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## Thermal evolution of onshore West Iberia: A better understanding of the ages of breakup and rift-to-drift in the Iberia-Newfoundland Rift

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1           1           **Thermal evolution of onshore west Iberia: a better**  
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14 **Abstract**

15

16 The age of breakup which formed the Central-North Atlantic has been debated for many  
17 decades and is still subject to debate: from ca. 150 Ma to 110 Ma. To address this issue, we  
18 have carried out a thermochronological study of the eastern margin of the rifted Iberia-  
19 Newfoundland sector. New apatite fission-track (AFT) data acquired on samples from the  
20 footwall (Variscan basement) and one sample (Triassic of the Lusitanian Basin) from the  
21 hanging wall of the principal normal fault bounding the basin. Thermal history of western Iberia  
22 can be then reconstructed since ca. 250 Ma. Fission-track ages of Variscan granitoids (whose  
23 crystallization age is  $> 275$  Ma) range between  $191\pm 8$  Ma and  $75\pm 5$  Ma, indicating that  
24 significant thermal events affected the study area during that period. Thermal inversion supports  
25 two main cooling events that we attribute to a major uplift and denudation of the Variscan  
26 basement, consistent with widespread basement-derived siliciclastic rocks of similar ages: Late  
27 Jurassic/Early Cretaceous (ca. 150-145 Ma) and late Early Cretaceous (ca. 110 Ma). From the  
28 sedimentary record in the Lusitanian Basin and the new AFT data, we deduce that: (1)  
29 temperatures  $> 70$ -110 °C affected west Iberia between ca. 190 and 150 Ma, which can be  
30 explained partly by a subsiding basement in both foot and hanging walls; (2) the main cooling  
31 event at ca. 150-145 Ma is interpreted as major rift flank uplift; (3) the ca. 110 Ma cooling event  
32 may be linked to the final evolution of the margin and onset of oceanic spreading. We infer that  
33 the AFT main cooling event at ca. 150-145 Ma reflects the break of the elastic core of the  
34 lithosphere (whole lithosphere failure = breakup) with significant rift shoulder uplift (end of  
35 Rift 1), which was followed by hyper-extension and mantle exhumation (Rift 2) and finally, by  
36 oceanic spreading (rift-to-drift) at ca. 110 Ma (onset of Rift 3).

37

38 **Keywords:** breakup; rift-to-drift; rift flank uplift and denudation; Iberia-Newfoundland  
39 Atlantic opening; whole lithosphere failure; apatite fission track thermal history

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## 42 1. Introduction

43  
44 Breakup of the continental lithosphere (rift) and sea-floor spreading (drift) are two  
45 significant phases of the supercontinent cycle (Allen and Allen, 2005). Timing is crucial but  
46 may still be subject to debate in some key areas. The Newfoundland-Iberia rift system has been  
47 considered the archetype of a magma-poor rift (Boillot et al., 1980, 1995; Manatschal and  
48 Bernoulli, 1999; Whitmarsh et al., 2001; Reston, 2009; Pereira et al., 2017) and its age of rift-  
49 to-drift stage is subject to debate as there is still no agreement on the position and the ages of  
50 the magnetic anomalies. This is partially due to the existence of a diffuse continent ocean  
51 transition due to ultra-slow opening (Dean et al., 2000). There are three main views on its  
52 timing: Late Jurassic (160 Ma-145 Ma) to Early Cretaceous (> 128 Ma) (e.g. Mauffret et al.,  
53 1989; Hiscott et al., 1990; Whitmarsh and Miles, 1995; Srivastava et al., 2000; Wilson et al.,  
54 2001; Russell and Whitmarsh, 2003; Tucholke et al., 2007), Early Cretaceous (ca. 112 Ma) (e.g.  
55 Driscoll et al., 1995), and spread over time (145–128 Ma and 112 Ma) (e.g. Péron-Pinvidic et  
56 al., 2007; Bronner et al., 2011). A strong argument has been put forward by Srivastava et al.  
57 (2000) in favour of an Upper Jurassic rift-to-drift stage in the south (rifting was diachronous  
58 along the strike, i.e. becoming younger as you move from south to north), because the oldest  
59 magnetic anomaly that has been found west of Iberia and in the Grand Banks of Canada is M20,  
60 i.e. ca. 147 Ma. Subsidence has been analysed to estimate the successive rift stages in western  
61 Iberia but has not brought decisive answers (e.g., Stapel et al., 1996; Rasmussen et al., 1998;  
62 Leinfelder and Wilson, 1998; Alves et al., 2002, 2009, 2018). Three rift phases have been  
63 recorded: Rift 1 (during the Late Triassic > 202 Ma), Rift 2 (during most of the Jurassic, ca.  
64 200 to 150 Ma) and Rift 3 (at the end of the Jurassic, especially in the transition from the Jurassic  
65 to the Cretaceous, ca. 145 Ma). Additional data is required to clarify this issue.

66 If the concept of rift-to-drift is clear and well established as the moment of first creation  
67 of basaltic oceanic crust, the concept of breakup is unclear, and is often assumed to be  
68 equivalent to rift-to-drift. Breakup may also be the stage when one continental lithosphere  
69 becomes two separate plates, followed by hyper-extension and an intervening transition zone  
70 made up of exhumed mantle and intrusive gabbro and mafic dykes (Soares et al., 2012). To  
71 resolve this ambiguity, in this study we have used the mechanical definition of lithosphere  
72 (elastic in time scales of Ma; e.g. Burov and Diament, 1995) to consider that a lithospheric plate  
73 is broken (breakup moment) when its elastic core vanishes, i.e. its elastic thickness becomes  
74 zero. This corresponds to the concept of “whole lithosphere failure” (WLF) (e.g. Kusznir and

75 Park, 1982; see also Marques and Podladchikov, 2009 and figure S1 in the supplementary  
76 material). Following this definition, we should see the effects of WLF on the vertical  
77 movements of rift shoulders (Beaumont et al., 1982; Weissel and Karner, 1989), encompassing  
78 the whole basin. Weissel and Karner (1989) suggested that rift flank uplift during extension  
79 may result from mechanical unloading of the lithosphere and a consequent isostatic rebound.  
80 This mechanism is preferred here as an alternative to explanations for rift flank uplift involving  
81 thermal processes and magmatic thickening of the crust, because we are studying a magma-  
82 poor rift.

83 The aim of this paper is to understand the behaviour of the rift shoulder, thus allowing us  
84 to deduce the age of WLF from rift shoulder uplift. We investigated this uplift, using low-  
85 temperature thermochronology coupled with the onshore geological records. AFT  
86 thermochronology has long been used to detect upper crustal rock uplift and cooling, in  
87 particular on passive margins (e.g., Gallagher and Brown, 1997).

## 88 89 90 **2. Geological context and previous LTT results**

91 In western Iberia, the basement is mostly made up of pre-Variscan rocks which have been  
92 deformed, metamorphosed and intruded by granitoids during the Variscan orogeny in the late  
93 Palaeozoic (e.g., Quesada and Oliveira, 2019). The core of the orogen, known as the Central  
94 Iberian Zone, is mostly composed of plutonic rocks (granitoids), which is where most of the  
95 samples were collected. Immediately to the south lies the Ossa-Morena Zone, which is a major  
96 terrane accreted to the southern margin of the Central Iberian Zone (Quesada and Oliveira,  
97 2019), and where we collected some samples from gabbro and granite. Available U-Pb data  
98 indicates that the sampled gabbro and various granites crystallized prior to 295 Ma (Table 1,  
99 and references therein).

100 The Iberian margin has been considered to be a lower plate, magma-poor, hyperextended  
101 continental lithosphere, the structure of which is variable both across and along the strike (e.g.,  
102 Péron-Pinvidic et al., 2007; Alves et al., 2009) across the strike (E-W) because of the distance  
103 to the rift shoulders (proximal and distal sedimentation and structure), and along the strike (N-  
104 S) because rifting seems to have been diachronic, propagating from south to north (e.g.  
105 Whitmarsh and Miles, 1995; Tucholke et al., 2007; Alves et al., 2009).

106 The sedimentary infilling can be summarized as follows. From the Late Triassic to the  
107 Late Cretaceous, fault-controlled regional subsidence lead to the deposition of a thick Mesozoic  
108 cover that reaches >6 km in the deepest parts of the onshore LB (Wilson et al., 1989; Kullberg,  
109 2000; FrogTech, 2012; Fig. 3). The master fault bounding the LB in its central-northern portion  
110 is the reactivated Porto-Tomar-Badajoz Fault (PTBF) striking N-S to NNW-SSE (PTBF, Fig.  
111 1) from Porto to Tomar, and in its southern portion is the reactivated NNE-SSW late-Variscan  
112 fault system (Marques et al., 2002). In the early stages of rifting (Hettangian), with restricted  
113 salt-water circulation and which was dominated by evaporation, a thick evaporitic sequence  
114 was deposited, which reaches almost 3,000 m thickness in some places. During the Early and  
115 Middle Jurassic (ca. 201 to 164 Ma), marine deposits accumulated in the current onshore  
116 central-northern LB. Rocks are mostly marls, marly limestones and limestones. The early Late  
117 Jurassic (Oxfordian to early Kimmeridgian) is still marine (limestones and marls), but with  
118 intercalations of fine calcarenites. The late Kimmeridgian and Tithonian lie unconformably (ca.  
119 20° angular unconformity) on previous Jurassic rocks, and are mostly composed of calcarenites,  
120 calcarenites with limestone boulders, sandstones with mixed siliceous and calcareous boulders,  
121 and some intercalated marls. This late Upper Jurassic detrital carbonate sequence demonstrates  
122 that erosion was occurring within the LB, either by basin uplift or eustatic sea level change.  
123 The total thickness of the Jurassic in the central-northern LB, measured at the outcrop and in  
124 deep wells, is ca. 3,000 m (Rocha et al., 1981). Together with the Hettangian evaporites, the  
125 total thickness of the pre-Cretaceous sediments may be between 4,000 and 5,000 m. The  
126 Cretaceous initiates with the most prominent unconformity in the LB lie on deformed pre-  
127 Cretaceous rocks (mostly Jurassic), where coarse and very coarse conglomerates, with siliceous  
128 clasts and boulders originated in the basement to the east. The conglomerates are overlain by  
129 siliceous sandstones, whose grain size decreases in the upper section (Trincão et al. 1989).

130 We have observed no Jurassic and Cretaceous sediments lying on the currently exposed  
131 Variscan basement east of the PTBF, except in a small area north of the western end of the  
132 thrust bounding the Serra da Estrela to the north (Fig. 1), and close to the eastern border of the  
133 LB. This small basin, the Lousã Basin, has deposits that include feldspathic sandstones from  
134 the Albian to the Late Campanian (Cunha, 1999).

