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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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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 ± 8 Ma and 75 ± 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 5×10^{15} n/cm². Etching conditions were 5 M HNO₃ 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 ζ -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.

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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 ± 8 to 191 ± 8 Ma), with
209 younger ages towards the north. The western unit is characterized by younger ages (75 ± 4 to
210 157 ± 8 Ma), but with similar even younger ages towards the north. The eastern unit displays
211 young ages (78 ± 4 to 110 ± 9 Ma), with two older ages (151 ± 8 and 162 ± 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 (~ 10 km) across the Chaves fault, between
214 sample AN (137 ± 8) located in the central unit and sample BAR (75 ± 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 ± 16 and 97 ± 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 μm , and 13.4 μm 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 μm , except for sample
226 SCD ($SD = 1.0$ μm). All samples have similar D_{par} (1.4-1.7 μm), 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 ± 8 Ma), and has the longest mean track length
246 (13.4 ± 0.1 μm) 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 ± 10 $^{\circ}\text{C}$) during Triassic times. As expected from basin evolution, the temperature increased
257 up to ~ 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 (~ 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 ~ 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 ~ 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).

- 14 339 3. Temperatures higher than surface temperature can be explained by two different models
15 340 considering granite rock samples or sedimentary Triassic rocks. For the Triassic
16 341 conglomerates, heating to 90 ± 10 °C must be by burial of Triassic deposits below thick
17 342 Jurassic ones. For the samples collected in the granitic basement, the temperature may be
18 343 explained in two different ways: (i) The sampled granites were at the surface in the Triassic
19 344 and were buried together with the Triassic sediments (Fig. 8A and 8B). Given the
20 345 geographical distribution of the samples, this would mean that a much wider Jurassic basin
21 346 should have formed extending much further east than the currently known LB boundary.
22 347 Wilson et al. (1989) also noted the absence on the eastern basin margin of significant
23 348 siliciclastic sands in the Lower and Middle Jurassic formations, and suggested that carbonate
24 349 deposition probably extended far into the interior of Iberia. However, there are no known
25 350 Jurassic sediments on top of the Variscan basement, and no detrital deposits are known in
26 351 the LB showing evidence of erosion of all the possible thick Jurassic sediments covering the
27 352 basement to the east. This assumption may however be true for the central unit samples
28 353 where the thermal histories are similar to that of the Triassic sample and have relatively old
29 354 FT ages (SCD, NEL, VTA). Consideration of this assumption on a wider scale has yet to be
30 355 done. (ii) The sampled granites were at a certain depth (equivalent to ca. 100 °C) at the time
31 356 of deposition of the Triassic conglomerate, the upper section of the granitic massif being at
32 357 the surface to feed erosion (Fig. 8A). During the Jurassic, the Triassic conglomerate was
33 358 buried below thick Jurassic sediments, and along the master bounding fault at a similar depth
34 359 to that of the basement samples were (Fig. 8B). Later, at ca. 150 Ma, footwall uplift occurred
35 360 differentially along the master bounding faults (Porto-Tomar-Badajoz, Chaves and Vilariça
36 361 faults) to produce a major rift shoulder uplift that exhumed the buried rocks that we sampled
37 362 in both Triassic conglomerates and the granitic basement (Fig. 8C). Consequent erosion
38 363 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
2
3 366 Triassic conglomerate, but some similar data has been measured in the LB: AFT ages of
4
5 367 169±8, 204±9, 132±16 and 147±10 Ma have been measured in four sandstones from
6
7 368 different stratigraphic levels, respectively Upper Carboniferous, Triassic, Upper Jurassic and
8
9 369 Lower Cretaceous (Pereira et al., 1998). These ages also show that Upper Jurassic and Lower
10
11 370 Cretaceous sediments were not significantly affected by annealing in the LB.

12
13 371 We suggest that differential burial of the Triassic (subsiding hanging wall) and the
14
15 372 Variscan basement (stable footwall with most of the granite samples at a certain depth) and the
16
17 373 later simultaneous uplift of both is the most plausible explanation for the high temperatures
18
19 374 recorded by FT data.

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

24
25 377 Two phases for the beginning of cooling have been defined by the thermal modelling of
26
27 378 different samples: ~160-150 Ma from the western and central units and ~120-110 Ma from the
28
29 379 eastern unit. The offset between these two phases can be interpreted in two different ways. Two
30
31 380 uplift episodes separated by ~40 Ma, or a single, slower, uplift episode. In the latter case, the
32
33 381 age offset would correspond to the duration required for the eastern unit to cross the isotherms,
34
35 382 with samples from this unit being originally much deeper in the crust than samples from the
36
37 383 western and central units. Coincidentally, these ages correspond to the limits of the interval
38
39 384 ascribed in the literature to rift-to-drift – 160 to 110 Ma. FT ages are younger towards the north
40
41 385 and this is consistent with the proposed diachronism for rifting in western Iberia. Jagoutz et al.
42
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
44
45 387 oceanic crust started to form. Our AFT data also indicates cooling at this age, and therefore we
46
47 388 agree with Jagoutz et al.'s (2007) interpretation that ca. 110 Ma marks the rift-to-drift stage.
48
49 389 Timing of the uplift in southern Portugal is different from northern Portugal as the probable age
50
51 390 is Late Cretaceous to early Palaeocene. Fernandez et al. (2013) showed that peak temperatures
52
53 391 were attained in the Mesozoic rocks during the latest Cretaceous-early Palaeocene times
54
55 392 (Fernandez et al., 2013). Their modelled palaeotemperatures suggest also that the Jurassic
56
57 393 sedimentary rocks entered the oil-window at the beginning of the Cretaceous, with peak oil
58
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.

