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1 Thermal imprint of rift-related processes in orogens as 2 recorded in the Pyrenees

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19 Abstract

20 The extent to which heat recorded in orogens reflects thermal conditions inherited from
21 previous rift-related processes is still debated and poorly documented. As a case study, we
22 examine the Mauléon basin in the north-western Pyrenees that experienced both extreme
23 crustal thinning and tectonic inversion within a period of ~30 Myrs. To constrain the time-
24 temperature history of the basin in such a scenario, we provide new detrital zircon fission-
25 track and (U-Th-Sm)/He thermochronology data. The role of rift-related processes in
26 subsequent collision is captured by inverse modeling of our thermochronological data, using
27 relationships between zircon (U-Th-Sm)/He ages and uranium content, combined with
28 thermo-kinematic models of a rift-orogen cycle. We show that the basin recorded significant
29 heating at about 100 Ma characterized by high geothermal gradients (~80°C/km). Our
30 thermo-kinematic modeling and geological constraints support the view that subcontinental

31 lithospheric mantle was exhumed at that time below the Mauléon basin. Such a high
32 geothermal gradient lasted 30 Myr after onset of convergence at ~83 Ma and was relaxed
33 during the collision phase from ~50 Ma. This study suggests that heat needed for ductile
34 shortening during convergence, is primarily inherited from extension rather than being only
35 related to tectonic and/or sedimentary burial. This should have strong implications on tectonic
36 reconstructions in many collision belts that resulted from inversion of hyper-extended rift
37 basins.

38 **1. Introduction**

39 The steady-state thermal structure of collisional orogenic belts is controlled by upward
40 advection of heat through the coupling between crustal shortening and erosion (Royden,
41 1993; Stüwe et al., 1994; Willett and Brandon, 2002). However, considering typical thermal
42 relaxation time of several 100 Myrs for thick lithospheres (Jaupart and Mareschal, 2007),
43 transient effects might not be negligible for continental margins that experienced both thermal
44 resetting during thinning and structural inversion over a relatively short period of time
45 (Mouthereau et al., 2013). This process might be even more significant for inverted distal
46 margins that have experienced extreme crustal thinning and mantle exhumation (Manatschal,
47 2004). The pre-orogenic temperature anomalies caused by crust/subcontinental lithospheric
48 mantle (SCLM) thinning, may therefore significantly impact the thermal history and thermal-
49 dependent ductile mechanisms of deformation in orogens, but their magnitude has yet to be
50 constrained. For instance, Mesalles et al. (2014) using low-temperature thermochronological
51 data in southern Taiwan demonstrated that an originally hot distal margin may record cooling
52 only ~20 Myrs after the end of rifting due to the onset of continental accretion.

53 Here, we focus on the Pyrenees, where geochronological and petrographic constraints indicate
54 that rifting exhumed the SCLM in Albian times (ca. ~110 Ma) (Vielzeuf and Kornprobst,
55 1984; Lagabrielle and Bodinier, 2008; Jammes et al., 2009; Lagabrielle et al., 2010; Clerc et
56 al., 2012) while plate convergence initiated at ~83 Ma (Rosenbaum et al., 2002). The
57 Mesozoic Mauléon basin in the north-western Pyrenees (Fig. 1A) is interpreted as a preserved
58 hyper-extended rift system formed during the Late Aptian – Early Albian, above a low-angle
59 detachment system (Johnson and Hall, 1989; Jammes et al., 2009; Masini et al., 2014). This is
60 supported by field evidences of breccias of mantle peridotites reworked in syn-/post-rift
61 sediments of Albo-Cenomanian age, or tectonically overlying the granulitic complex of the
62 Labourd-Ursuya Massif (Jammes et al., 2009).

63 To establish the time-temperature history of the Mauléon basin, we inverse modeled detrital
64 zircon fission-track and (U-Th-Sm)/He thermochronological data collected for this study.
65 Model results were then compared to thermal patterns predicted from a 1D thermo-kinematic
66 numerical model of the evolution of hyper-extended rift basins that are inverted during
67 collision. Our results reveal that high geothermal gradients, inherited from hyper-extension,
68 are maintained over 30 Myrs after convergence initiated.

69 **2. Hyper-extension in the Pyrenees and thermal constraints**

70 The Pyrenean belt resulted from the inversion of previously extended domains of the Iberian
71 and European plates from the Late Cretaceous to the Late Oligocene (Choukroune et al., 1989
72 and references therein). The Mauléon basin, located in the western part of the North Pyrenean
73 Zone, consists of folded Mesozoic sedimentary units, thrust northward during the Tertiary
74 along the Saint-Palais Thrust and the North-Pyrenean Frontal Thrust (Fig. 1). The basin is a
75 Late Aptian to Albo-Cenomanian sag basin interpreted as a hyper-extended rift basin formed
76 above a low-angle extensional detachment system, which is identified on the northern flank of
77 the Labourd-Ursuya massif (Jammes et al., 2009) and at the base of the Igountze-Mendibelza
78 massif (Johnson and Hall, 1989). In such a hyper-extended system, middle-lower crustal
79 rocks and the SCLM were exhumed (Jammes et al., 2009; Masini et al., 2014), but age
80 constraints on the timing of exhumation are still lacking. The only age associated to this
81 extension phase is obtained in the eastern part of the Mauléon basin, where a gabbroic dyke,
82 intruding the exhumed mantle body of Urdach, is sealed by Cenomanian sediments (Jammes
83 et al., 2009; Debroas et al., 2010), and yields a relative flat Ar-Ar on biotite spectrum in the
84 105-108 Ma range (Masini et al., 2014).