135  
136 Three main Mesozoic rifting stages have been considered in the literature for the on and  
137 offshore LB (see Fig. 2; e.g., Wilson et al., 1989; Rasmussen et al., 1998; Alves et al., 2002,  
138 2006, 2009; Pereira and Alves, 2012): Rift 1 – Triassic (> 201 Ma); Rift 2 – Sinemurian–early

139 Pliensbachian (ca. 190 Ma) and Wilson et al. (1989) suggests a single extensional event from  
140 Triassic to Callovian; Rift 3 – late Oxfordian (ca. 155 Ma). All of them indicate the progressive  
141 subsidence and deposition of marine sediments. The latter has been considered the main  
142 subsidence episode in the LB (Leinfelder and Wilson, 1998; Stapel et al., 1996; Alves et al.,  
143 2002), and the precursor of ocean spreading in the Tagus Abyssal Plain (Wilson et al., 1989).  
144 Pinheiro et al. (1992) suggested a Valanginian age (ca. 135 Ma) for the initiation of ocean  
145 spreading. Alternatively, the oldest magnetic anomaly in western Iberia may be M20  
146 [Srivastava et al., 2000], which may indicate a late Tithonian age (ca. 147 Ma) for the rift-to-  
147 drift stage in southwest Iberia. A Rift 4 event has been suggested (e.g. Alves et al., 2009), which  
148 lasted from the Jurassic/Cretaceous boundary to mid Lower Cretaceous. The onset of seafloor  
149 spreading between Iberia and Newfoundland has long been debated. Based on the first  
150 undisputed oceanic magnetic anomaly (M3), Whitmarsh and Miles (1995) suggested an age of  
151 ca. 128 Ma for the rift-to-drift stage, whereas Tucholke et al. (2007) and Péron-Pinvidic et al.  
152 (2007) proposed a ca. 112 Ma age based on drilling results and seismic stratigraphic arguments.

153 Little thermochronological data is available to record the post-Variscan history of the  
154 basement in western Iberia. Pereira et al. (1998) presented AFT ages for two Variscan granites  
155 of the northern domain (Central Iberian zone) and four sandstones from different stratigraphic  
156 levels (Upper Carboniferous, Triassic, Upper Jurassic and Lower Cretaceous) from the LB, and  
157 concluded that there had been active erosion of the basement throughout the Jurassic and the  
158 Cretaceous, but the driving forces are not considered. Very little data is available for the  
159 southern domain (Ossa Morena Zone), but the data does confirm younger ages and significant  
160 cooling (112 to 70 Ma; Stapel 1999; Vázquez-Vílchez et al., 2015).

### 161 162 163 **3. Sampling and methodology**

164 Given the objective of analysing the vertical motion of the rift shoulder, we collected 17  
165 samples of Variscan granites along transects in the basement east of the westward-dipping  
166 master bounding fault (footwall of the reactivated PTB shear zone as normal fault - PTBF), and  
167 one granitic sample included in a very coarse Triassic conglomerate in the LB (hanging wall of  
168 the PTBF). The sampled plutonic rocks include monzonite, granite, granodiorite and gabbro  
169 related to syn-tectonic and post-tectonic Variscan magmatic events. Granitic boulders (COI3)  
170 within conglomerates attributed to the Triassic were collected in Coimbra, in the hanging-wall



171 of the master fault bounding the LB, very close to the contact with the Variscan basement.

172 AFT analysis was performed using the external detector method and zeta calibration  
173 (Hurford and Green, 1983) at the GEOPS laboratory (Université Paris Saclay). Classical  
174 mineral separation procedures were followed, and apatite grains were picked using a binocular  
175 microscope. Samples were then irradiated in the Garching facility of the Technische Universität  
176 München (Germany) with a fluence of  $5 \times 10^{15}$  n/cm<sup>2</sup>. Etching conditions were 5 M HNO<sub>3</sub> for  
177 20 seconds at 20±0.5 °C for apatite crystals, and 40% HF for 20 minutes at 20±1 °C for  
178 muscovite external detectors. Central ages were calculated using a  $\zeta$ -value for the dosimeter  
179 glass CN5 of 359±8, calibrated by multiple analyses of IUGS apatite age standards (Durango  
180 and Fish Canyon). Only crystals with sections parallel to the *c*-axis were analysed. Confined  
181 track-lengths were measured using only TINTs (track-in-track) under a 100x dry objective with  
182 a digitizing tablet linked via a drawing tube attached to the microscope. The angle to the *c*-axis  
183 was measured for each track length, and Dpar values are available to determine the  
184 structural/chemical control of apatite on fission-track annealing (Carlson et al., 1999; Barbarand  
185 et al., 2003).

186 The thermal history was determined by QTQt modelling (Gallagher, 2012), which allows  
187 for the inverting of AFT data using a Markov chain Monte Carlo sampling method. The  
188 inversion code incorporates the multi-compositional fission-track annealing model (Ketcham  
189 et al., 2007). The modelling starts with an initial randomly chosen time-temperature path and  
190 set of kinetic parameters, for which a probability that the model fits the data is calculated, and  
191 the model with the highest probability is retained. This procedure is repeated many times (100  
192 000 iterations were used in this study), providing many models with their associated  
193 probabilities that allow us to calculate statistical models, and ultimately the best-fit model.

194

195

#### 196 4. Results

197 AFT ages in granites range between 75±4 and 191±8 Ma (Figs. 1 and 4; Table 1) with  
198 high homogeneity within samples (see radial plot distributions in the supplementary  
199 information, figures S2a-b-c). AFT ages are then much younger than the crystallization age of  
200 the granites (> 295 Ma; Table 1). This is consistent with old rocks being dated by a low-  
201 temperature dating technique (AFT). They appear also younger than the ages measured for  
202 samples lying at the surface since Permo-Triassic times. The measured age variations can be

203 grouped by geographical distribution of the samples considering (1) rapid change on three main  
204 units bounded by main faults, and (2) younger ages towards the north for samples located in the  
205 western and central units. Three units where FT ages are coherent have been defined and they  
206 are separated by main late-Variscan faults (Fig. 1): a western unit, west of the Chaves fault; a  
207 central unit between the Chaves and Vilariça faults; and an eastern unit located east of the  
208 Vilariça fault. The central unit is characterized by the oldest ages ( $137\pm 8$  to  $191\pm 8$  Ma), with  
209 younger ages towards the north. The western unit is characterized by younger ages ( $75\pm 4$  to  
210  $157\pm 8$  Ma), but with similar even younger ages towards the north. The eastern unit displays  
211 young ages ( $78\pm 4$  to  $110\pm 9$  Ma), with two older ages ( $151\pm 8$  and  $162\pm 8$  Ma). A sharp age  
212 transition is observed between the central and eastern units, across the main Vilariça fault. The  
213 same observation can be made over a short distance ( $\sim 10$  km) across the Chaves fault, between  
214 sample AN ( $137\pm 8$ ) located in the central unit and sample BAR ( $75\pm 4$  Ma) in the western unit.  
215 The definition of the different units depends on the number and distribution of samples, but we  
216 are confident that the density of our sample size is enough to build a pertinent age distribution  
217 and identify different units. Samples from the southern domain (Ossa-Moreno Zone) have been  
218 considered separately, as they are far away from the other samples, but display similar young  
219 ages ( $88\pm 16$  and  $97\pm 7$  Ma). All samples except sample T pass the chi-square test, indicating  
220 that a single age population exists for each sample. No relationship was found between AFT  
221 ages and current topographic elevation.

222 Mean horizontal confined track lengths in samples for which enough tracks have been  
223 measured (minimum 50, 13 samples) range between 12.0 and 12.8  $\mu\text{m}$ , and 13.4  $\mu\text{m}$  for just  
224 only one sample (SCD, which also displays the oldest age) and have a unimodal distribution.  
225 The standard deviation of the mean length ranges between 1.3 and 1.8  $\mu\text{m}$ , except for sample  
226 SCD ( $SD = 1.0$   $\mu\text{m}$ ). All samples have similar  $D_{\text{par}}$  (1.4-1.7  $\mu\text{m}$ ), characteristics of apatite  
227 crystals with a high fluorine content.

### 228 229 *Thermal modelling*

230 Inverse modelling of the AFT data was produced for all samples. For most of the samples,  
231 two boxes have been input for initial thermal inversion at the beginning and the end: (1) The  
232 start of the low-temperature history at 300 Ma, and at a minimum temperature of 120  $^{\circ}\text{C}$   
233 considering crystallization and cooling ages (Hildenbrand et al. 2021). The strict definition of  
234 a crystallization age is not necessary because FT ages are below 200 Ma and crystallization  
235 ages are  $>295$  Ma (Table 1). (2) At the surface (temperature of ca. 20  $^{\circ}\text{C}$ ) at present-day (time

236 = 0 Ma). To the south of the central unit, near to the city of Coimbra (Fig. 1) and east of the  
237 PTBF, the presence of Aptian (Trincão et al., 1989) continental siliciclastic rocks are deposited  
238 unconformably on top of the basement, and it is here considered to be a geological constraint  
239 for samples SCD and NEL to be close to the surface before the Aptian.

240 Results of inverse modelling are presented for the different units, starting with the central  
241 unit where the presence of Lower Cretaceous sedimentary rocks unconformably on the Variscan  
242 basement helps to understand some possible scenarios.

243

#### 244 *Central unit*

245 Sample SCD is the oldest in age ( $191\pm 8$  Ma), and has the longest mean track length  
246 ( $13.4\pm 0.1$   $\mu\text{m}$ ) with a narrow distribution (confined horizontal track-length histograms are  
247 available in figure S3 of the SI). Modelling shows that this sample underwent low temperatures  
248 ( $<40$   $^{\circ}\text{C}$ ) during the Mesozoic (Fig 5-A). Modelling of sample NEL, also close to the surface  
249 in the Aptian, produced a different trend, with a high temperature during the beginning of its  
250 history (Fig. 5-B). Sample VTA produced a non-indicative thermal history, the best fits  
251 corresponding to linear cooling from the starting point to the present-day. These contrasting  
252 thermal histories are not consistent with the small size of the area in which these samples  
253 outcrop, so begging for an alternative explanation is needed.

254 Whilst considering the thermal history of sample SCD, we tested thermal inversion for  
255 samples NEL and VTA assuming the presence of these samples at near surface temperatures  
256 ( $20\pm 10$   $^{\circ}\text{C}$ ) during Triassic times. As expected from basin evolution, the temperature increased  
257 up to  $\sim 90$   $^{\circ}\text{C}$  during rifting after the Triassic, and then decreased as discussed in this work (Figs.  
258 5-C and 5-D). The northernmost sample (AN) recorded a higher temperature ( $\sim 100$   $^{\circ}\text{C}$ ) that has  
259 erased the early history, and then underwent a decrease in temperature during the Cretaceous  
260 (Fig. 5-E). The thermal history obtained for samples NEL and VTA is very similar to the one  
261 obtained for sample COI3, which corresponds to granitic boulders in Triassic conglomerates  
262 outcropping currently in Coimbra, very close to the PTBF. After deposition, the temperature  
263 increased during most of the Jurassic to reach  $\sim 100$   $^{\circ}\text{C}$ , followed by a cooling period during  
264 Cretaceous and Cainozoic times.

265 Modelling indicates that rocks from the Central Unit underwent post-Triassic heating up  
266 to  $\sim 90$ - $100$   $^{\circ}\text{C}$ , except for sample SCD for which the temperature is probably lower ( $<50$   $^{\circ}\text{C}$ ).  
267 Cooling is not well understood, which may be due to partial annealing of most of the samples,

268 in which tracks which formed before the temperature peak were partly preserved, thus blurring  
269 the thermal history. The signal is clearer for sample AN (where most of the tracks which formed  
270 prior to the temperature peak were erased) and cooling started at ~160-150 Ma.

271

#### 272 *Western unit*

273 Three samples (BAR, GRJ, ALH) with a minimum measured length of 50 tracks have  
274 been processed. All models are characterized by high temperatures during the Jurassic (~100  
275 °C), capable of resetting the FT system. Cooling occurred from ~160-150 Ma (ALH) or later at  
276 ~130-110 Ma (Fig. 6). Although less constrained, a similar thermal history is obtained for  
277 sample SVG. Cooling appears complex for sample BAR with two episodes: Lower (in common  
278 with the other samples) and beginning in the Upper Cretaceous and Neogene. This may be due  
279 to the relatively low MTL ( $12.0 \pm 0.1 \mu\text{m}$ ), implying that the sample stayed at relatively high  
280 temperature for longer.

