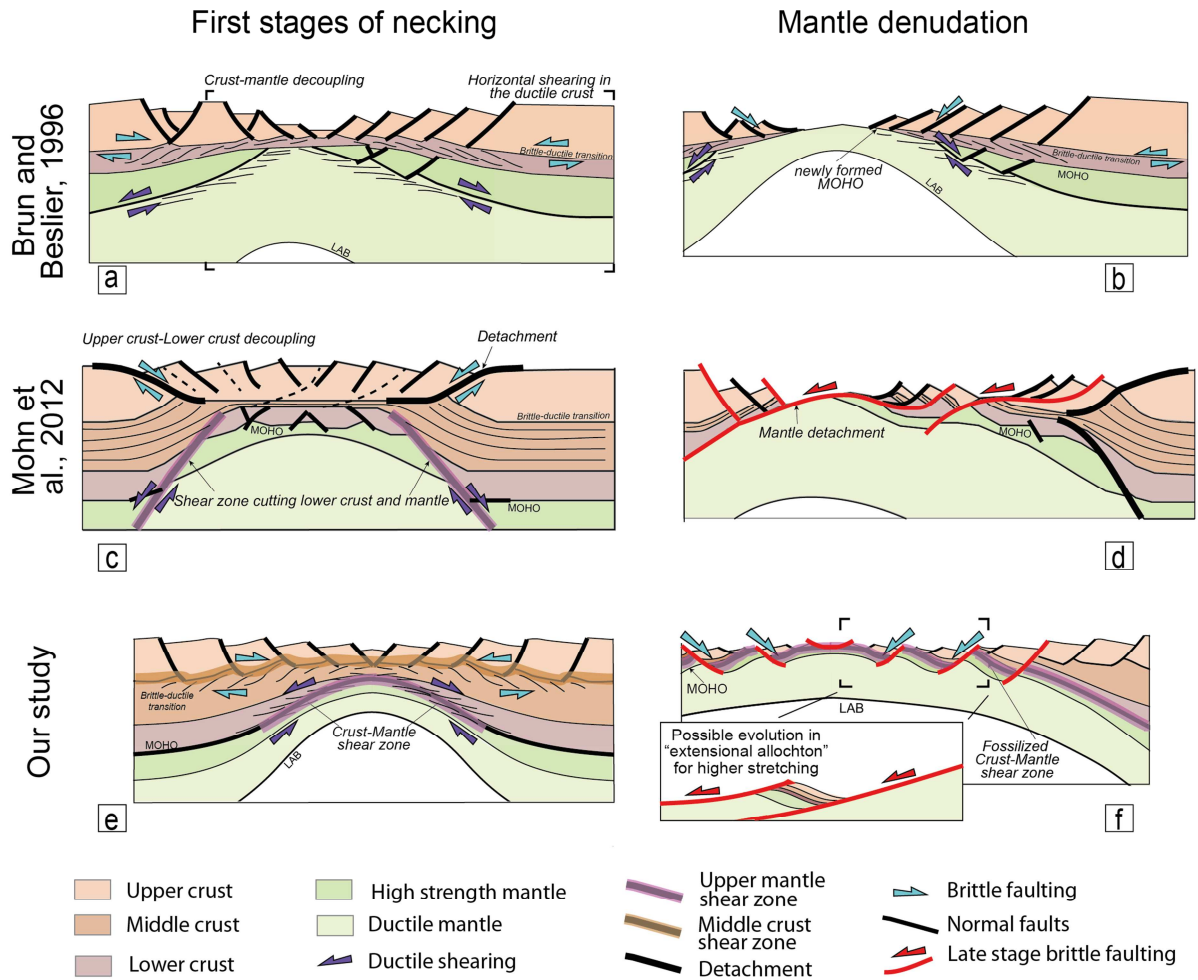


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**a) At the onset of extension****b) After a significant amount of thinning (hyper-stretching)**

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**a** 33-25 Ma; Early stages of lithosphere necking*Crust-mantle decoupling, crust heating by exhuming mantle***b** 25-22 Ma; Advanced stages of lithosphere necking*Crust-mantle coupling, localisation (hyper-stretching), onset of cooling***c** 22-20 Ma; Late stages of lithosphere necking*Mantle faulting & block tilting, cooling*



ACCEPTED TEL



**Highlights**

- Western Betics (S Spain): exceptional exposures of thinned continental lithosphere
- Progressive lithosphere necking leads to crustal stretching values larger than 2000%
- 1) mid-crustal and crust-mantle shear zones act with opposite senses of shear;
- 2) ductile crust disappears, and upper crust touches the subcontinental mantle
- 3) high-angle normal faults end mantle exhumation where stretching is localized.