85 Alkaline magmatism, high-temperature (up to 600°C) low-pressure (HT-LP) metamorphism
86 (Montigny et al., 1986; Golberg and Leyreloup, 1990), and hydrothermal fluid circulation
87 associated with talc-chlorite mineralization (Boulvais et al., 2006) are observed elsewhere in
88 the North Pyrenean Zone. These constraints indicate a heating episode from 110 Ma to 85 Ma
89 (Montigny et al., 1986). Raman spectroscopy of carbonaceous material (RSCM) shows that
90 the Albian to Cenomanian series of the Mauléon basin experienced shallow to mid-crustal
91 temperatures of 180 to 295°C (Clerc and Lagabrielle, 2014). Determining whether these
92 temperatures reflect thinning processes is not trivial and requires a thermal modeling
93 approach.

94 An important delay occurred between the onset of plate convergence at ~83 Ma, (chron A34)
95 used in plate reconstructions (Rosenbaum et al., 2002) and the exhumation in the belt
96 recorded from ~50 Ma to ~20 Ma, as constrained by low-temperature thermochronological
97 data mainly in the Axial Zone (Yelland, 1990, 1991; Morris et al., 1998; Fitzgerald et al.,
98 1999; Sinclair et al., 2005; Jolivet et al., 2007; Maurel et al., 2008; Gunnell et al., 2009;
99 Metcalf et al., 2009). Time-temperature history (burial, heating and cooling) during the initial
100 accretionary stage are therefore largely unknown. It may involve underestimated competing
101 cooling processes such as syn-orogenic thermal relaxation or cooling by underthrusting as
102 suggested recently in Taiwan (Mesalles et al., 2014).

103 **3. Sampling and methods**

104 **3.1. Strategy**

105 Determining thermal histories of crustal rocks is classically done using multiple low-
106 temperature thermochronometers on bedrock samples, but, in the Pyrenees, published bedrock
107 low-temperature data are only consistent with episodes of collision-related cooling. To gain
108 resolution on syn- to post-rift time-temperature history, a direct approach relies on examining
109 syn-rift basins that recorded both extension and compression in the North Pyrenean Zone. In
110 this aspect, the Mauléon basin is suitable as it experienced temperatures in the 110-295°C
111 range (Fig. 1A). Its time-temperature evolution may therefore be resolved by combining
112 zircon fission-track (ZFT) and (U-Th-Sm)/He (ZHe) thermochronology, which have their
113 Partial Annealing Zone and Partial Retention Zone (PRZ) between 160-270°C and 140-
114 220°C, respectively (e.g. Brandon et al., 1998; Guenther et al., 2013). When zircons reside
115 in these temperature intervals, the resulting ages are highly dependent on their time-
116 temperature histories and diffusion kinetics. Recently published helium diffusion models
117 reveal that apparent (U-Th-Sm)/He ages are controlled by the amount of α -recoil damage,
118 which is proportional to the effective U concentration [eU] (Flowers et al., 2009; Gautheron et
119 al., 2009; Guenther et al., 2013). Following these models, the thermal history explaining
120 these ZHe age and eU correlations can be deduced from inverse modeling.

121 The dataset consists of five detrital sandstone samples from deep-water turbidites of the
122 western part of the Mauléon basin, deposited at 101 ± 4 Ma in Albo-Cenomanian (Su-1, Ar-
123 2, Lu-1, Ch-1, and Mi-1, Fig. 1A, B). Two basement samples from a granitic intrusion (Itx-1)
124 and gneiss (Lag-1) were collected (Fig. 1) to complete these data with apatite (AHe) and ZHe

125 analyses, and to add independent constraints on collision-related cooling. Lu-1 was analysed
126 with Raman Spectroscopy following the protocol developed by Lahfid et al. (2010). Samples
127 were prepared at CRPG (Nancy, France). They were crushed and zircon and apatite grains
128 were separated for low-temperature thermochronological analyses using standard heavy-
129 liquid and magnetic separation from the 61-280 μm fraction.

130 **3.2. Zircon Fission track analysis**

131 Zircon grains were handpicked for fission track dating and analysed at the thermochronology
132 laboratory of ISTERre (Université Joseph Fourier, Grenoble). We used standard ZFT
133 preparation procedures as described by Bernet and Garver (2005). Grains were mounted in a
134 teflon sheet, polished to expose internal surfaces and etched with NaOH-KOH at 228°C
135 between 8 and 13 hrs. Irradiation was performed in the FRM II Research Reactor at the
136 Technische Universität München (Germany) with CN1 dosimeter glasses and Fish Canyon
137 Tuff age standards. Mica detectors used for external detector method (Gleadow et al., 1976)
138 and standards were then etched in 48% HF at 21°C for 18 min. Counting was carried out on a
139 Olympus BH2 optical microscope using FTStage 4.04 system of Dumitru (1993). Individual
140 fission-track ages were obtained using zeta factors following approach of Hurford and Green
141 (1983).