281

#### 282 *Eastern unit*

283 Similarl to the modelling of samples from the western unit, modelling of samples MTG1,  
284 SRT and CST from the eastern unit shows the existence of a high temperature (~100 °C) during  
285 the Jurassic, which has fully reset the former thermal history of the samples (Fig. 7). Peak  
286 temperature was followed by cooling that started at ~120-110 Ma. Cooling does not appear to  
287 be linear for samples MTG1 and SRT, with faster rates during the Neogene. Modelling of  
288 samples AC and T, which have older FT ages, is not indicative, as best fits correspond to linear  
289 cooling between 300 Ma and present-day. The addition of a constraint assuming the presence  
290 of sample T close to the surface during the Triassic generates a history compatible with the  
291 other samples and shows that the temperature was higher during the Jurassic.

292 Fission-track data and models support overall (except samples SCD and T) high  
293 temperatures during the Jurassic and cooling during Late Jurassic and Cretaceous times with  
294 cooling starting at ~160-150 Ma or ~120-110 Ma.

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## 297 **5. Discussion**

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### 5.1. Origin of post-Variscan high (~100 °C) temperatures

Fission-track data and thermal models indicate overall high temperatures during the Jurassic, except for samples SCD and T. Most of the samples are basement rocks that may have risen towards the surface after the Jurassic, but this explanation ignores the fact that some samples were already close to the surface at the end of the Triassic. Three alternative processes are generally considered to explain the temperature increase and are discussed below: (1) heat introduced into the system by magmatic activity, (2) circulation of thermal fluids in disequilibrium with the host rock, and (3) burial.

1. Three magmatic events have been described on the onshore sector of the West Iberian Margin: the first event was at the Triassic-Jurassic transition (~200 Ma), the so-called Central Atlantic Magmatic Province (CAMP) of the Atlantic large igneous province, which has generated large volumes of basalts that have affected the whole Atlantic margin (e.g. Marzoli et al., 1999). The ca. 600 km long Messejana-Placencia dyke crossing Iberia from SW to NE is significant evidence of this episode, which may have significantly affected the neighbouring basement temperature (Dunn et al., 1998). Except for sample SCD, which shows rapid cooling initiated close to ~200 Ma, all samples have recorded a high temperature during the Jurassic, i.e. long after the CAMP event. If we assume that the origin of these temperatures is related to the CAMP, we also have to consider a protracted and pervasive period of heating too. Given that no evidence of pervasive heating has been reported, it thus appears difficult to have a near-surface temperature of ~100°C in such a large continental area and over such a long period of time, at least for the entire Jurassic. A second thermal event has been described around 145 Ma (Ferreira and Macedo, 1983; Grange et al., 2008; Mata et al., 2015), but it corresponds to low volumes of igneous intrusions restricted to dykes far from the basement studied here, and it is therefore unable to reset the FT ages. A third, Upper Cretaceous (ca. 85 to 75 Ma), event (Rui Miranda et al., 2009; Grange et al., 2010) is quite voluminous, but very localised west and south of Lisbon (Sintra, Sines and Monchique igneous massifs), and cannot have affected the whole study area at the periods of interest in this article.
2. Circulation of hot fluids might also be considered, but it appears difficult to imagine and model fluid circulation on such a large area across rocks with low permeability without any robust driving forces. Numerous mineralized sites have been recognized in the Central Iberian Zone and three periods for mineralization have been reported (Tornos et al. 2000):

331 late-Variscan (295-275 Ma), Mesozoic (150 and 100 Ma) and Alpine (20 Ma), with the late-  
1 332 Variscan deposits representing most of the dated ores. Their occurrence is associated with  
2 333 shear zones (late-Variscan) or faults (younger), mostly as a result of very localised  
3 334 hydrothermal circulation, whose effects are therefore restricted to the fault zone and the  
4 335 nearby host rocks. Given that the ages of the Mesozoic ores (150 and 100 Ma) coincide with  
5 336 the cooling ages deduced from the AFT data (150 and 80 Ma), we infer that Mesozoic  
6 337 remobilisation may be the consequence, not the cause, of the movements that we detected in  
7 338 the Mesozoic (assuming that cooling occurs at the same time as uplift).

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15 339 3. Temperatures higher than surface temperature can be explained by two different models  
16 340 considering granite rock samples or sedimentary Triassic rocks. For the Triassic  
17 341 conglomerates, heating to  $90 \pm 10$  °C must be by burial of Triassic deposits below thick  
18 342 Jurassic ones. For the samples collected in the granitic basement, the temperature may be  
19 343 explained in two different ways: (i) The sampled granites were at the surface in the Triassic  
20 344 and were buried together with the Triassic sediments (Fig. 8A and 8B). Given the  
21 345 geographical distribution of the samples, this would mean that a much wider Jurassic basin  
22 346 should have formed extending much further east than the currently known LB boundary.  
23 347 Wilson et al. (1989) also noted the absence on the eastern basin margin of significant  
24 348 siliciclastic sands in the Lower and Middle Jurassic formations, and suggested that carbonate  
25 349 deposition probably extended far into the interior of Iberia. However, there are no known  
26 350 Jurassic sediments on top of the Variscan basement, and no detrital deposits are known in  
27 351 the LB showing evidence of erosion of all the possible thick Jurassic sediments covering the  
28 352 basement to the east. This assumption may however be true for the central unit samples  
29 353 where the thermal histories are similar to that of the Triassic sample and have relatively old  
30 354 FT ages (SCD, NEL, VTA). Consideration of this assumption on a wider scale has yet to be  
31 355 done. (ii) The sampled granites were at a certain depth (equivalent to ca. 100 °C) at the time  
32 356 of deposition of the Triassic conglomerate, the upper section of the granitic massif being at  
33 357 the surface to feed erosion (Fig. 8A). During the Jurassic, the Triassic conglomerate was  
34 358 buried below thick Jurassic sediments, and along the master bounding fault at a similar depth  
35 359 to that of the basement samples were (Fig. 8B). Later, at ca. 150 Ma, footwall uplift occurred  
36 360 differentially along the master bounding faults (Porto-Tomar-Badajoz, Chaves and Vilariça  
37 361 faults) to produce a major rift shoulder uplift that exhumed the buried rocks that we sampled  
38 362 in both Triassic conglomerates and the granitic basement (Fig. 8C). Consequent erosion  
39 363 brought the LB with thick and coarse siliciclastic sediments. This second scenario is

364 consistent with geological evidence and the fact that AFT data which shows a similar story  
1 365 for granites in the basement and in the Triassic conglomerates. Our data includes only one  
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3 366 Triassic conglomerate, but some similar data has been measured in the LB: AFT ages of  
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5 367 169±8, 204±9, 132±16 and 147±10 Ma have been measured in four sandstones from  
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7 368 different stratigraphic levels, respectively Upper Carboniferous, Triassic, Upper Jurassic and  
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9 369 Lower Cretaceous (Pereira et al., 1998). These ages also show that Upper Jurassic and Lower  
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11 370 Cretaceous sediments were not significantly affected by annealing in the LB.

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13 371 We suggest that differential burial of the Triassic (subsiding hanging wall) and the  
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15 372 Variscan basement (stable footwall with most of the granite samples at a certain depth) and the  
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17 373 later simultaneous uplift of both is the most plausible explanation for the high temperatures  
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19 374 recorded by FT data.

## 20 21 375 22 23 376 *5.2. Age and mechanisms of Mesozoic cooling*

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25 377 Two phases for the beginning of cooling have been defined by the thermal modelling of  
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27 378 different samples: ~160-150 Ma from the western and central units and ~120-110 Ma from the  
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29 379 eastern unit. The offset between these two phases can be interpreted in two different ways. Two  
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31 380 uplift episodes separated by ~40 Ma, or a single, slower, uplift episode. In the latter case, the  
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33 381 age offset would correspond to the duration required for the eastern unit to cross the isotherms,  
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35 382 with samples from this unit being originally much deeper in the crust than samples from the  
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37 383 western and central units. Coincidentally, these ages correspond to the limits of the interval  
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39 384 ascribed in the literature to rift-to-drift – 160 to 110 Ma. FT ages are younger towards the north  
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41 385 and this is consistent with the proposed diachronism for rifting in western Iberia. Jagoutz et al.  
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43 386 (2007) argued that ca. 110 Ma is the age of the rift-to-drift stage, i.e. it is the age at which new  
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45 387 oceanic crust started to form. Our AFT data also indicates cooling at this age, and therefore we  
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47 388 agree with Jagoutz et al.'s (2007) interpretation that ca. 110 Ma marks the rift-to-drift stage.  
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49 389 Timing of the uplift in southern Portugal is different from northern Portugal as the probable age  
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51 390 is Late Cretaceous to early Palaeocene. Fernandez et al. (2013) showed that peak temperatures  
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53 391 were attained in the Mesozoic rocks during the latest Cretaceous-early Palaeocene times  
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55 392 (Fernandez et al., 2013). Their modelled palaeotemperatures suggest also that the Jurassic  
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57 393 sedimentary rocks entered the oil-window at the beginning of the Cretaceous, with peak oil  
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59 394 generation in the Late Cretaceous time.

396 5.3. Consistency between new and previous data/interpretations

397 Three to four rift phases have been proposed for the LB (see Fig. 2; e.g., Wilson et al.,  
398 1989; Rasmussen et al., 1998; Alves et al., 2009). We do not discard the hypothesis that minor  
399 rift pulses may have existed, like Rifts 2 and 3 of Rasmussen et al. (1998) and Alves et al.  
400 (2002, 2006, 2009), or extensional event two of Wilson et al. (1989). However, from the main  
401 stratigraphic trends in the geological record and their geodynamic meaning, we do not see the  
402 need for three or four rift phases. For instance, Rifts 1 and 2 are separated by a gap in which  
403 the thick Hettangian evaporites were deposited (e.g., Rasmussen et al., 1998; Alves et al., 2002,  
404 2006, 2009; Alves and Cunha, 2018). Wilson et al. (1989) also did not find arguments in favour  
405 of two rift stages for the period Triassic-Callovia. Deposition of marly limestones and  
406 limestones marks unrestricted opening to sea water, thus again indicating continued subsidence  
407 until the late Upper Jurassic (ca. 160 Ma). This continued subsidence is consistent with the  
408 modelled heating event between 190 and 160 Ma, which was suddenly interrupted ca. 150-145  
409 Ma ago, as attested by the major Jurassic/Cretaceous unconformity with deposition of thick  
410 siliciclastic sediments derived from the uplifted rift shoulder.

411 Based on large unconformities, we suggest that there were only two continental rift stages,  
412 one (Rift 1) from the Late Triassic to the latest Jurassic, and the other (Rift 2) from the  
413 Jurassic/Cretaceous boundary till the first formation of the basaltic oceanic crust. This rift-to-  
414 drift stage marks the shift from continental (Rift 2) to oceanic rifting, which can be considered  
415 Rift 3 lasting until present-day. We note that, despite disagreements regarding rift stages, there  
416 is a major coincidence between our AFT cooling age of ca. 150-145 Ma age (Rift 1 to Rift 2  
417 boundary) and the age of the base of Breakup Sequence A (ca. 150 Ma) of Alves and Cunha  
418 (2018), or the age of shifting from pre-tectonic to syn-tectonic sedimentation (ca. 145 Ma) of  
419 Sutra et al. (2013).