26 487

27 488

28 489 **6. Conclusions**

29 490

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)

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.
10 502

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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17 509

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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; ρ – track density ($\times 10^5$ tracks/cm²); N - number of tracks counted; $P(\chi^2)$ - probability
744 of obtaining Chi-square value (χ^2) for n degrees of freedom (where n = number of crystals - 1);
745 Age $\pm 1\sigma$ - central age ± 1 standard error (Galbraith and Laslett, 1993); MTL - mean track
746 length (μm); SD - standard deviation of track length distribution (μm); 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 ± 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 μm). 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
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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.

804

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).

812

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

818

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.

823

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

825

Code	Lithotype	U-Pb age (Ma)	Latitude (°)	Longitude (°)	Altitude (m)	n	r _s	N _s	r _i	N _i	r _d	N _d	P(c ²)	RE	U (ppm)	FT age (Ma)	N	L (μm)	SD (μm)	Dpar (μm)	
Western Unit																					
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	75±4	87	12.0±0.1	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	87±4	58	12.6±0.2	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	118±5	8	12.6±0.5	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	124±5	100	12.6±0.1	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	118±6	22	12.9±0.2	1,0	1.3±0.1	
Central Unit																					
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	137±8	100	12.5±0.1	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	156±8	100	12.2±0.1	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	172±7	100	12.7±0.2	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	191±8	100	13.4±0.1	1,0	1.5±0.1	
Eastern Unit																					
AC	monzonite		40.76379	-6.85296	711	16	2.882	1568	2.079	1131	6.14	6213	7.3	9.9	41	151±8	100	12.8±0.2	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	110±9	16	13.3±0.3	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	100±5	85	12.6±0.2	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	94±4	101	12.3±0.1	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	78±4	92	12.6±0.2	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	162±8	101	12.2±0.2	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	157±8	100	12.0±0.1	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	97±7	10	13.2±0.4	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	88±16	-	-	-	1.4±0.1

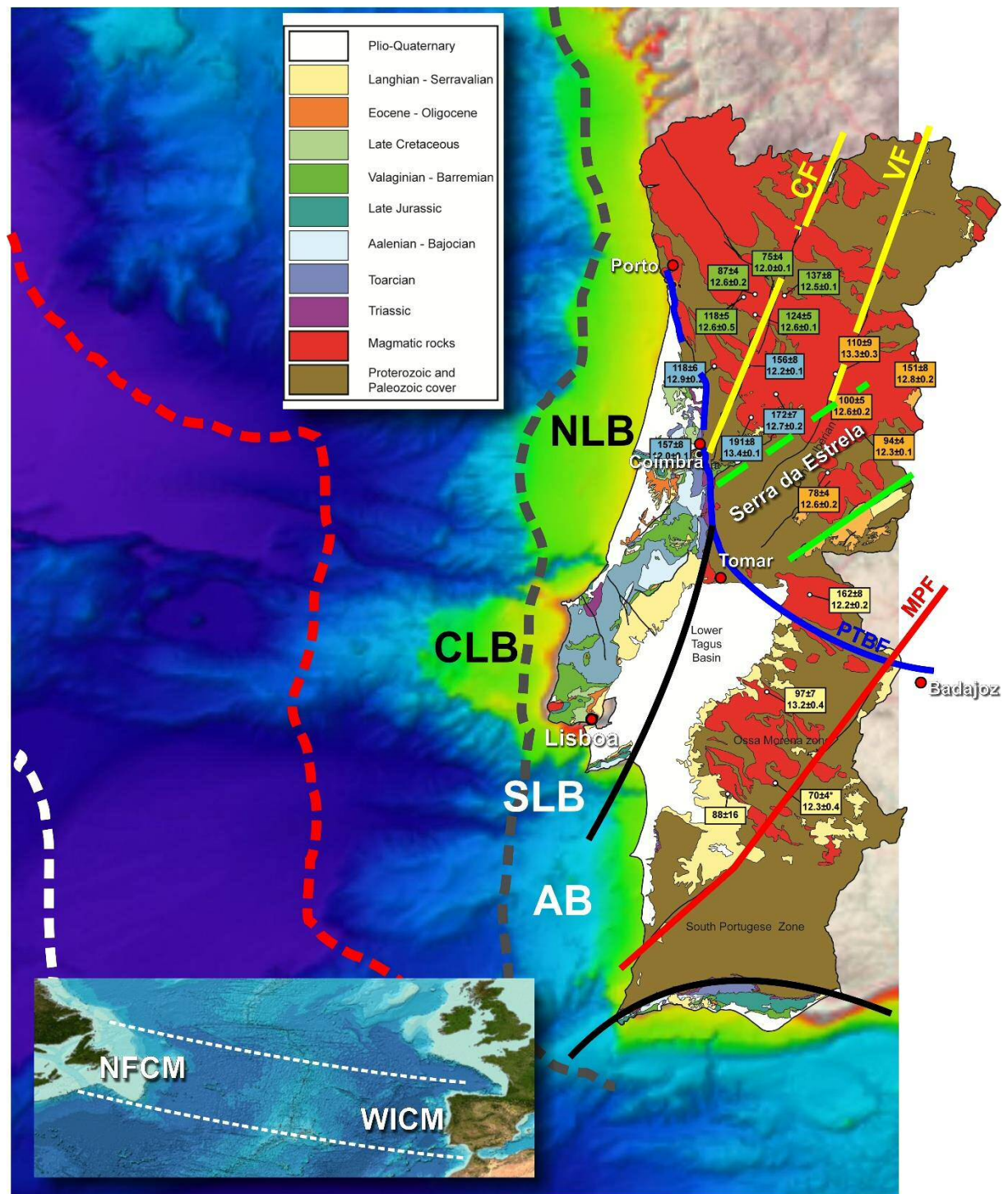


Figure 1. Simplified geological map with squares marking sample location, and showing sample reference, AFT ages (top, in Ma) and track length (bottom, in μm). Color of the squares (blue, green, orange and yellow) corresponds respectively to the western unit, central unit, eastern unit and samples from 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