142 **3.3. Zircon (U-Th-Sm)/He analysis**

143 For (U-Th-Sm)/He dating, we handpicked 5 to 10 zircon grains per bedrock sample and
144 between 60 and 110 zircon grains per detrital sample. Detrital zircons were mounted in epoxy
145 and polished for future U/Pb analyses. Among those zircons, we retrieved from the epoxy
146 mounts between 20 and 35 zircons per detrital samples for (U-Th-Sm)/He analysis, chosen so
147 as to represent the main peaks individualized in the U/Pb age distributions. All these detrital
148 zircons were measured as single grains. Additionally, 4 to 5 replicates of 5 to 10 zircon grains
149 were analysed in bedrock samples (Itx-1 and Lag-1). Zircon grains (prismatic to round-
150 shaped, 0 to 2 pyramids, with an equivalent spherical radius ranging from 35 to 60 μm) were
151 then measured, and loaded in Pt capsules for He extraction at CRPG. They were outgassed at
152 1500°C for 20 min, and analyzed for He concentrations with a VG603 noble gas mass
153 spectrometer (Pik et al., 2003; Godard et al., 2009). After total helium extraction, Pt capsules
154 containing zircon grains were retrieved for U, Th, and Sm content measurements at SARM
155 (Nancy, France). Pt capsules were opened, and loaded in Pt crucibles along with ultra-pure

156 LiBO₂ and ultra-pure B(OH)₃ for 2 hrs at 990°C in an automatic tunnel oven. Then, the Pt
157 crucibles were digested 12 hrs into acid. The preparation was then analysed using an
158 inductively coupled plasma mass spectrometer. The overall precision of He ages determined
159 with this procedure is within 5-6% (1σ). Zircon grains whose He and/or U concentrations are
160 too close from the blank (e.g. for He content less than 1.10⁻¹³ moles and for U concentration in
161 the solution less than 100 ppb after blank correction) were not considered for this study.
162 Zircon ages were corrected for α-ejection (F_T) following Ketcham et al. (2011) (Table DR2,
163 DR3). To account for the abrasion of the detrital zircon single grains, we consider that we
164 deleted ~20 μm (the mean stopping distance (Ketcham et al., 2011)) and the half of the mean
165 width of our zircon grains (~45 μm). Following Reiners et al. (2007), we corrected F_T
166 considering an abrasion of 45 μm for each detrital grain.

167 **3.4. Apatite (U-Th-Sm)/He analysis**

168 Apatites were prismatic, with 0 to 2 pyramids, and with an equivalent spherical radius ranging
169 from 60 to 160 μm. We performed AHe analyses at Paris-Sud University (Orsay, France) on
170 bedrock samples (Lag-1 and Itx-1, Fig. 1) following the procedure described by Fillon et al.
171 (2013). Four single grain replicates were analyzed for Itx-1 and Lag-1 with 8% precision (1σ).
172 AHe ages were corrected for α-ejection following Ketcham et al. (2011) and apatites with
173 outlier Th/U ratios were excluded (Table DR4).

174 **4. Results**

175 ZFT analyses performed on samples Su-1, Ar-2, Ch-1, and Mi-1 yielded 23 to 63 dated grains
176 per sample (Table DR1). Each sample with an identical depositional age shows a similar age
177 distribution. We therefore only present age component distributions for the combined samples
178 (n=171) (Fig. 2A). Most of the grains (97%) are older than the depositional age, indicating
179 very minor resetting after deposition.

180 We decomposed our age distribution into age components using DensityPlotter (Vermeesch,
181 2009, 2012). The software represents distribution of ages using KDE (Kernel Density
182 Estimation), which is determined by stacks of Gaussian curves on top of each measurement,
183 whose standard deviation is determined by the local probability density. Deconvolution for
184 combined data returned three age components (errors are given as ±2σ): two majors at 134 ±

185 46 (P1, 17%) and 236 ± 40 (P2, 79%), considered as cooling events, and a minor population
186 at 1005 ± 886 Ma (4%) characterized by a too important error to be statistically meaningful
187 (Fig. 2A).

188 ZHe analyses were carried out on the same Su-1, Ar-2, Ch-1 samples and on Lu-1. Ten to 27
189 grains were dated per sample (Table DR2) and show similar ages and eU distributions. The
190 age distribution from the combined data ranges from 36 Ma to 131 Ma (n=75, Fig. 2B) and
191 yields four age peaks at 39 ± 4 (16%), 50 ± 3 (38%), 68 ± 4 (33%), and 116 ± 7 Ma (13%).
192 Most of these detrital ZHe grain ages (87%) are younger than the depositional age, suggesting
193 that they have been, at least, partially reset by post-deposition burial. We will test the timing
194 and amount of burial and exhumation through numerical inversion of the data in the next
195 section.

196 ZHe analyses on two bedrock samples from the Labourd-Ursuya massif (Lag-1 and Itx-1)
197 give ages ranging from 51 ± 5 to 74 ± 7 Ma, and from 61 ± 6 to 86 ± 9 Ma, respectively
198 (Table DR3). AHe single grain analyses performed on the same samples yield ages ranging
199 from 42 ± 3 to 49 ± 4 Ma and from 35 ± 3 to 43 ± 4 Ma for Lag-1 and Itx-1, respectively
200 (Table DR4).