420 Boillot et al. (1980) showed that the lithosphere underwent drastic thinning from the  
421 Tithonian to the Barremian, which finally led to the unroofing of the subcontinental mantle in  
422 the distal margin to form the COT. Boillot et al. (1988) suggested that tectonic unroofing and  
423 serpentinization of mantle rocks occurred just before seafloor spreading started at 114 Ma ago  
424 in this part of the North Atlantic, which is also consistent with the 110 Ma age proposed for the  
425 rift-to-drift stage. In order for the lithosphere to undergo major thinning, its elastic core must  
426 have vanished, thus allowing for whole lithosphere failure (WLF). Therefore, WLF must  
427 precede hyper-extension and mantle exhumation, which means that there is no inconsistency  
428 between our new data (WLF age) and previous data and ages. Rift 2 was not as simple as Rift



1 (continued subsidence) and led to a complexity that characterizes the rift in western Iberia, which has attracted so much investment in research over several decades. When the elastic core of the continental lithosphere broke (WLF stage between Rifts 1 and 2), three main geodynamic reconfigurations occurred during Rift 2: (1) the rift shoulder underwent a major uplift, (2) the continental viscous “lithosphere” was hyper-extended, which finally led to (3) unroofing and mantle exhumation.

Jagoutz et al. (2007) concluded that the rift-to-drift stage is not a well-defined event, but rather a transitional process occurring over a period of time lasting from ca. 128 Ma (oldest age of intrusive gabbro) to ca. 113 Ma in the rift in western Iberia. These ages are consistent with the AFT data reported here, which indicates a rift-to-drift stage at around 110 Ma. Given that 128 Ma is the minimum age of mantle exhumation, then hyper-extension and mantle exhumation must have taken place before 128 Ma. In the Iberia Abyssal Plain, which roughly corresponds to the area of interest in this study, mantle exhumation likely started at ca. 136 Ma (Valanginian) (e.g., Manatschal et al. 2001; Sutra et al. 2013). This is consistent with the age of 150-145 Ma for WLF. Wilson et al. (2001) suggested that the maximum age of the deformation which led to continental breakup in the Newfoundland-Iberia Rift is given by tilted Tithonian sediments (ca. 152-145 Ma) deposited over thinned continental crust drilled at ODP Sites 901, 1065 and 1069. This is also consistent with the age we propose for breakup. More recently, Alves et al. (2009) and Alves and Cunha (2018), based on the concept of breakup unconformity, suggested early Berriasian (ca. 145-140 Ma) for the age of continental breakup, which again is consistent with our new data for WLF.

The new AFT data shows younging of the cooling ages to the north in the western and central units, in agreement with the diachronous evolution of the rift proposed in previous studies (e.g. Whitmarsh and Miles, 1995; Tucholke et al., 2007; Alves et al., 2009). Thermochronology data also show that the margin remained relatively stable after the Central-North Atlantic opening and that minor Alpine deformation is recorded during Tertiary (erosion <1 km), in contrast to central Iberia where vertical movements are significant.

#### *5.4. Regional implications*

From the new AFT data and the geological context, we infer the following evolution of the eastern margin of the Newfoundland-Iberia Rift (carton in Fig. 9):

(1) Rift 1 – first stage lasting from the Triassic until ca. 150 Ma, which corresponds to a general

461 subsidence of the margin that is responsible for the initial heating stage in AFT data (Fig.  
1 462 9A). Vertical movements were accommodated along lithospheric scale shear zones  
2 463 inherited from the Variscan orogeny.

3 464 (2) Cooling at ca. 150 Ma – this stage is characterized by a general uplift of the margin and  
4 465 associated major erosional angular unconformity (Fig. 9B). The Western and Central units  
5 466 in the Variscan basement and eastern margin of the Lusitanian Basin were eroded as  
6 467 indicated by cooling of the samples. The critical change in sediment nature, from carbonate  
7 468 dominated sediments to coarse siliciclastic rocks, also attests to the uplift of the margin.  
8 469 This uplift seems to have been generalised to the entire Lusitanian Basin, as attested by the  
9 470 thick siliciclastic sediments in the Alentejo Basin (Alves et al., 2002), which diachronously  
10 471 progressed northwards so covering the entire Iberian margin both on and offshore.

11 472 (3) Rift 2 – this stage is bracketed between ca. 145 and 110 Ma (Fig. 9C), and mostly  
12 473 corresponds to the period of hyper-extension and of the margin and consequent mantle  
13 474 upwelling. In the Iberia Abyssal Plain, which roughly corresponds to the area of interest in  
14 475 this study, mantle exhumation likely started at ca. 136 Ma (Valanginian) (e.g., Manatschal  
15 476 et al. 2001; Sutra et al. 2013). The Eastern Unit in the Variscan basement was effectively  
16 477 eroded and the whole Variscan basement knew vertical movements, as inferred from the  
17 478 AFT data.

18 479 (4) Rift 3 – this stage is bracketed between ca. 110 Ma and present-day (Fig. 9C). Rift 2 of the  
19 480 continental rifting process came to an end with the onset of oceanic spreading (oceanic  
20 481 rifting), which occurred in the late Early Cretaceous (ca. 110 Ma) according to the new  
21 482 AFT data. The general trend of the sedimentation was upward finning during most of Rift  
22 483 2, but the Aptian-Albian was again coarse silicic sandstones (Rasmussen et al., 1998;  
23 484 Wilson et al., 2001), from which we infer relatively high topography of the basement east  
24 485 of the Porto-Tomar Fault around 120-110 Ma that could be responsible for the AFT cooling  
25 486 age at ca. 110 Ma.

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## 27 488

## 28 489 **6. Conclusions**

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30 491 Based on geological data and the new AFT results, we infer the following geodynamic  
31 492 evolution of the Newfoundland-Iberia Rift in its eastern margin (western Iberia) (Fig. 9): (1)

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493 Rift 1 (>200 to ca. 150 Ma) – slow and continued subsidence from Upper Triassic to uppermost  
1 494 Jurassic (Kimmeridgian); (2) Breakup (150-145 Ma) – whole lithosphere failure by vanishing  
2 of its elastic core, with consequent major rift shoulder uplift and formation of a basin-wide  
3 495 erosional unconformity followed by deposition of thick siliciclastic rocks in the subsequent  
4 496 rifting stage; (3) Rift 2 (ca. 145 to 110 Ma) – without elastic resistance, the viscous  
5 497 “lithosphere” could be hyper-extended, which allowed for unroofing and mantle exhumation  
6 498 after ca. 136 Ma; (4) Rift-to-drift (ca. 110 Ma) – rise of the asthenosphere to shallow depth and  
7 499 first formation of basaltic oceanic crust, thus marking the transition from continental to oceanic  
8 500 rifting; (5) Rift 3 (ca. 110 Ma to present-day) – oceanic spreading with development of the Mid-  
9 501 Atlantic Ridge.  
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11 503 Contrary to previous work, we distinguish between breakup (whole lithosphere failure)  
12 504 and rift-to-drift (onset of oceanic spreading) and attribute an age to each of the stages based on  
13 505 the new AFT data: breakup at ca. 150 Ma and rift-to-drift at ca. 110 Ma, with an intervening  
14 506 stage of hyper-extension.  
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738 **Tables**

739

740 Table 1. Sample details (locality, lithology, U-Pb age, coordinates in WGS84 projection system,  
741 and altitude) and apatite fission-track results.

742 n - number of apatite crystals counted; s, i and d subscripts denote spontaneous, induced and  
743 dosimeter;  $\rho$  – track density ( $\times 10^5$  tracks/cm<sup>2</sup>); N - number of tracks counted;  $P(\chi^2)$  - probability  
744 of obtaining Chi-square value ( $\chi^2$ ) for n degrees of freedom (where n = number of crystals - 1);  
745 Age  $\pm 1\sigma$  - central age  $\pm 1$  standard error (Galbraith and Laslett, 1993); MTL - mean track  
746 length ( $\mu\text{m}$ ); SD - standard deviation of track length distribution ( $\mu\text{m}$ ); N(L) - number of  
747 horizontal confined tracks measured ; Dpar - average etch pit diameter parallel to c. Ages were  
748 calculated using the zeta calibration method (Hurford and Green, 1983), glass dosimeters CN-  
749 5, and a zeta value of  $359 \pm 8$  (JB). U/Pb ages inferred to represent crystallization ages are from  
750 (1) Neiva et al. (2009); (2) Valle Aguado et al. (2005); (3) Neiva et al. (2012); (4) Dias et al.  
751 (1998); (5) Valle Aguado et al. (2017); (6) Antunes et al. (2008); (7) Pin et al. (2008); (8) Solá  
752 et al. (2009); (9) Lima et al. (2012).

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754

755 **Figures**

756

757 Figure 1. Simplified geological map with squares marking sample location, and showing sample  
758 reference, AFT ages (top, in Ma) and track length (bottom, in  $\mu\text{m}$ ). Color of the squares (blue,  
759 green, orange and yellow) corresponds respectively to the western unit, central unit, eastern unit  
760 and samples from the south of the PTBF. CF and VF – Chaves and Vilariça faults, respectively;  
761 PTBF – Porto-Tomar-Badajoz Fault. MPF – Messejana-Placencia Fault with mafic dyke  
762 intruded in most of its extension. Cyan lines – reverse faults bounding the Serra da Estrela pop-  
763 up. Inset top left – lithostratigraphy, NLB, CLB and SLB – northern, central and southern LB,  
764 respectively. AB – Alentejo Basin. GB – Galicia Bank. IAP and TAP – Iberian and Tagus  
765 Abyssal Plains, respectively. Inset bottom left – Western Iberia-Newfoundland conjugate  
766 margins (WICM and NFCM, respectively). Black, red and white dashed lines mark the  
767 boundaries between proximal, thinned, exhumed and oceanic domains (after Nirrengarten et  
768 al., 2017). Image downloaded from EMODnet (<https://portal.emodnet-bathymetry.eu/>).

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2 770 Figure 2. Synthetic lithostratigraphic logs of the onshore and offshore Lusitanian Basin between  
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4 771 Figueira da Foz (north) and Nazaré (south) (see Fig. 3 for location). Onshore data are from  
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6 772 Rocha et al. (1981), Wilson et al. (1989) and Azerêdo (2007). Offshore data from figure 14 in  
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8 773 Alves et al. (2009). R1 to R4 in black are from Alves et al. (2009).

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12 775 Figure 3. Depth to basement map obtained in the Lusitanian Basin by modelling of seismic  
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14 776 velocities and magnetic and gravity anomalies (SEEBASE, FrogTech, 2012). Note that in this  
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16 777 model the deepest depocenter in the onshore LB is around -8,000 m.

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21 779 Figure 4. FT results: track length vs FT age by geographic region. Colours correspond to the  
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23 780 location of the samples: blue for western unit, green for central unit, and orange for eastern unit.  
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25 781 Sample J (in black) is located in the south of the studied area (Ossa Morena Zone).