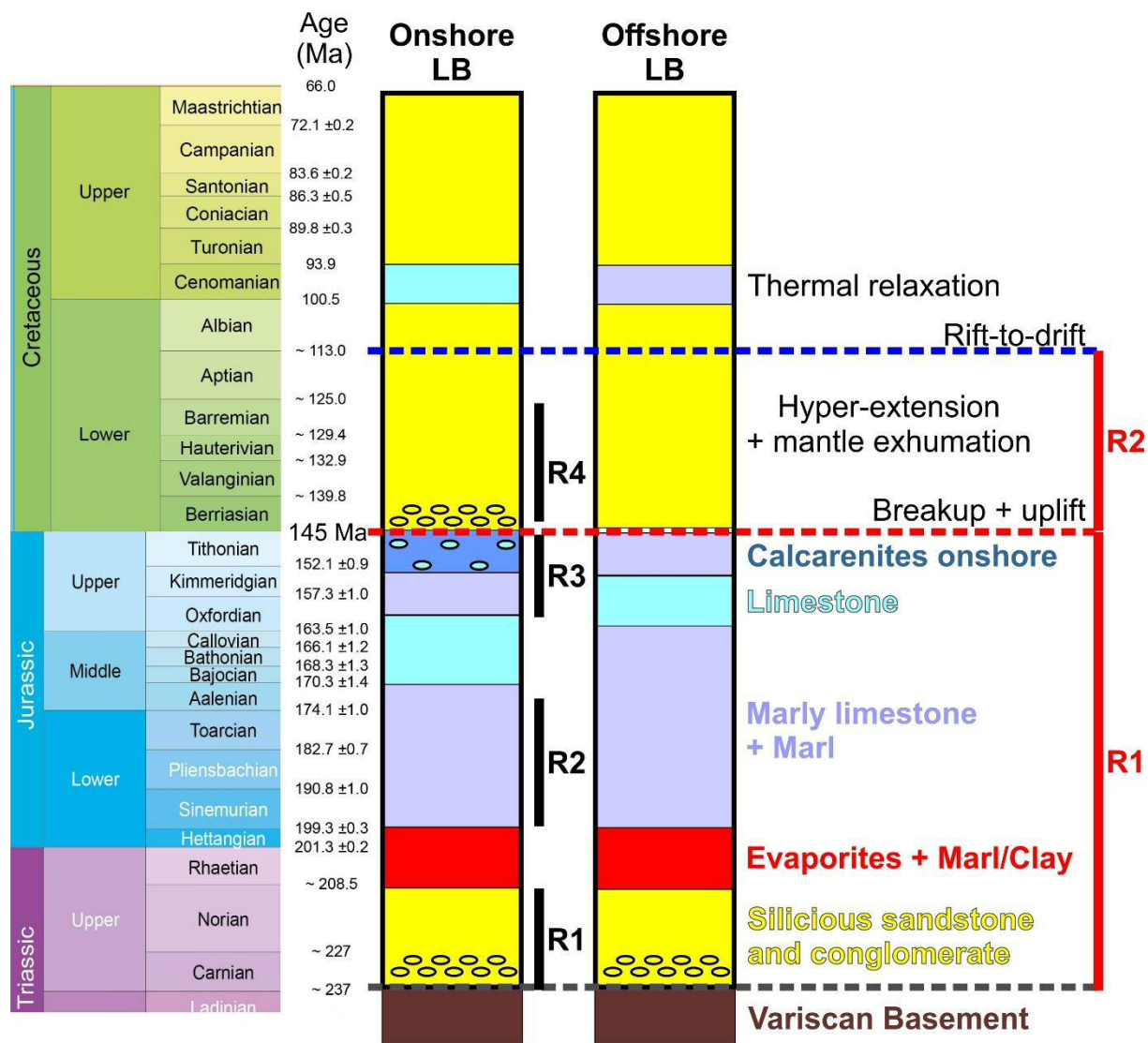
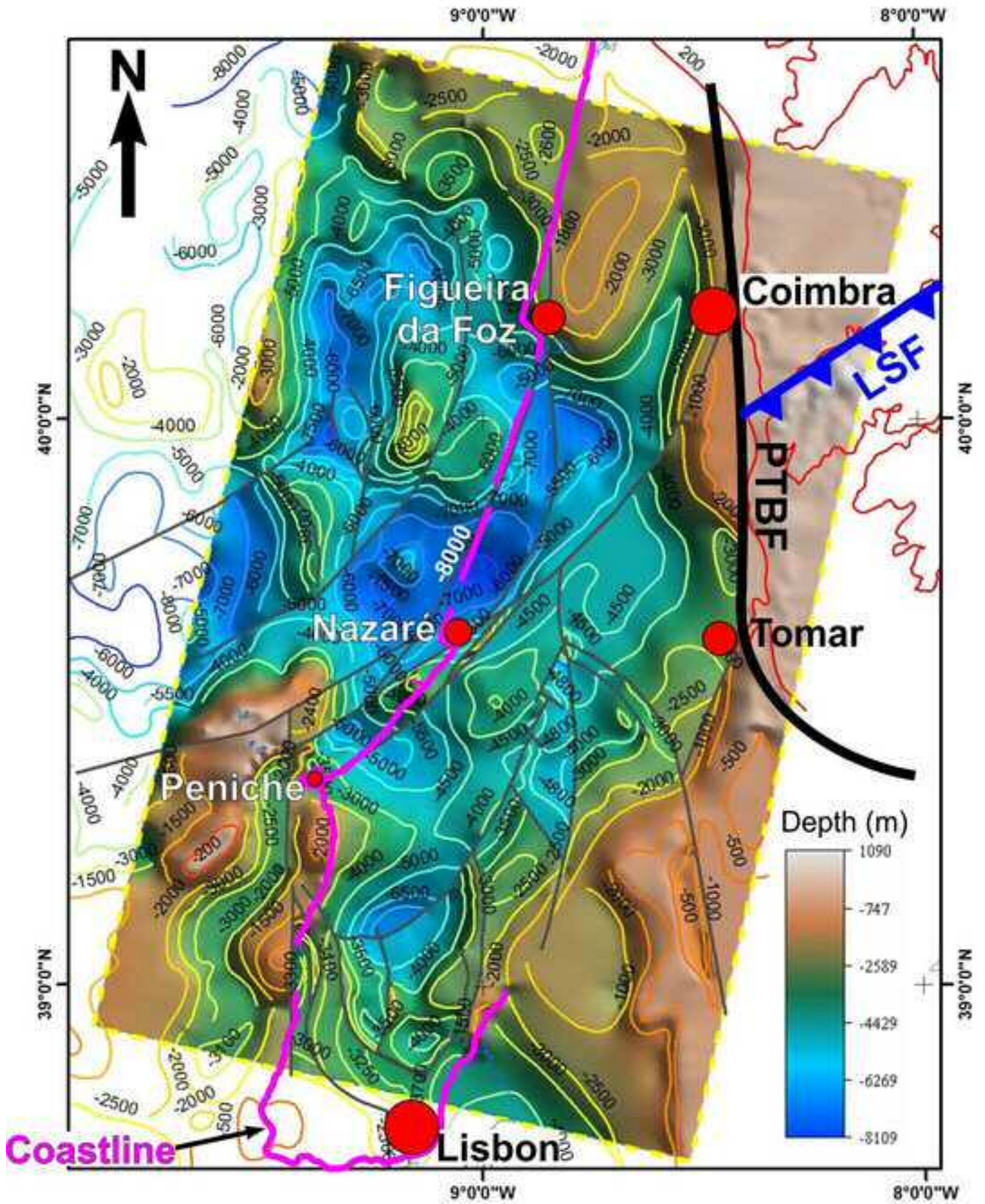