201 **5. Thermal modeling of partially reset ages**

202 It has been demonstrated that α -recoil damages associated to U and Th decay, and their
203 respective concentration (eU) could affect He diffusion in apatites (Shuster et al., 2006;
204 Flowers et al., 2009; Gautheron et al., 2009). For high eU concentrations, the amount of α -
205 recoil damages increases with He retentivity and closure temperature. Guenther et al. (2013)
206 highlighted the same trend in zircons but only for relatively low concentrations of eU. For
207 very high eU concentrations, the He retentivity rapidly decreases. These authors hypothesized
208 that for very high eU content, the amount of α -recoil damage is high enough so that damaged
209 areas in the crystal are interconnected and form through-going fast diffusion pathways for He.
210 Guenther et al. (2013) showed that the evolution of He retentivity in zircons, which depends
211 on the eU content, controls both variations of the closure temperature and individual
212 annealing behaviors. This effect can lead to large ZHe age distributions, under a given time-
213 temperature path. The non-random distribution of our ZHe age-eU dataset is supported by
214 their statistical distribution in Figure 3 (see caption for details concerning the density function
215 used) and suggests such a control. Originating from a dense zone of young ZHe ages and low-

216 eU grains two opposite trends can be identified as ZHe-eU groups. A first group A (red area)
217 consists in young ZHe ages (from 36 to 65 Ma) associated with a large eU distribution (from
218 400 to 4000 ppm). The group B (blue area) corresponds to older ZHe age (from 65 to 131
219 Ma) associated with low eU values only (from 0 to 1100 ppm).

220 Following Guenther et al. (2013), the oldest ages could correspond to zircon grains that have
221 been less resetted due to a higher closure temperature ($\sim 220^{\circ}\text{C}$). Such zircons require a longer
222 residence time in the PRZ to be reset. In contrast, the young ZHe grain ages that display
223 $\text{eU} > 1100$ ppm would correspond to a lower closure temperature of $< 140^{\circ}\text{C}$.

224 To determine the time-temperature paths of these zircon grains, we used the HeFTy soft
225 (Ketcham, 2005) that includes the kinetic model of Guenther et al. (2013). The limited
226 number of grains (seven) that can be input in the HeFTy inverse modeling procedure do not
227 allows direct inversion of the entire dataset and requires to identify representative individual
228 ZHe age-eU pairs within the two groups observed in Figure 3. These two trends which
229 originate in the red high density zone of Figure 3 can be easily and robustly described by a
230 couple of representative samples. In order to describe the entire range of age-eU distribution,
231 seven representative samples have been taken along the A and B groups and used for distinct
232 sets of inversion. Various tests demonstrated that the use of representative samples is not an
233 issue in this inversion procedure.

234 Because the A and B groups have been potentially controlled by distinct closure temperatures
235 linked to the amount of radiation damage accumulated in the zircon grains (Guenther et al.
236 2013) it is crucial to take into account the ZFT data obtained for these zircons in the inversion
237 modeling. The ZFT data also exhibit two distinct populations characterized by peak ages at
238 P1 (~ 134 Ma) and P2 (~ 236 Ma) that represent two independent cooling histories prior to
239 deposition (at ~ 100 Ma). At that time, the amount of accumulated damages was therefore
240 significantly higher for the P2 population and could have triggered differential He diffusion
241 when the sediments have been subsequently buried and re-heated. It is however not possible
242 to directly relate one ZFT population with one ZHe ages group. Consequently both P1 and P2
243 ZFT populations have been used alternatively as input parameters for the inversion modeling.
244 Four sets of inversion models have therefore been tested (Fig. 4), corresponding to the various
245 combinations using ZFT populations (P1, P2) and ZHe ages groups (A, B). Models are
246 characterized by: (i) different pre-deposition histories constrained by the P1 and P2 ZFT
247 characteristics (134 ± 15 Ma / $240 \pm 40^{\circ}\text{C}$ and 236 ± 20 Ma / $240 \pm 40^{\circ}\text{C}$ respectively), (ii)
248 identical depositional age at 110 Ma, (iii) a free post-depositional time-temperature history
249 inverted for 7 representative ZHe ages of groups A and B independently. To reproduce the

250 partial reset signature of the ZHe data, we constrained the software to search post-deposition
251 time-temperature paths from shallow to mid-crustal temperatures, corresponding to a range
252 from 20°C and 200°C. The inversion consisted of randomly testing 300,000 time-temperature
253 paths for each model.

254 The P1-A model returned 439 acceptable and 180 good solutions. The P1-B model returned
255 only 70 acceptable and no good solutions. The P2-A model only returned 35 acceptable and
256 no good solutions. The P2-B model returned 405 acceptable and 62 good solutions. The best
257 time-temperature path of each model corresponds to a ZHe age-eU correlation, which is
258 compared to the data in Figure 4B. The P1-A model better fits the group A than the P2-A
259 model, which fails to reproduce the data with eU > 2000 ppm. The P2-B model is in better
260 agreement with the group B than the P1-B model. Models P1-A and P2-B return the best
261 correlations between eU and ZHe ages that reproduce the data (Fig. 4B). We infer that all
262 Albo-Cenomanian ZHe detrital data are obtained by the combination of these two time-
263 temperature models (Fig. 4).

264 These models show a consistent post-100 Ma thermal history. In particular, time-temperature
265 paths of our zircon grains are consistent with heating to temperatures of ~180°C soon after
266 ~100 Ma at an average heating rate of ~5°C/Myr. Several of these pathways show a nearly
267 isothermal stage established at ~80 Ma which lasted maximum ~30 Myr. This
268 heating/isothermal stage was followed by a relatively rapid cooling stage from 50 Ma to
269 present (~3°C/Myr). This cooling path is not very precisely described by the inverse
270 modeling because of the lack of additional lower temperature thermochronometers.