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29 783 Figure 5. FT modelling of samples from the central unit. Two boxes at the beginning and at the  
30  
31 784 end of the history have been set; additional boxes depend on the samples. A – sample SCD with  
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33 785 constraints only at the beginning and the end ; B – sample NEL with a box during Early  
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35 786 Cretaceous justified by the presence of Lower Cretaceous sedimentary rocks on top of the  
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37 787 basement east of the PTBF; C – sample NEL with an additional box during Triassic times,  
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39 788 consistent with the presence of basement of this domain at the surface during that period, as  
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41 789 attested by the thermal history of sample SCD; D – sample VTA with modelling conditions  
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43 790 similar to sample NEL; E – sample AN; F – sample COI3, for which residence at the surface  
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45 791 during Triassic times is necessary as the stratigraphic age of the sample is Triassic. Colors of  
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47 792 single thermal history correspond to the relative probability of the posterior model, red color  
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49 793 being the higher probability (Gallagher, 2012). Color scale is presented on the right of the  
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51 794 models.

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54 796 Figure 6. FT modelling of samples from the western unit. No additional constraint has been set  
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56 797 as no Mesozoic sedimentary cover is preserved. A: sample BAR, B: sample GRJ, C: sample  
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58 798 ALH and D: sample SVG.

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800 Figure 7. FT modelling for samples from the eastern unit. A – sample CST; B – sample MTG1;  
801 C – sample SRT; and D – sample T, for which the model assumes the presence of the sample  
802 close to the surface during Triassic times, and shows lower temperature than for the other  
803 samples during the Jurassic.

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805 Figure 8. Interpreted evolution of the LB's eastern margin since the Triassic, supported by the  
806 geological and AFT data reported here. A – late-Variscan orogenic collapse to Triassic early  
807 rifting of Pangaea. B – main subsidence during the Jurassic, including the Hettangian  
808 (evaporites up to 3,000 m thick) and overlying rocks up to the Lower Kimmeridgian  
809 (Carbonated rocks ca. 3,000 m thick). C – Basin uplift in the Upper Kimmeridgian and  
810 Tithonian. D – major rift shoulder uplift, with injection of alkaline dykes (thick red line) into  
811 Jurassic sediments following lithospheric scale discontinuities (Grange et al., 2008).

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813 Figure 9: Geodynamic model suggested from data for the evolution of the western Iberia  
814 margin. MAR: middle Atlantic ridge.

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817 Supplementary materials

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819 Figure S1: Sketch with the yield strength envelope (YSE) for continental lithosphere. Note that,  
820 prior to whole lithosphere failure, two elastic cores exist that can vanish due to increase in time,  
821 and/or heat and/or stress. For rifting, the YSE must be flipped around the ordinate axis.

822 Figure S2a-b-c: representation of fission-track ages in radial plot distributions.

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824 Figure S3: histograms of horizontal confined track-length distribution for all samples.

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Code	Lithotype	U-Pb age (Ma)	Latitude (°)	Longitude (°)	Altitude (m)	n	r <sub>s</sub>	N <sub>s</sub>	r <sub>i</sub>	N <sub>i</sub>	r <sub>d</sub>	N <sub>d</sub>	P(c <sup>2</sup> )	RE	U (ppm)	FT age (Ma)	N	L (μm)	SD (μm)	Dpar (μm)
<b>Western Unit</b>																				
BAR	biotite granodiorite	319±4 (4)	41.12307	-7.88465	278	23	0.565	971	0.946	1626	7.024	6686	91.1	0.1	16	<b>75±4</b>	87	<b>12.0±0.1</b>	1.3	1.4±0.1
GRJ	monzonite		41.10244	-8.00113	118	22	1.27	960	1.753	1325	6.717	6686	79.1	0.7	32	<b>87±4</b>	58	<b>12.6±0.2</b>	1.8	1.4±0.1
CIF	two mica granite		41.07861	-8.1011	357	21	2.332	1770	2.427	1842	6.901	6686	60.7	0.8	43	<b>118±5</b>	8	<b>12.6±0.5</b>	1.5	1.5±0.1
ALH	porphyroid monzonite		40.97636	-8.0209	1105	20	2.452	1910	2.452	1936	7.086	6686	87.6	0,0	43	<b>124±5</b>	100	<b>12.6±0.1</b>	1.4	1.4±0.1
SVG	two mica granite	308±1 (2)	40.67345	-8.33757	440	22	1.863	1099	1.78	1050	6.348	6686	97.7	0,0	34	<b>118±6</b>	22	<b>12.9±0.2</b>	1,0	1.3±0.1
<b>Central Unit</b>																				
AN	biotite granodiorite	319±4 (4)	41.08813	-7.7996	530	20	1.459	1005	1.158	798	6.174	6213	11.7	12.1	23	<b>137±8</b>	100	<b>12.5±0.1</b>	1.2	1.7±0.1
VTA	two mica granite	308±1 (2)	40.69365	-8.08034	612	20	0.888	1089	0.635	779	6.286	6213	89.7	0.2	12	<b>156±8</b>	100	<b>12.2±0.1</b>	1.4	1.4±0.2
NEL	porphyroid monzonite	297±3 (3)	40.53177	-7.86412	405	20	3.239	2212	2.184	1492	6.532	6686	86.1	0,0	41	<b>172±7</b>	100	<b>12.7±0.2</b>	1.6	1.5±0.1
SCD	monzonite	198±1 (5)	40.39512	-8.099	206	18	5.592	2220	3.358	1333	6.471	6686	49,0	2.3	63	<b>191±8</b>	100	<b>13.4±0.1</b>	1,0	1.5±0.1
<b>Eastern Unit</b>																				
AC	monzonite		40.76379	-6.85296	711	16	2.882	1568	2.079	1131	6.14	6213	7.3	9.9	41	<b>151±8</b>	100	<b>12.8±0.2</b>	1.5	1.8±0.1

CEL	porphyroid monzonite	305±4 (2)	40.62911	-7.41561	448	19	0.376	534	0.418	594	6.963	6686	20.3	15.4	7	<b>110±9</b>	16	<b>13.3±0.3</b>	1.0	1.4±0.1
MTG1	monzonite	301±3 (1)	40.41253	-7.55279	1224	23	0.684	1031	0.808	1218	6.655	6686	69.1	1.1	15	<b>100±5</b>	85	<b>12.6±0.2</b>	1.4	1.4±0.1
SRT	porphyroid monzonite		40.31532	-7.28675	507	21	2.743	1684	3.349	2056	6.409	6686	96,0	0,0	64	<b>94±4</b>	101	<b>12.3±0.1</b>	1.3	1.4±0.1
CST	two mica granite	310±1 (6)	40.07818	-7.49373	590	17	1.997	1206	3.104	1875	6.778	6686	31.7	6,0	56	<b>78±4</b>	92	<b>12.6±0.2</b>	1.7	1.7±0.1
T	porphyroid monzonite	305±6 (8)	39.41024	-7.62334	331	19	6.447	2824	4.406	1930	6.23	6213	1.7	10.9	86	<b>162±8</b>	101	<b>12.2±0.2</b>	1.6	1.8±0.1

#### Lusitanian Basin

COI3	granite in Triassic conglomerate		40.244	-8.39941	149	20	2.256	1135	1.748	879	6.84	6686	79.1	0.1	31	<b>157±8</b>	100	<b>12.0±0.1</b>	1.3	1.4±0.1
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#### Ossa-Morena Zone

J	Porphyrid 2-micas granite	324±1 (9)	38.84073	-7.89542	210	18	0.543	415	0.619	473	6.185	6213	60.8	3.1	13	<b>97±7</b>	10	<b>13.2±0.4</b>	1.2	1.6±0.1
TOR	gabbro	352±2 (7)	38.29957	-8.22976	61	15	0.248	75	0.315	238	6.286	6686	90.5	0	6	<b>88±16</b>	-	-	-	1.4±0.1



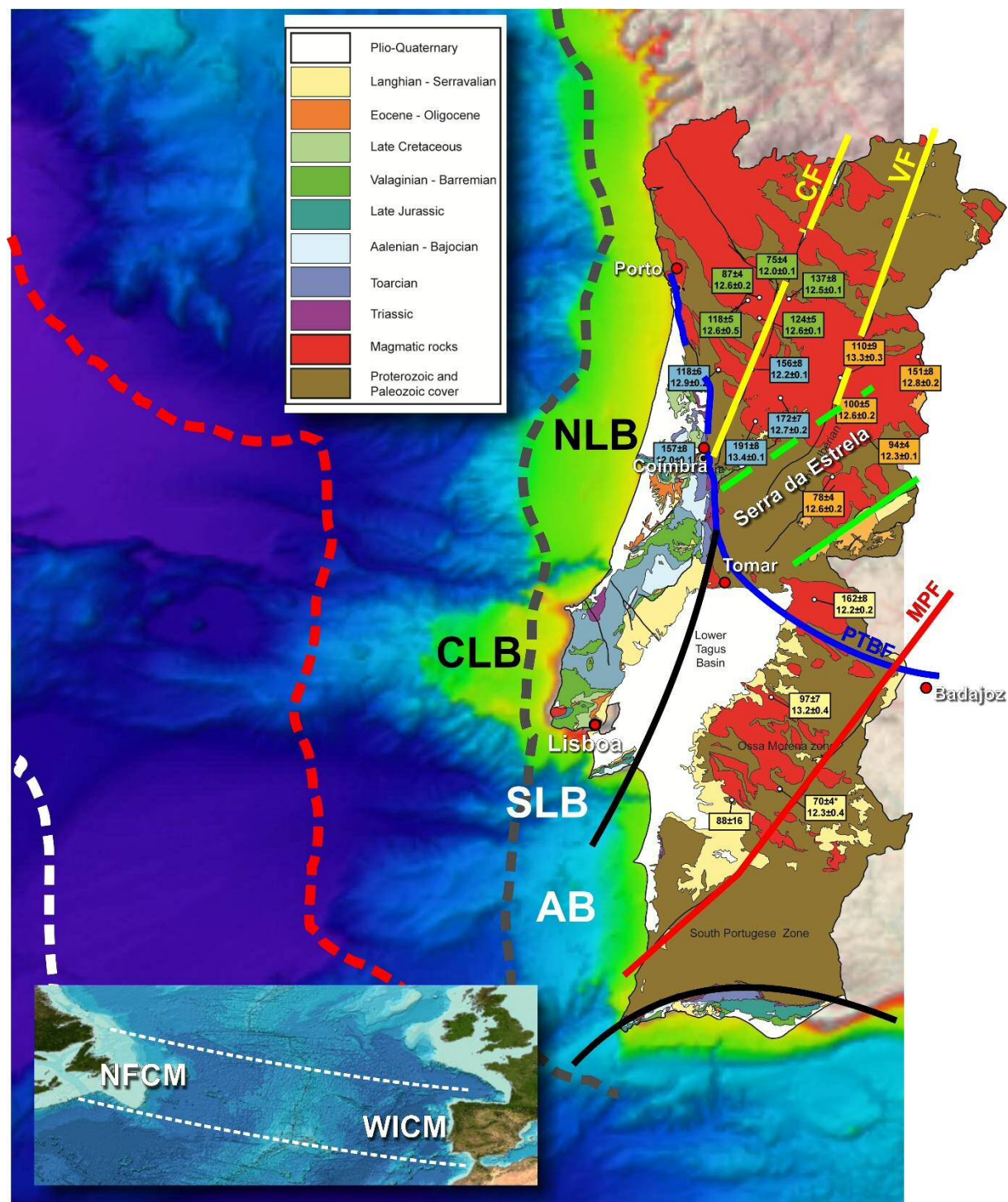


Figure 1. Simplified geological map with squares marking sample location, and showing sample reference, AFT ages (top, in Ma) and track length (bottom, in  $\mu\text{m}$ ). Color of the squares (blue, green, orange and yellow) corresponds respectively to the western unit, central unit, eastern unit and samples from the south of the PTBF. CF and VF – Chaves and Vilarica faults, respectively; PTBF – Porto-Tomar-Badajoz Fault. MPF – Messejana-Placencia Fault with mafic dyke intruded in most of its extension. Cyan lines – reverse faults bounding the Serra da

Estrela pop-up. Inset top left – lithostratigraphy, NLB, CLB and SLB – northern, central and southern LB, respectively. AB – Alentejo Basin. GB – Galicia Bank. IAP and TAP – Iberian and Tagus Abyssal Plains, respectively. Inset bottom left – Western Iberia-Newfoundland conjugate margins (WICM and NFCM, respectively). Black, red and white dashed lines mark the boundaries between proximal, thinned, exhumed and oceanic domains (after Nirrengarten et al., 2017). Image downloaded from EMODnet (<https://portal.emodnet-bathymetry.eu/>).