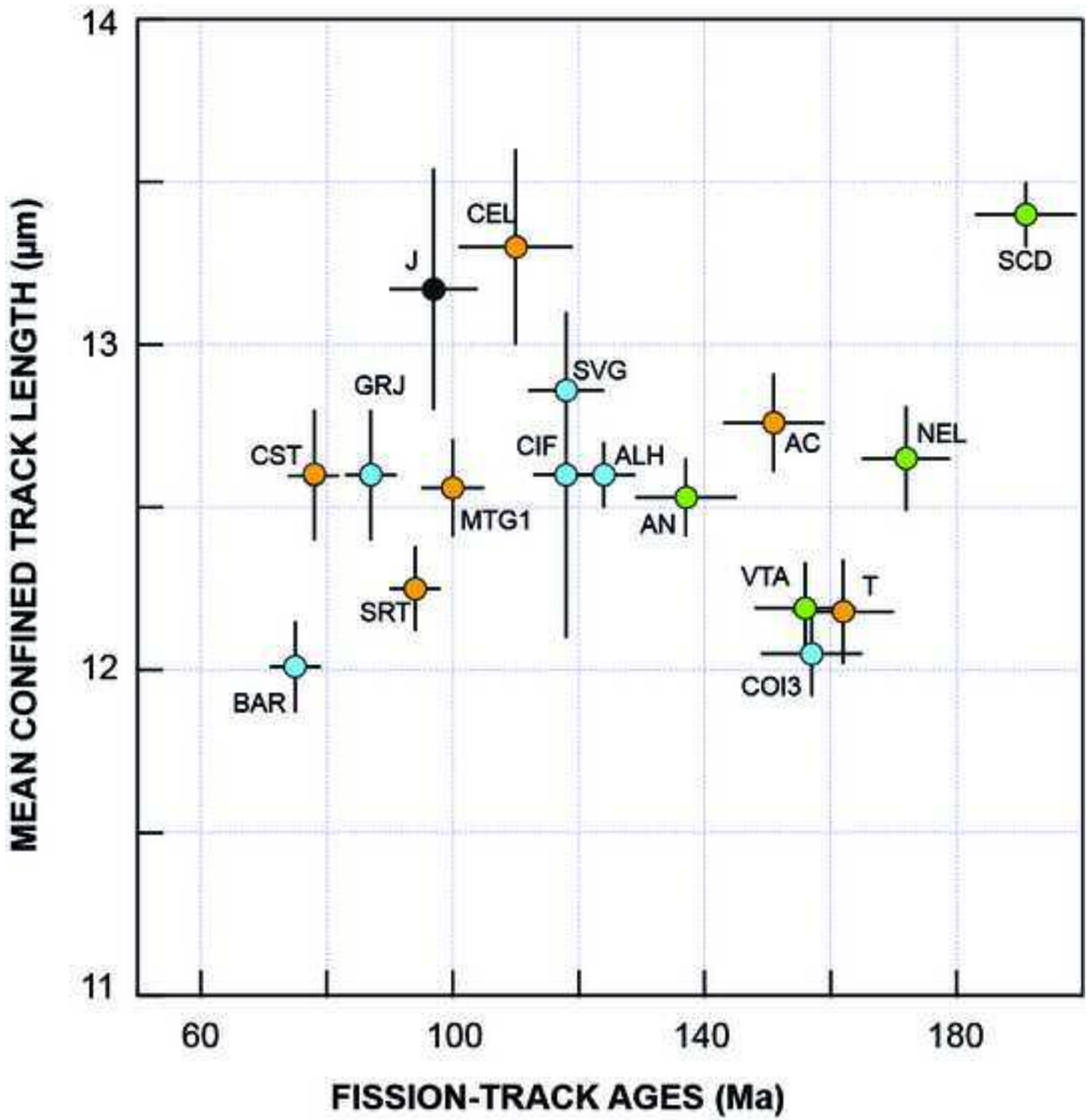