271 **6. 1D thermo-kinematic modeling of rift-to-collision evolution**

272 The results of modeling (Fig. 4) show that the Albo-Cenomanian zircon grains were heated to
273 a temperature of ~180°C during the post-breakup evolution of the Mauléon basin. To
274 constrain the geothermal gradient associated with this heating episode, a few Myrs only after
275 deposition of the sampled Albo-Cenomanian rocks (see Figure 4A), we first need to
276 determine the thickness of the entire pile of sediments in the Mauléon basin. The complete
277 burial history shown in Figure 5A was resolved by combining well data from different
278 boreholes, including boreholes in the Arzacq basin, North of the Mauléon basin (for instance
279 Lacq 301, Brunet (1984)), and wells drilled in the Mauléon basin. They are from South to
280 North: Ainhice 1, Chéraute 1, Uhart-Mixe 1 and Saint-Palais 1 for the Triassic to Late

281 Cretaceous history, and Lahontan 1bis, Lacq 301, and Nassiet 1 for the Late Cretaceous to the
282 Late Eocene deposits (Fig. 5B). Estimates of minimum and maximum thicknesses at time of
283 deposition of the studied samples from 105 to 70 Ma are indicated in Figure 1B. We estimate
284 that a mean sediment thickness of ~2 km was deposited above the studied samples during this
285 period of time (Fig. 5A). A temperature of 180°C at ~2 km depth, as suggested from the ZHe
286 data, allows to define a geothermal gradient as high as ~80°C/km (assuming a surface
287 temperature of 20°C).

288 To examine the tectonic conditions that led to the observed cooling history, we consider two
289 different end-member thinning processes (Figs. 6A, B, and 7) that are thought to embody
290 most of the fundamental characteristics of rifted margins as summarized by Huisman and
291 Beaumont (2011). A stepwise tectonic evolution from 130 Ma to 0 Ma of a lithosphere
292 section below the Mauléon basin involves thinning through a rifting phase from the Early
293 Barremian (130 Ma) until the Late Cenomanian (95 Ma) with a breakup occurring at 110 Ma.
294 This is followed by inversion and underthrusting of the thinned lithosphere from 83 Ma until
295 the Early Eocene (50 Ma) that marks the onset of crustal thickening and thrust-related
296 exhumation. In order to test these two hypotheses, we adopt a 1D forward thermo-kinematic
297 modeling approach. The thermochronological modeling showed that the Albo-Cenomanian
298 series in the Mauléon basin were heated to a temperature of ~180°C, which was maintained
299 during ~30 Myrs (Fig. 4). We therefore attempt to retrieve from the model the time-depth
300 evolution of the 180°C isotherm by varying thinning factors for crust and mantle.

301 The thickness of the Mauléon sedimentary layers at time of deposition of Albo-Cenomanian
302 is constrained by the subsidence history of the Mauléon basin defined in Figure 5A. We
303 consider an initial thickness of 2 km of sediments above the continental crust that increases to
304 a final thickness of 6 km. The current Moho depth is estimated to 32 km in the region of the
305 Mauléon basin (Daignieres et al., 1982; Jammes et al., 2010; Chevrot et al., in revision)
306 leading to consider a final thickness of continental basement of 28 km. We hypothesize that
307 the thickness of the continental basement was the same before the onset of rifting. At the
308 initial and final stages we consider that the lithosphere is in equilibrium stage and adopt a
309 typical thickness value for a Phanerozoic continental lithosphere of 130 km (e.g., Poudjom
310 Djomani et al., 2001) (Fig. 6).

311 In model A, the crust is thinned until breakup occurs (Fig. 6A). After crustal breakup, the
312 SCLM is thinned and exhumed at the base of the Mauléon basin, according to variable
313 amount of thinning factors β_A ranging from 4, 10 and higher (SCLM breakup). In model B,
314 the SCLM is thinned until its breakup, leading to the rise of hot asthenosphere below the

315 continental crust (Fig. 6B). Similarly to model A, model B is run for the same variable
316 amount of thinning factor β_B for the continental crust. We also test the impact of the
317 thickening of the SCLM (model A) or the crust (model B), after convergence initiated at 83
318 Ma, on the thermal evolution of Mauléon basin. This was performed by taking into account
319 the accumulation of syn-orogenic sediments under local isostatic conditions.

320 For both scenarios, the role of transient diffusive heat relaxation and advection related to
321 basin subsidence history, isostasy and rock uplift is quantified. It accounts for realistic
322 diffusivity and heat production distribution. To simulate the effect of a high thermal
323 conductivity layer represented by the Triassic evaporites, the basin rests above a 1 km-thick
324 Triassic salt layer, which thickness is kept constant during the simulation. Although fluid
325 circulations may play a key role during extension by maintaining high temperatures below the
326 basin and favoring heat transfers, we kept the model as simple as possible so as to depend on
327 a minimum of unknown parameters, as the basin evolved from extensional to compressional
328 tectonic settings.

329 Model results show that the depth of the 180°C isotherm is controlled to first order by the
330 amount of thinning of the SCLM (Fig. 7). This effect is most significant for model A (SCLM
331 exhumation) in which the depth of the 180°C isotherms is seen to vary between 1 and 5 km as
332 a function of the amount of thinning. In the model B, this is less apparent because the
333 asthenosphere is kept closer to the surface (from 7 km depth to surface depending on the β
334 considered) during all the experiment.

335 Prior to crustal breakup at 110 Ma, model A and B show very different thermal responses to
336 rifting. An upward deflection of isotherms is observed for the model B, while model A
337 indicates a cooling phase before heating. Thermal evolution in model B reflects the upward
338 advection of the base of the lithosphere during thinning, which is maximum when the SCLM
339 breakup is achieved. In model A, a delay is observed between the onset of crustal and SCLM
340 breakup. This reflects the loss of the radiogenic heat source caused by crustal breakup, which
341 is not instantaneously compensated by advection of heat caused by SCLM thinning.