Figure 2

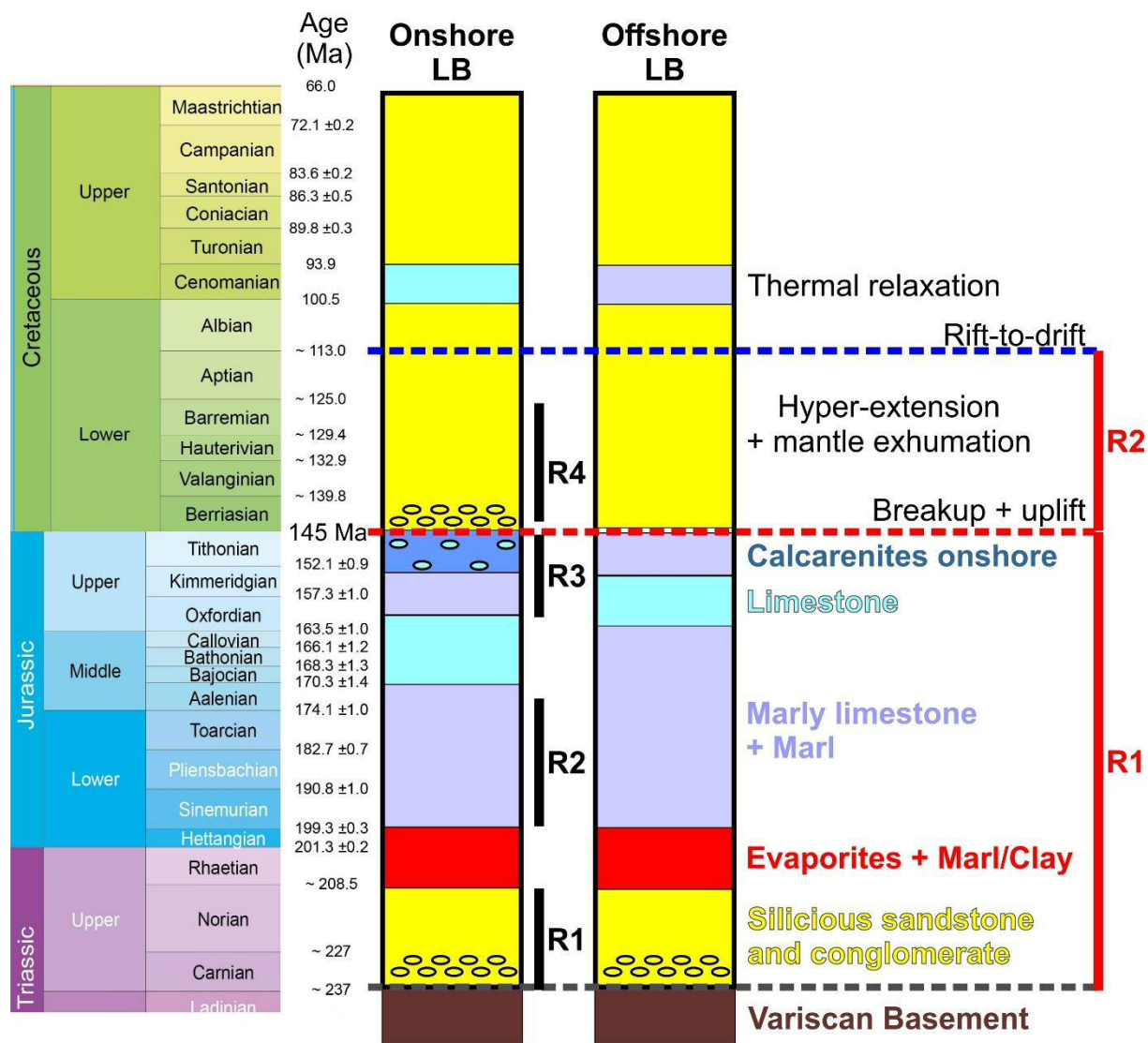
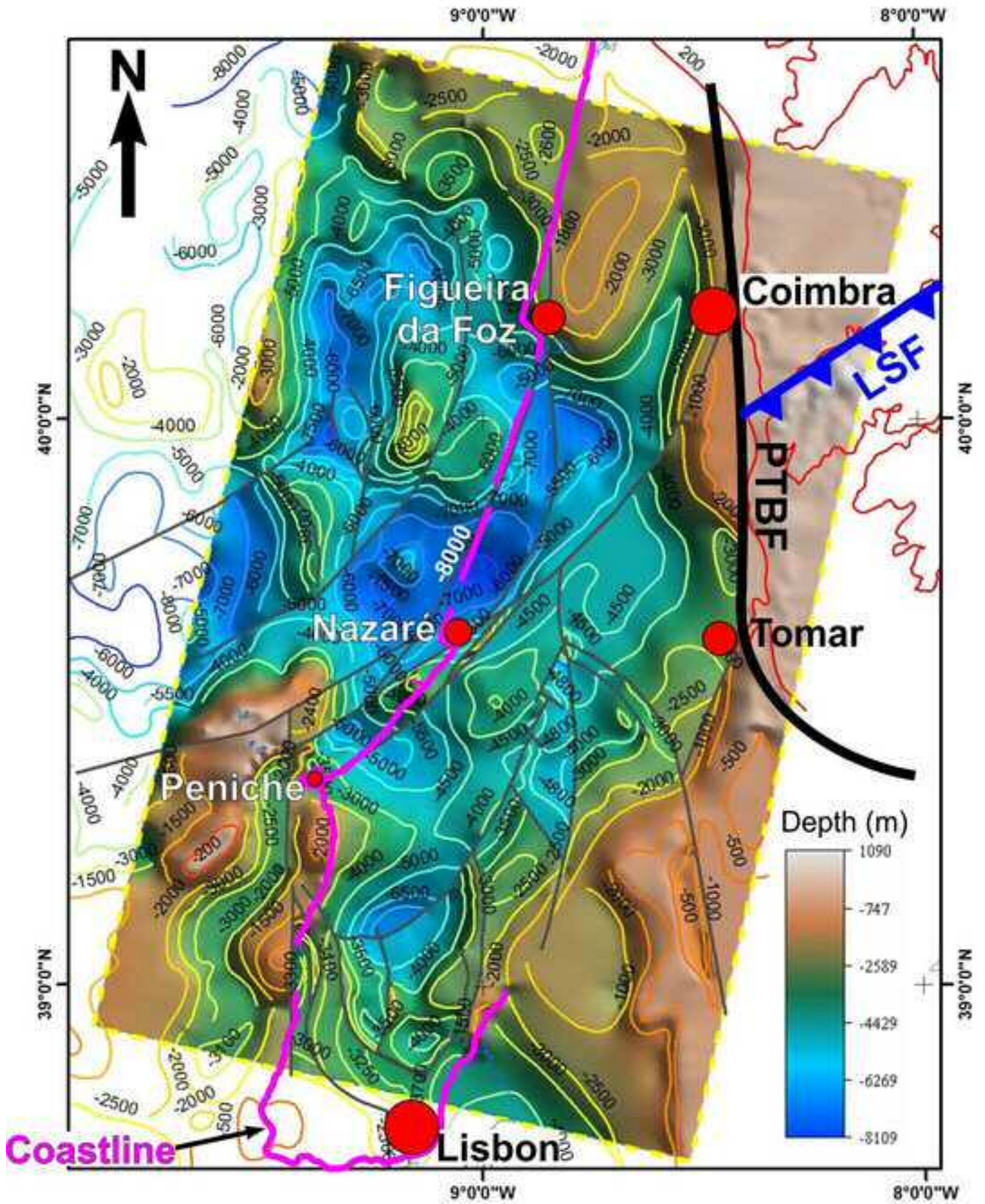


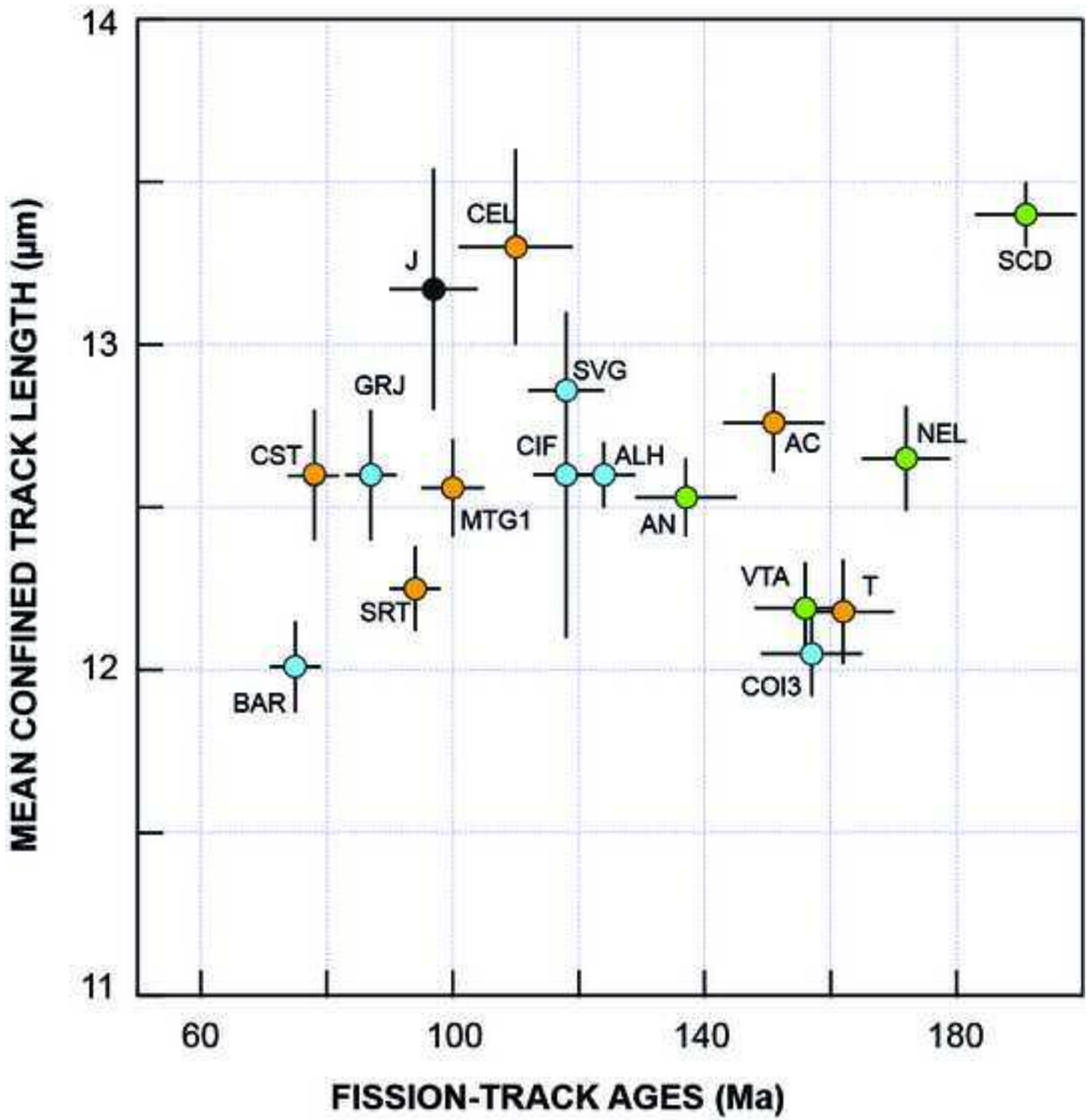
Figure 2. Synthetic lithostratigraphic logs of the onshore and offshore Lusitanian Basin between Figueira da Foz (north) and Nazaré (south) (see Fig. 3 for location). Onshore data are from Rocha et al. (1981), Wilson et al. (1989) and Azerêdo (2007). Offshore data from figure 14 in Alves et al. (2009). R1 to R4 in black are from Alves et al. (2009).