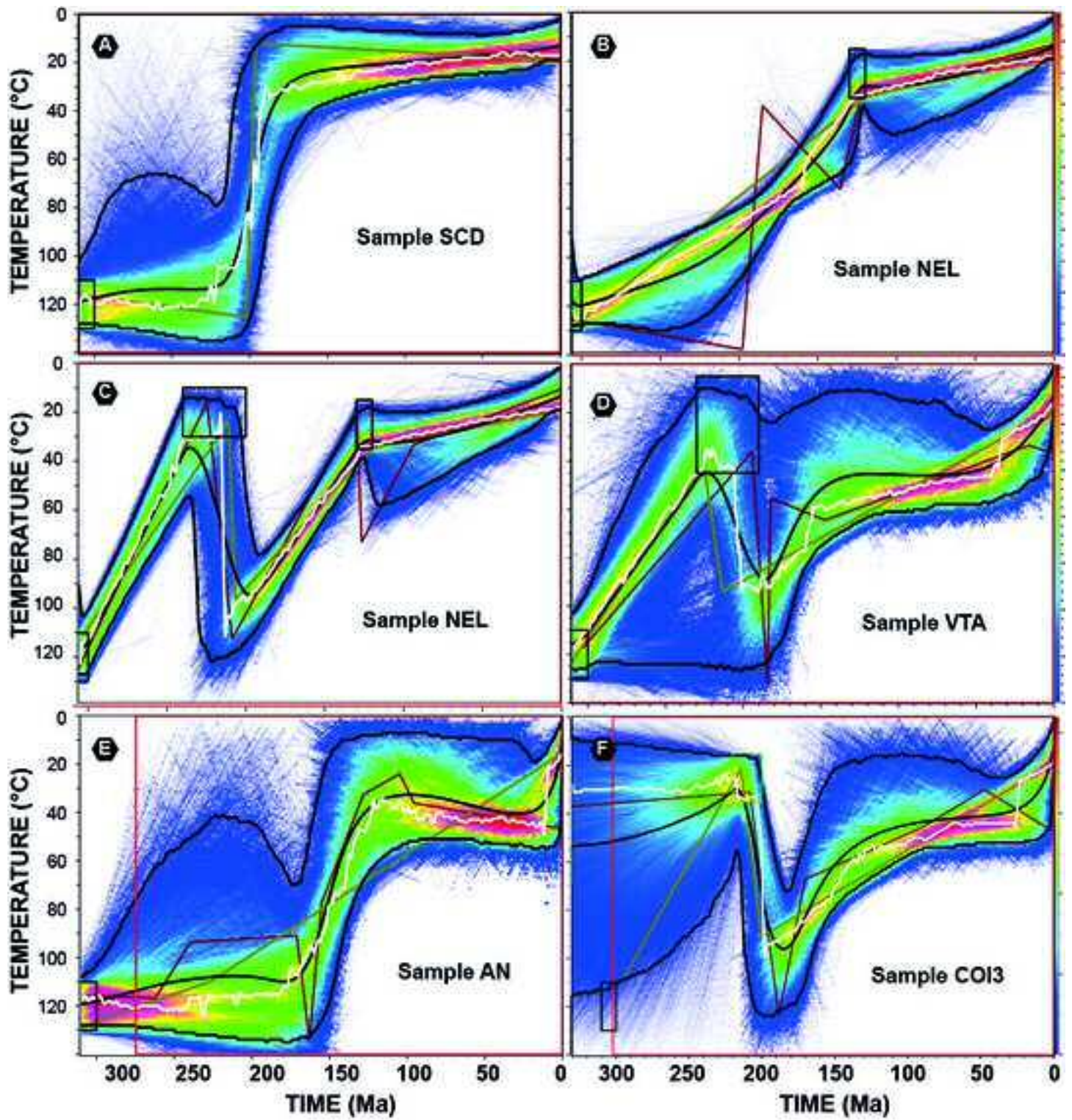
Figure 2. Synthetic lithostratigraphic logs of the onshore and offshore Lusitanian Basin between Figueira da Foz (north) and Nazaré (south) (see Fig. 3 for location). Onshore data are from Rocha et al. (1981), Wilson et al. (1989) and Azerêdo (2007). Offshore data from figure 14 in Alves et al. (2009).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