342 During the inversion phase, the 180°C isotherms are maintained to the same depth from 95 to
343 50 Ma for both models (red and black curves on figure 7) corresponding to a limited linear
344 increase of heat with respect to the base of the Mauléon basin. When we account for the
345 thickening of the SCLM or the crust below the basin during the underthrusting phase, the
346 180°C isotherms remain flat from 83 Ma to 50 Ma for both models A and B. The progressive

347 deepening of the 180°C isotherms after 50 Ma reflects the primary effect of thermal relaxation
348 and the deepening of the SCLM, relative to the heat advection due to erosion.

349 **7. Discussion**

350 **7.1. Comparison between thermochronological data and thermo-kinematic** 351 **modeling**

352 In this study, we focus on the thermal history of the Mauléon basin from Albo-Cenomanian
353 times until today. However, because our ZFT ages are only slightly reset (10%) with no
354 significant influence on age populations, we can assume that P1 (~134 Ma) and P2 (~236 Ma)
355 reflect two cooling events that occurred prior to deposition. The P1 cooling event appears to
356 be consistent with extension recorded in the Early Cretaceous at ca. 145–132 Ma (Vergés and
357 García-Senz, 2001). Zircons cooled between 150 Ma and 100 Ma (P1) from mid-crustal to
358 surface temperatures may reflect denudation in the footwall of a rolling-hinge normal fault
359 (Axen and Baertley, 1997). The P2 event is also coherent with a magmatic (Rossi et al., 2003)
360 and/or exhumational event during the Triassic, as recognized in Albian sediments in the
361 southern Pyrenees (Filleaudeau et al., 2011).

362 On the other hand, ZHe age data show a complex and large distribution from 36 to 131 Ma
363 with most of them younger than depositional age. This is typical of partial resetting and, for a
364 given duration of thermal event, its amplitude could have been controlled by various factors
365 including: (i) the size of the grains, (ii) the initial age distribution of grains, (iii) the position
366 in the PRZ during re-heating, (iv) the residence time above the PRZ and the amount of α -
367 recoil damages accumulated before re-heating (Guenther et al., 2013). Inversion of ZHe data
368 with thermochronological models (Fig. 4) suggests that zircon grains have been heated to
369 temperatures up to ~180°C soon after deposition ~100 Ma ago (Fig. 4A, B). This is consistent
370 with our thermo-kinematic models A and B showing that the basin was already hot at the end
371 of the rifting phase (95 Ma), due to upward deflection of the 180°C isotherm reaching the
372 depth of the Albo-Cenomanian series at 2 km for $\beta=10$ or higher (Fig. 7). After this heating
373 phase, both our thermochronological models P1-A and P2-B require that zircon grains were
374 maintained at this temperature of 180°C through a nearly isothermal stage until 50 Ma. This
375 period corresponds to the inversion phase of the thermo-kinematic models, where the 180°C
376 isotherm depth remains constant from 95 Ma to 50 Ma.

377 The youngest ZHe population from 60 to 40 Ma is associated to the largest eU concentration
378 distribution (from ~0 to 4000 ppm, Fig. 3) and corresponds to the lower limit of the He-PRZ
379 (closure temperature lower or equal to 140°C, Guenthner et al., 2013). These youngest ages
380 are directly related to the main episode of cooling that affected the Mauléon basin since the
381 Eocene. This is consistent with our thermo-kinematic models that indicate a progressive
382 cooling driven by mantle subduction and thermal relaxation during the orogenic phase (Fig.
383 6). This directly led to the compensation of the hot thermal anomaly previously emplaced, as
384 plate collision and crustal thickening initiated at 50 Ma. In the absence of very-low-
385 temperature thermochronological constraints, the results of the inversion models (Fig. 4A) do
386 not lead to precise t-T scenario concerning this late phase of cooling. Whether such cooling
387 through the He-PRZ of zircons was mostly achieved early (50 - 40 Ma) and driven by thermal
388 relaxation or whether part of this cooling occurred later in the Pyrenean orogenesis (40 - 25
389 Ma) and was driven by exhumation is not precisely expressed in the models. However
390 thermo-kinematic models conducted in this study clearly demonstrate that thermal relaxation
391 during exhumation, following transient upward deflection of isotherms, represent a significant
392 cooling process that must be taken into account in the interpretation of thermochronological
393 data in this range of temperature. In the Pyrenean belt this is particularly true for the North
394 Pyrenean Zone which experienced large-scale hyper-extension related high geothermal
395 gradients.

396 Our simple approach did not allow the evaluation of the role of the fluids effect in the
397 Mauléon basin, but the good agreement between model and data suggests its role might be
398 minor at least from a thermal perspective. However, fluid flow and serpentinization of the
399 exhumed mantle in such settings may be prominent processes allowing the localization of
400 deformation during extension.

401 **7.2. Implications for the evolution of the Pyrenees**

402 After deposition, Albo-Cenomanian zircon grains were heated to a temperature of ~180°C
403 during the post-breakup evolution of the Mauléon basin. At this time, the basin was
404 presumably floored by the exhumed mantle as shown by geological evidences summarized in
405 Jammes et al. (2009): reworked granulites and mantle peridotites in Albian sediments, and
406 tectonic relationship with SCLM exhumation. These geological data best support a model A
407 hyper-extended rift basin (Fig. 6A) even if both models A and B rift basins reproduce the
408 thermal history of the basin (Fig. 6A, B). This heating phase was characterized by a

409 geothermal gradient as high as $\sim 80^{\circ}\text{C}/\text{km}$ consistent with RSCM temperatures (180°C to
410 295°C) and HT-LP metamorphism of pre-Cenomanian sedimentary units (Fig. 1).