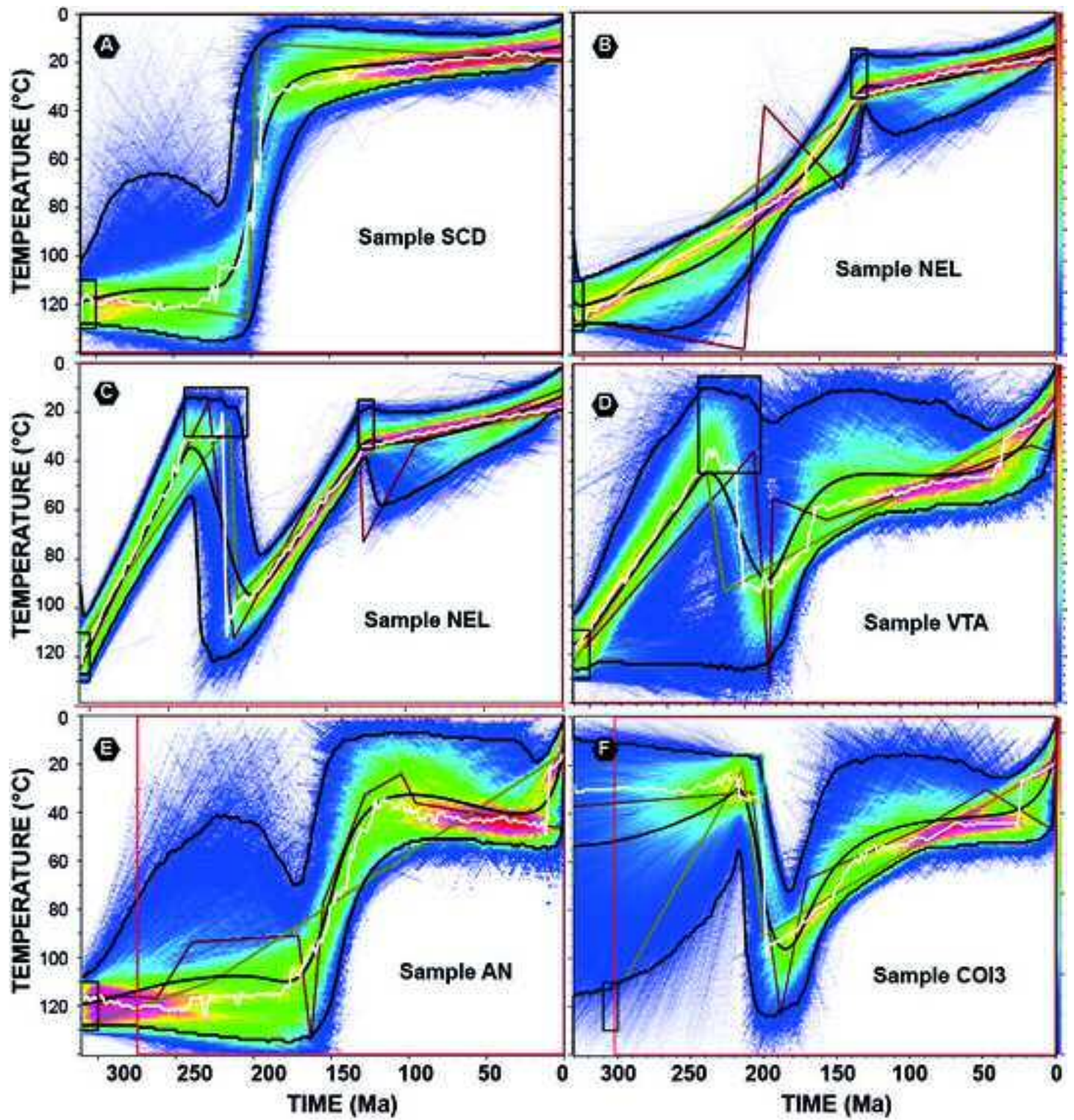
Figure 3

Click here to access/download;Figure (with caption below and on the same page);Figure 3 Lusitanian Basin.jpg