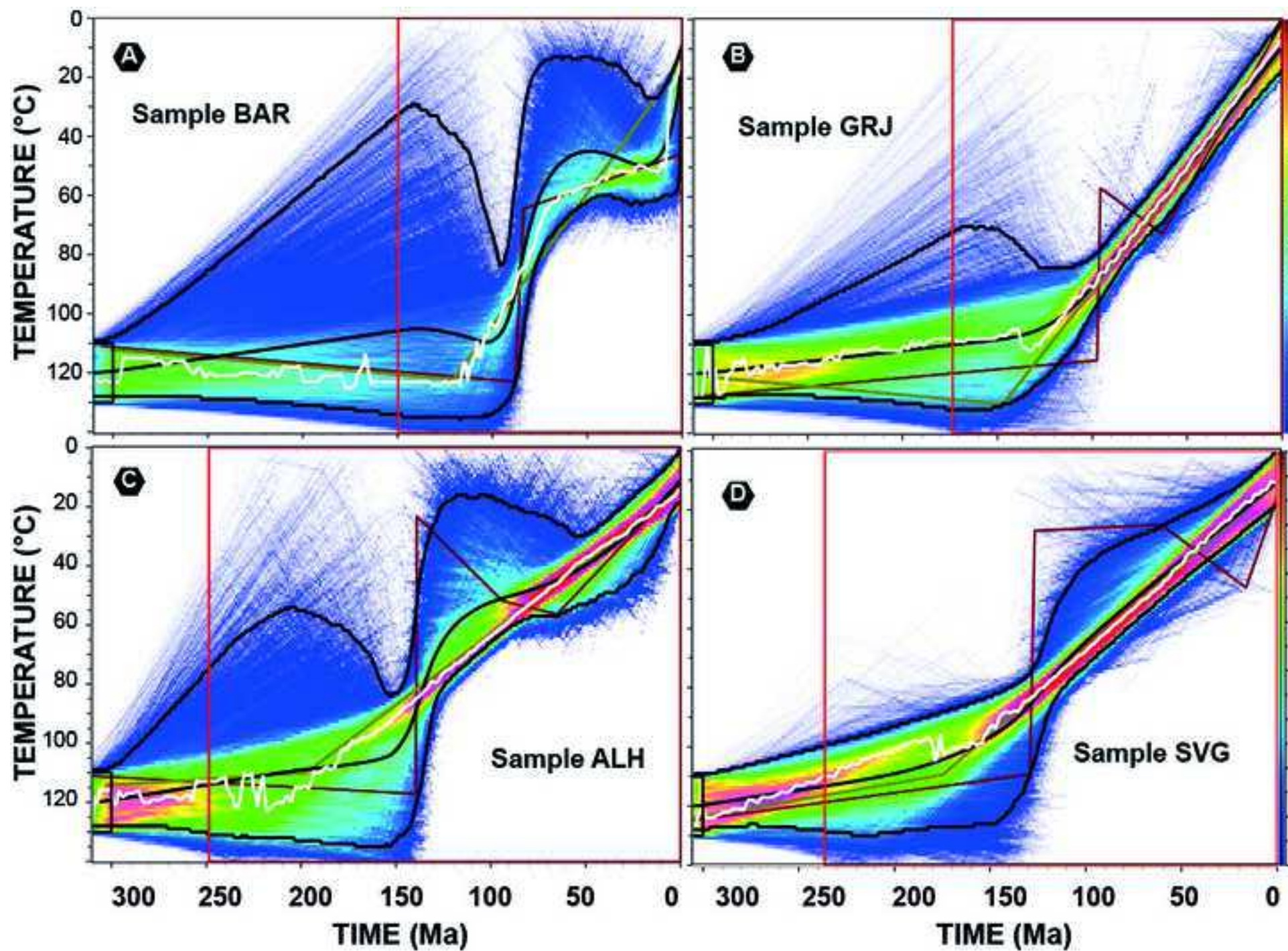
Figure 3

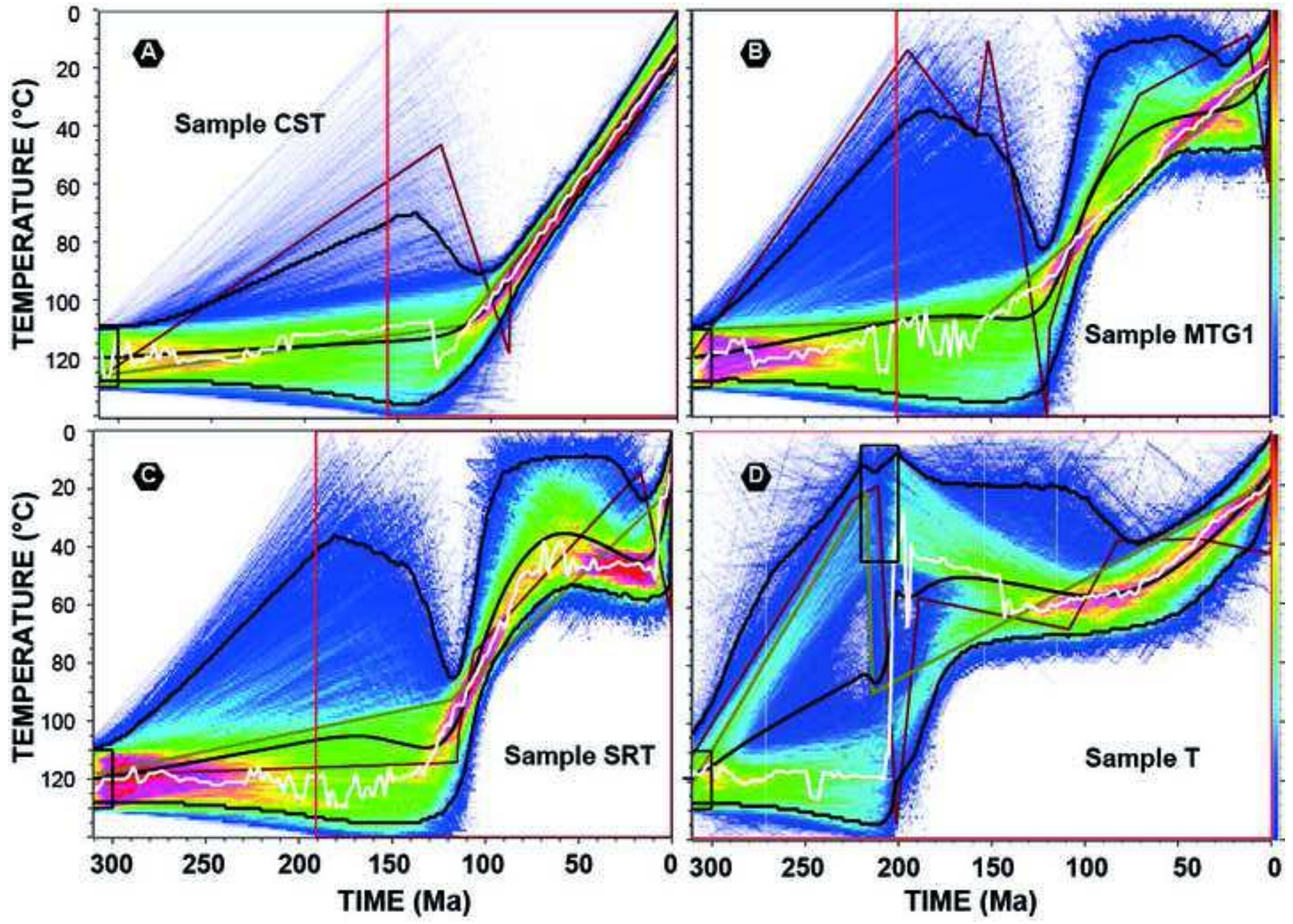
Click here to access/download;Figure (with caption below and on the same page);Figure 3 Lusitanian Basin.jpg