411 Heating in the basin ceased rapidly from ~ 80 Ma on. This stage was followed by a rather
412 isothermal period that initiated coevally with the onset of plate convergence at 83 Ma. Both
413 temperature and geothermal gradient were then kept at a high level for 30 Myrs, until 50 Ma
414 when cooling/exhumation started associated with mountain building. The persistence of high
415 surface thermal flow and geothermal gradients 18 Myrs after sea-floor spreading has been
416 reported in present-day rifted margins of the Gulf of Aden (Lucazeau et al., 2010; Rolandone
417 et al., 2013). In the case of the Mauléon basin, the temperature structure acquired during the
418 rift phase prevailed at the earliest stage of continental accretion. This is in marked contrast
419 with thermal evolution reported, e.g., in Taiwan (Mesalles et al., 2014) where rapid
420 underthrusting of the lower plate (50-80 mm/yr) at onset of continental accretion led to
421 downward deflection of isotherms. This cooling phase is not detected in the early accretionary
422 prism stage of the Pyrenees. We interpret this difference as a consequence of limited lateral
423 heat advection induced by a much slower plate convergence of only 3-4 mm/yr (Mouthereau
424 et al., 2014).

425 Our result reveals that onset of shortening in the Mauléon basin occurred in an abnormally hot
426 basin. Due to the absence of significant nappe stacking in the region, we argue that ductile
427 shortening documented in the inverted rifted basin results from high temperatures inherited
428 from rifting rather than syn-convergence burial. It is characterized by axial-planar and
429 crenulation cleavages in folded Albian to Cenomanian units of the Mauléon basin that reveal
430 ambient temperatures of $100\text{-}200^{\circ}\text{C}$ (Choukroune, 1974), consistent with our models.

431 Fission track analyses on the Labourd-Ursuya Massif yield two ages at 42.2 ± 2.4 Ma and
432 48.3 ± 2.3 Ma on apatites, and an age at 81.8 ± 3.1 Ma on zircons (Yelland, 1991). Thus, ZFT
433 and ZHe ages from the Labourd-Ursuya Massif indicate initial cooling from probably deeper
434 crustal temperatures at 80-50 Ma, showing a different thermal history from the Mauléon
435 basin. AFT and AHe ages (ranging from 49 ± 4 Ma and 35 ± 3 Ma) in the Western Pyrenees
436 suggest that cooling/thermal relaxation of high temperatures after 50 Ma occurred
437 synchronously with the North Pyrenean massifs in the Eastern and Central Pyrenees (Morris
438 et al., 1998; Fitzgerald et al., 1999; Yelland, 1991), as a result of crustal thickening and
439 erosion. The thermal relaxation observed after 50 Ma in the Mauléon basin therefore appears
440 related to a major and regional exhumational phase in the Pyrenees. Because erosion is one of
441 the main agent in orogenic belts bringing heat closer to the surface it may seem
442 counterintuitive that thermal relaxation occurred during the main exhumational phase.

443 Processes other than erosion may therefore explain the thermal relaxation. Heat advection
444 recorded in the Mauléon basin remained limited first because only 2 km of basin sediments
445 were eroded since 50 Ma. In addition, our thermo-kinematic experiments (Fig. 6) show that
446 the emplacement the Mauléon basin onto a thicker and colder foreland lithosphere
447 compensates heating due to exhumation.

448 **8. Conclusions**

449 This study demonstrates that the analysis of low-temperature thermochronological constraints
450 performed on pre-/syn-rift sediments preserved in mountain belts is effective in resolving the
451 long-term post-rift and syn-convergence thermal evolution of rifted margins and hyper-
452 extended rift basins. When combined with thermal-kinematic models of rift-to-collision
453 evolution, our data allowed to test hypotheses on the thinning processes between crust and the
454 lithospheric mantle that cause the reconstructed time-temperature history.

455 Our low-temperature thermochronological data show that the sediment succession of the
456 Mauléon basin recorded a phase of heating following breakup in the Albo-Cenomanian as a
457 result of extreme extension. The Albo-Cenomanian sandstones reached temperatures of
458 180°C at only ~2 km depth, corresponding to a geothermal gradient of ~80°C/km.

459 Using this approach we demonstrate that the thermal structure of the Mauléon basin is
460 consistent with extreme thinning, although the relative thermal effect of breakup of the SCLM
461 and crustal breakup can hardly be differentiated. The temperature anomaly inherited from
462 extreme thinning lasted 30 Myrs, from ~80 Ma to ~50 Ma. This inherited thermal anomaly
463 explains ductile shortening identified in the inverted basin. It provides a mechanism for
464 explaining the observations of abnormally high temperatures (relative to inferred burial), syn-
465 convergence MT or HT metamorphism and ductile deformation in post-rift sediments. On the
466 other hand, these tectono-metamorphic characteristics are diagnostic of highly extended rift
467 basin inverted relatively soon after its emplacement. Thermal relaxation of the rift-related
468 heat anomaly occurred during the main stage of the orogenic development, when the hyper-
469 extended rift basin was thrust over the colder and thicker European plate ~50 Myrs ago.
470 The Pyrenees give us a vivid example of how high temperatures inherited from the rifting can
471 affect the thermal structure of the early stages of the collision, and how these temperatures are
472 relaxed during the late stage of orogenic processes.