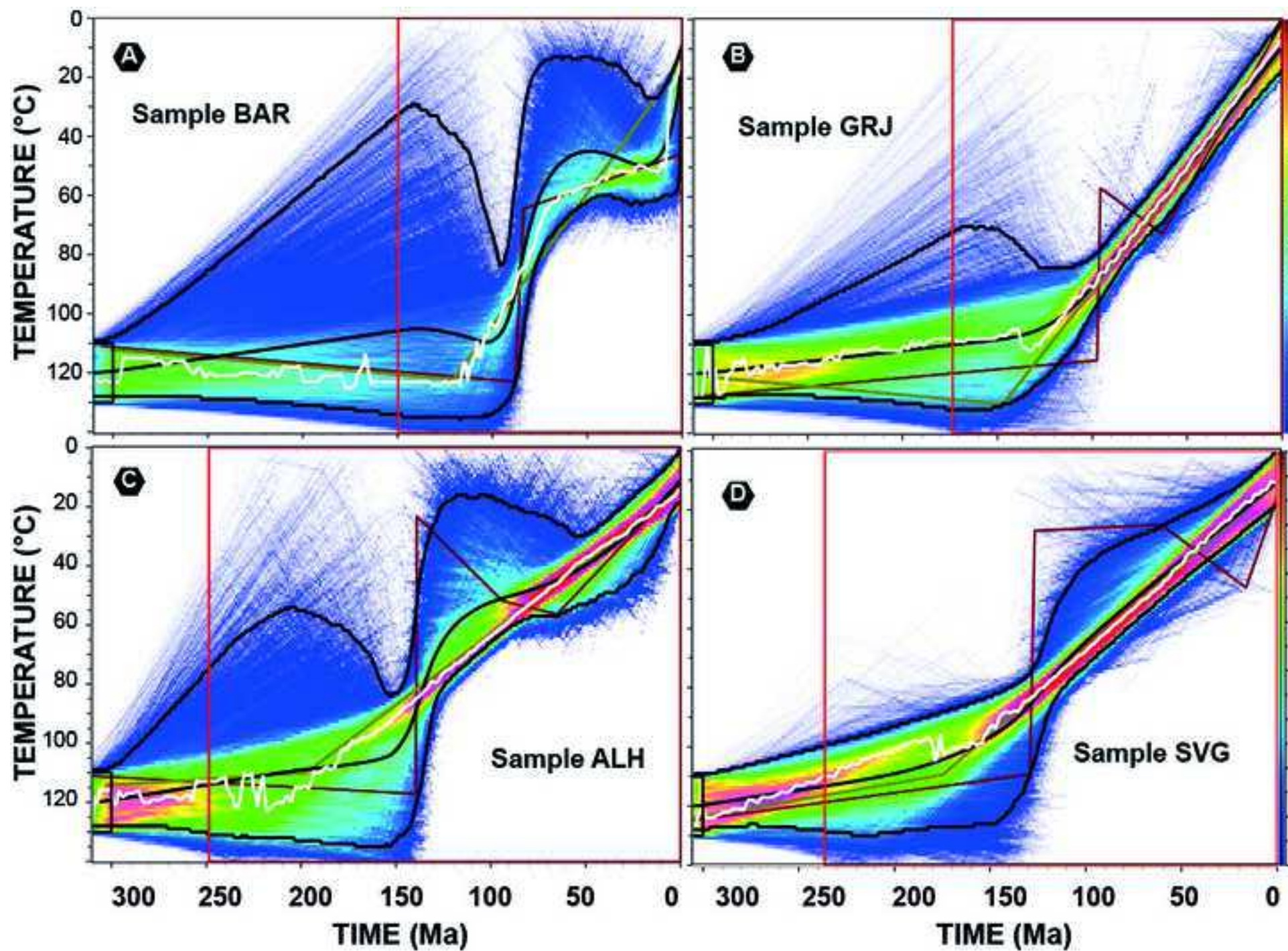




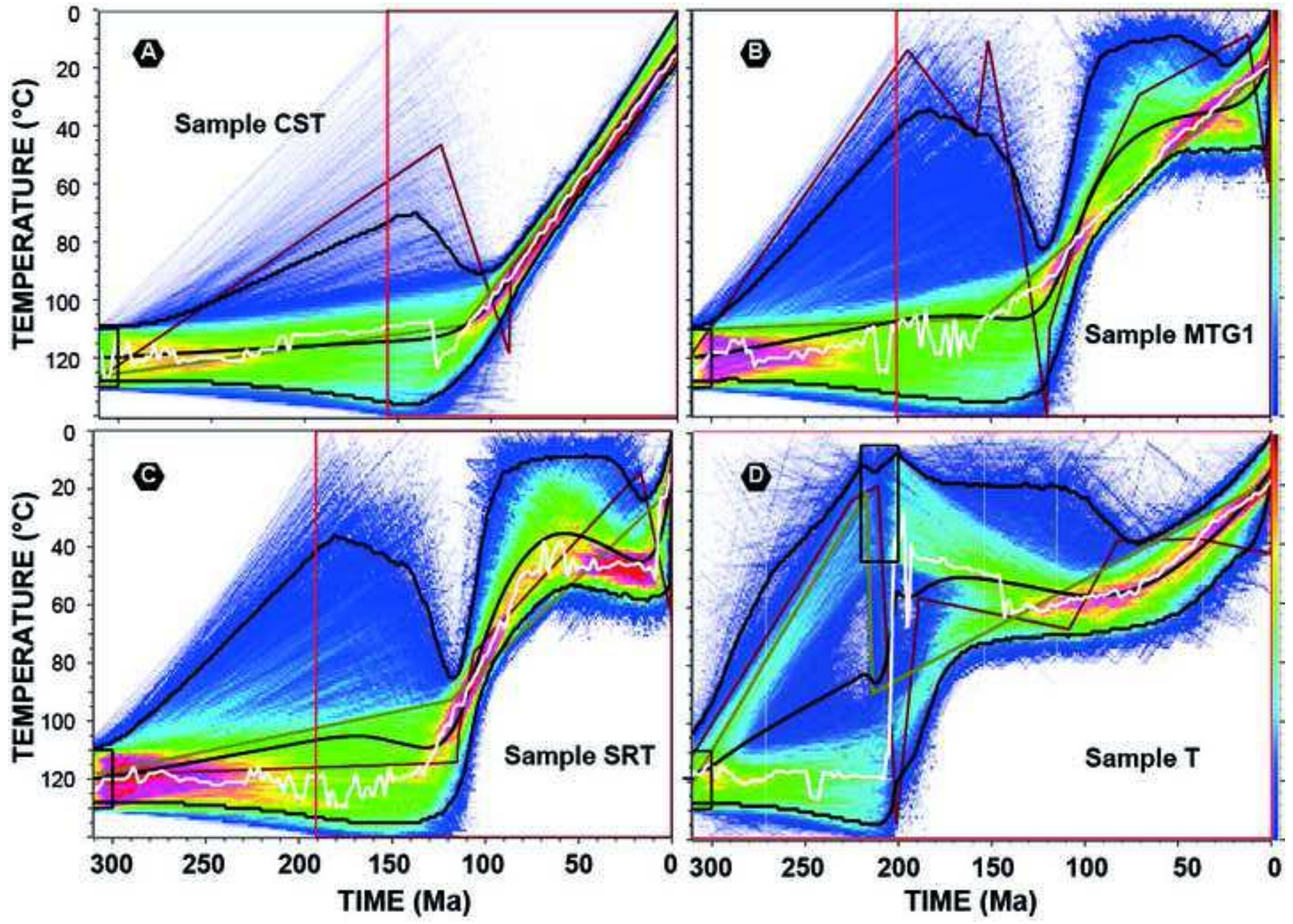














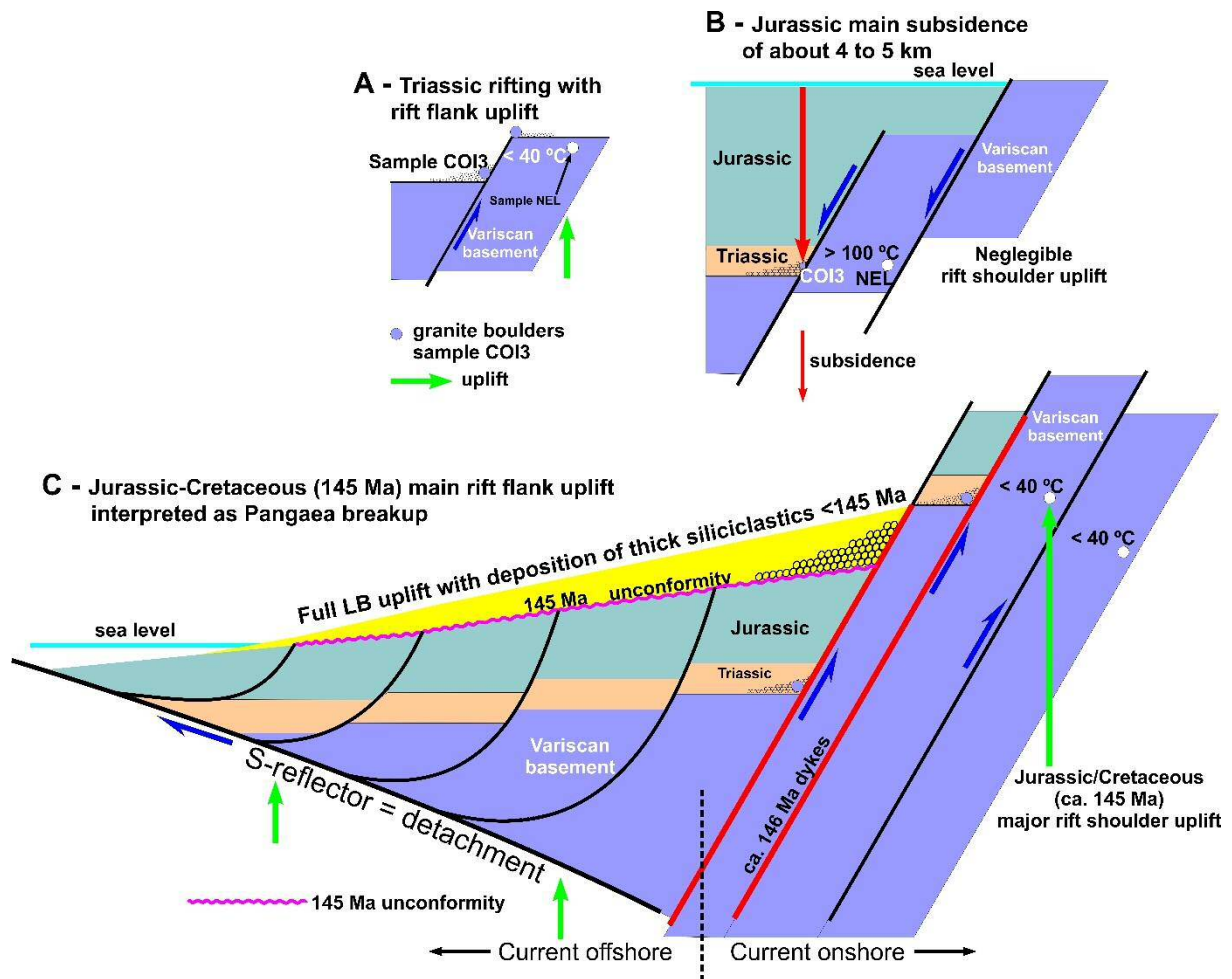


Figure 8. Interpreted evolution of the LB's eastern margin since the Triassic, supported by the geological and AFT data reported here. A – late-Variscan orogenic collapse to Triassic early rifting of Pangaea. B – main subsidence during the Jurassic, including the Hettangian (evaporites up to 3,000 m thick) and overlying rocks up to the Lower Kimmeridgian (Carbonated rocks ca. 3,000 m thick). C – Basin uplift in the Upper Kimmeridgian and Tithonian. D – major rift shoulder uplift, with injection of alkaline dykes (thick red line) into Jurassic sediments following lithospheric scale discontinuities (Grange et al., 2008).

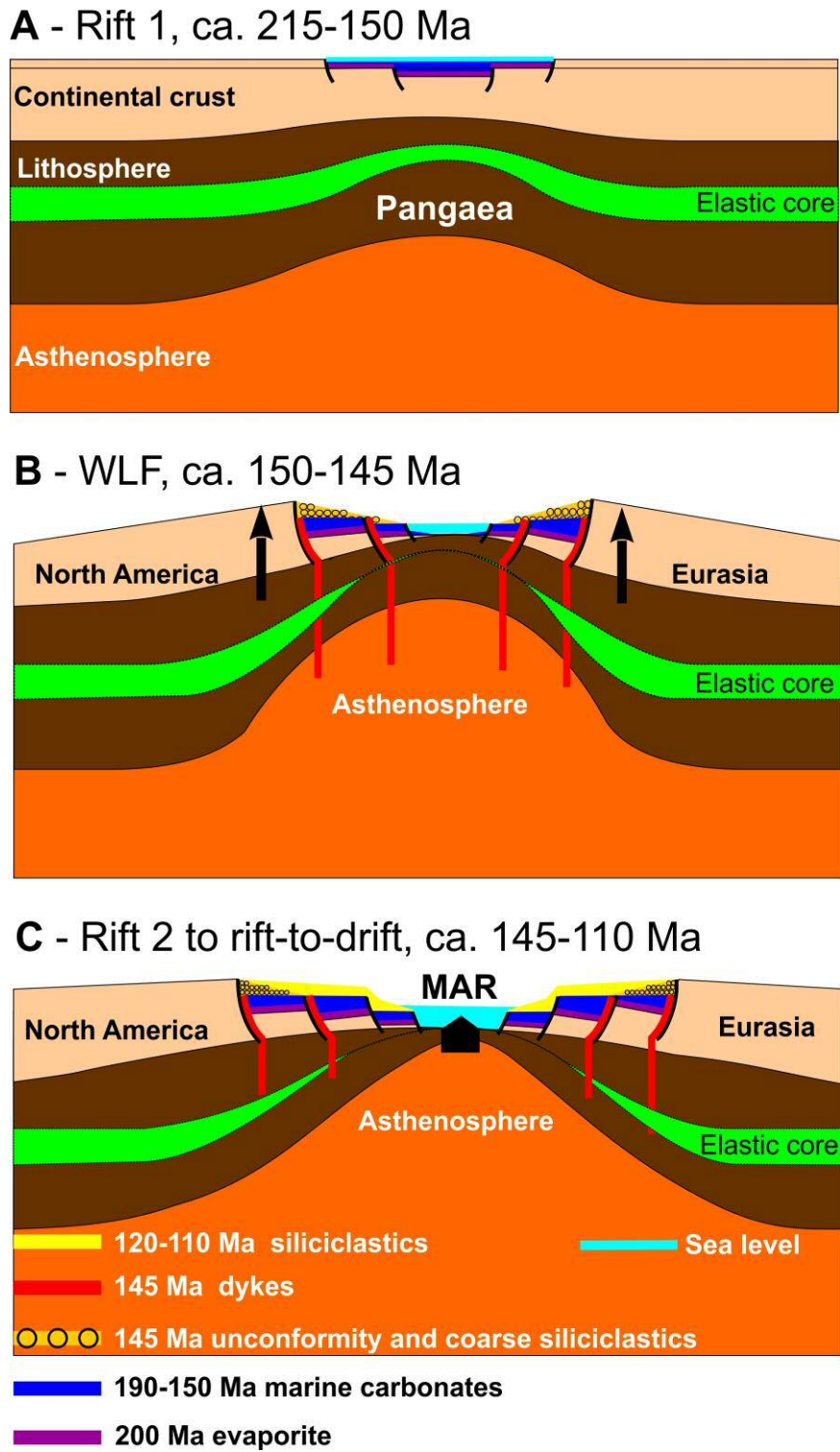
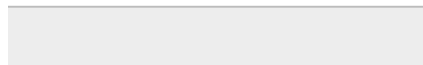
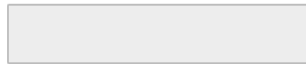


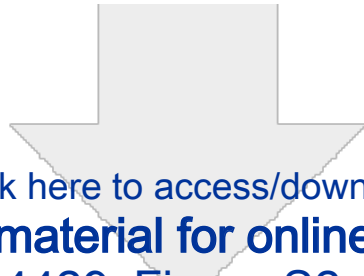
Figure 9: Geodynamic model suggested from data for the evolution of the western Iberia margin. MAR: middle Atlantic ridge.



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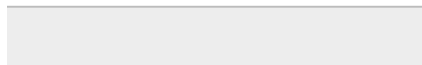
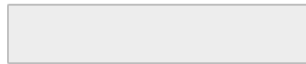
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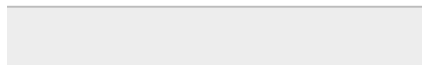
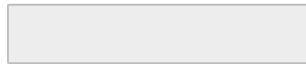
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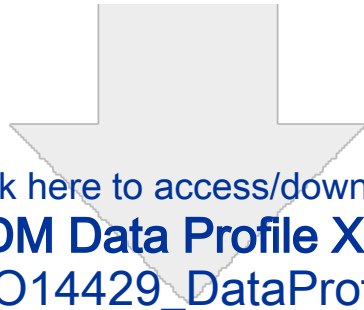




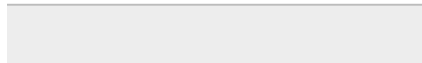
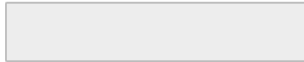
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**Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

J. Barbarand : methodology - investigation – data curation - Writing - Review & Editing

F.O. Marques : conceptualization - Writing - Review & Editing – funding acquisition

A. Hildenbrand : resources - Writing - Review & Editing

R. Pinna Jamme : investigation – data curation

C.R. Nogueira : resources