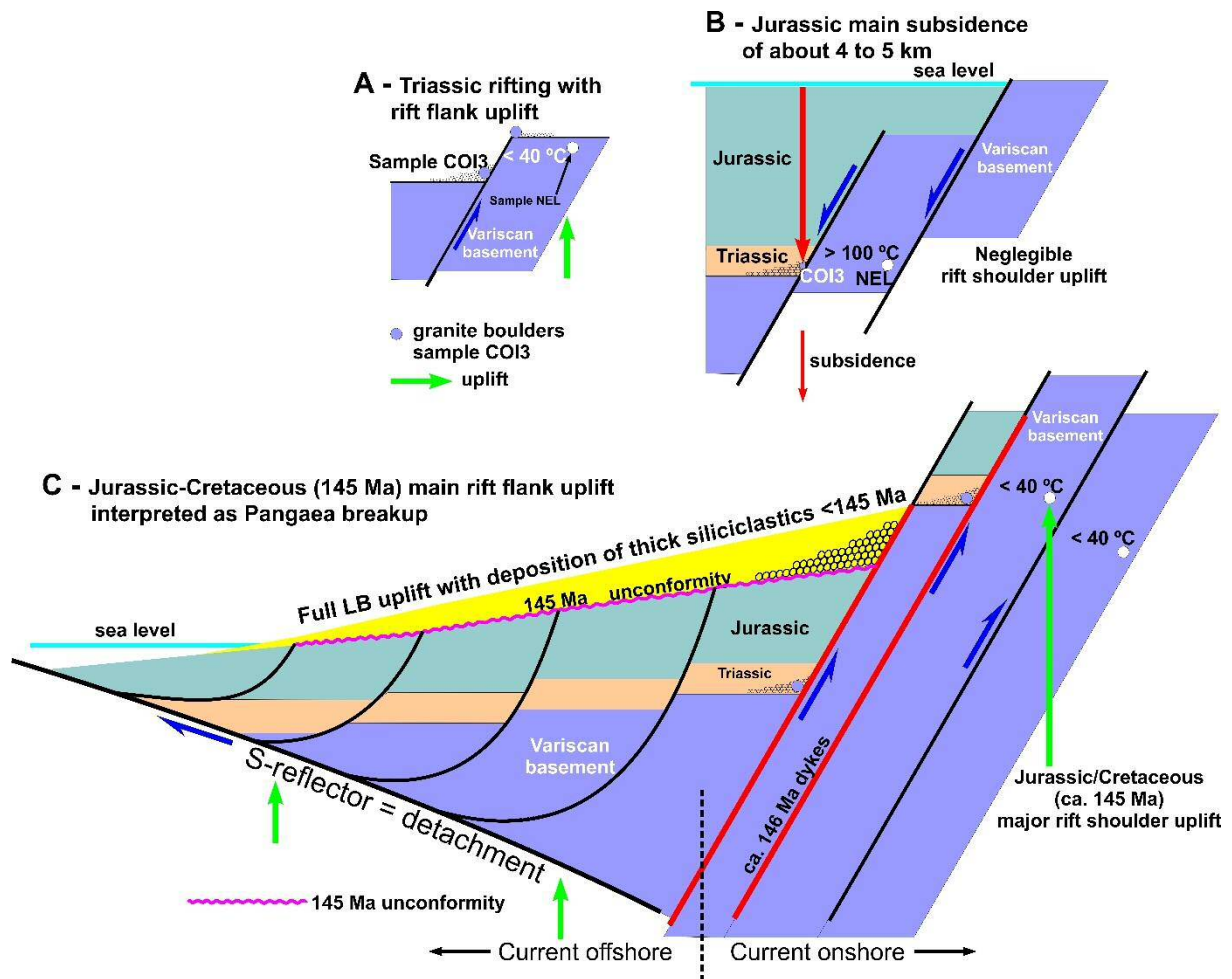


Figure 8. Interpreted evolution of the LB's eastern margin since the Triassic, supported by the geological and AFT data reported here. A – late-Variscan orogenic collapse to Triassic early rifting of Pangaea. B – main subsidence during the Jurassic, including the Hettangian (evaporites up to 3,000 m 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).

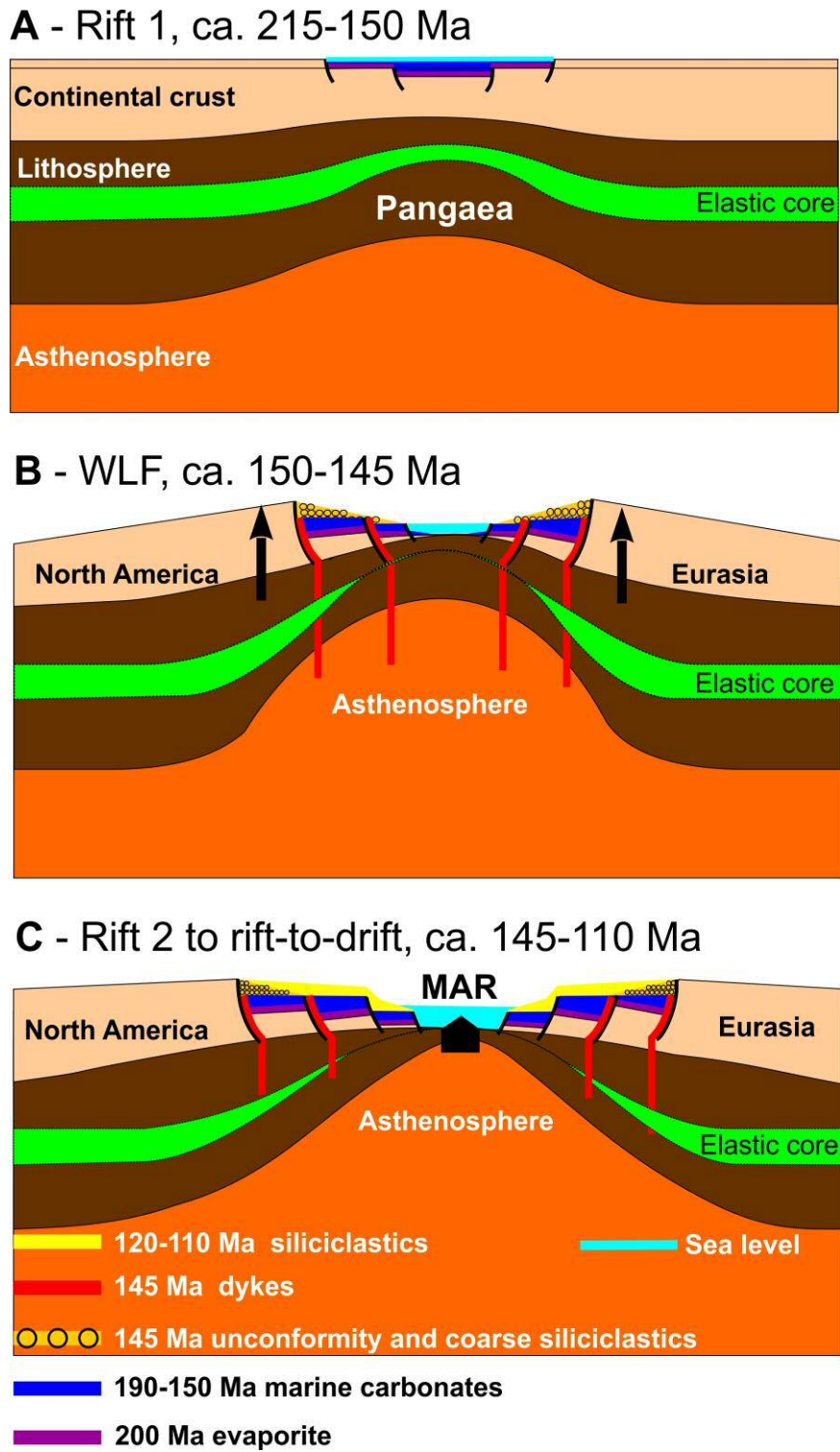
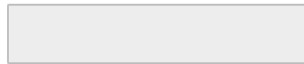


Figure 9: Geodynamic model suggested from data for the evolution of the western Iberia margin. MAR: middle Atlantic ridge.



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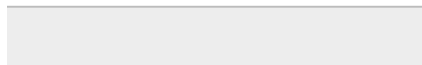
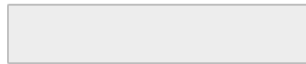
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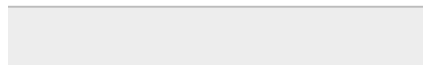
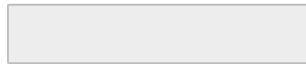
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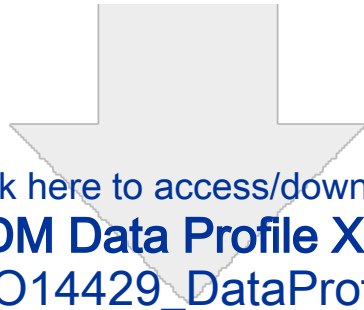




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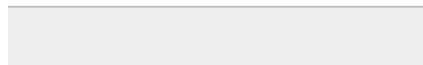
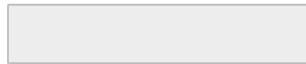




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RDM Data Profile XML

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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