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480 Appendix

481 The code of our 1D thermo-kinematic model solves the transient heat advection diffusion
482 equation (A1), including heat production in one dimension:

$$483 \quad -\frac{\partial}{\partial z} k \frac{\partial T}{\partial z} + \rho C_p v_z \frac{\partial T}{\partial z} = \rho C_p \frac{\partial T}{\partial t} + \rho H \quad (A1)$$

484 Density, ρ , heat capacity, C_p , heat production H and the heat conductivity, k , are given
485 constant values for each rock type and are listed in Table A1. The solution is obtained using a
486 standard implicit in time centered finite difference scheme at each time step. However, in
487 order to allow for advection of the 1330°C isotherm, or to allow for erosion and
488 sedimentation, the model domain is remeshed at every time step.

489 The material advection parameter is treated independently of the mesh using pre-computed
490 level-set functions that define the limit between each material phase (sediment, basement
491 crustal rocks, mantle rocks), excluding artificial diffusion of material properties with time.

492 In the models, we assume that velocity, v_z , in the rock column can be interpolated linearly
493 between each petrologic interface. Beneath the lowest interface, velocity is constant and equal
494 to the velocity of that interface. This ensures that the 1330°C isotherm imposed at the base of
495 the model is not tight to rock uplift and allows for thermal relaxation to occur. Similarly, to
496 enable effects of rock uplift or sedimentation, the 20°C isotherm is imposed at the surface of
497 the Earth, but velocity at the surface is equal to that of the shallowest rock interface.

498 Initial conditions are obtained by solving the heat diffusion equation (A2) at steady state using
499 defined material properties:

$$500 \quad -\frac{\partial}{\partial z} k \frac{\partial T}{\partial z} = \rho H \quad (A2)$$

501 This avoids artificial thermal re-equilibration, which would relate to ill-defined initial
502 geothermal gradients that would not be consistent with the material properties and particularly
503 heat production distribution.

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686 **Figure Captions**

687 Figure 1: A) Geologic map of the study area. Red stars and circles depict the position of
688 studied samples and samples from which RSCM temperatures were obtained (Clerc and
689 Lagabrielle 2014, including one new estimate from this study (Lu-1), respectively. The extent
690 of cleavage domain is shown as red dashed area. B) Synthetic lithostratigraphy of meso-
691 cenozoic successions of the Mauléon basin and layer thickness inferred from borehole data
692 (Fig. 5B). C) Geological cross-section of the western part of the Mauléon basin, including the
693 location of samples and the extent of cleavage domain, same as shown in A. Note that the
694 ductile deformation domain is observed at the base of the basin. NPFT: North Pyrenean
695 Frontal Thrust, SPT: Saint-Palais Thrust, GRH: Grand-Rieu High.

696 Figure 2: Results of thermochronological analyses and decomposition of statistically
697 representative age component of Albo-Cenomanian detrital samples. A) ZFT results from Su-

698 1, Ar-2, Ch-1, and Mi-1 samples. B) ZHe age distribution from Su-1, Ar-2, Ch-1, and Lu-1
699 samples.

700 Figure 3: The statistical distribution of our ZHe age-eU data (blue empty squares) is resolved
701 from a 2D Kernel probability density function using a Parzen window approach (Matlab code
702 available on request). Low eU content from 0 to 1100 ppm correspond to a large distribution
703 of ZHe ages from 65 Ma to 131 Ma (group B in blue). These oldest ages are associated to a
704 closure temperature of $\sim 220^{\circ}\text{C}$. By contrast, high eU content (>1100 ppm) only show young
705 ZHe ages from 36 Ma to 65 Ma (group A in red), corresponding to lower closure
706 temperatures ($<140^{\circ}\text{C}$).

707 Figure 4: A) Time-temperature histories extracted from HeFTy inverse modeling constrained
708 by ZFT data, ZHe age-eU pairs, and depositional ages, for each model (P1-A, P1-B, P2-A and
709 P2-B). B) ZHe age-eU statistical distribution of Albo-Cenomanian detrital zircon grains. We
710 compare the ZHe age-eU correlation corresponding to the best time-temperature path of each
711 model with the data.

712 Figure 5: A) Total decompacted thickness of sediments in the Mauléon basin, as obtained by
713 combining seven boreholes within or close to the Mauléon basin for maximum, minimum and
714 weighted averaged (red curve) estimations. Temporal influence intervals resolved from each
715 borehole is shown in grey. Vertical dashed lines represent the period of heating highlighted in
716 model of Figure 4 and the horizontal dashed lines correspond to the thickness of sediments
717 deposited during this period. A mean value of 2 km of sediments was deposited between 105
718 and 70 Ma. B) Map showing the location of the different boreholes.

719 Figure 6: 1D thermal-kinematic models tested for the Mauléon basin. A) Model A: crustal
720 breakup at 110 Ma and SCLM is thinned and exhumed to the base of the basin until 95 Ma.
721 Dashed red lines on model A correspond to tests considering SCLM thickening from onset of
722 convergence at 83 Ma to mature collision and exhumation after 50 Ma. B) Model B: SCLM
723 breakup occurs at 110 Ma and the continental crust thins until 95 Ma, lying in contact with the
724 asthenosphere. Tested thinning factors β_A and β_B are 4, 10, and ∞ (breakup) in both models.

725 Figure 7: Comparison of the Mauléon basin burial history with the depth of the 180°C
726 isotherm predicted from 1D rift-to-collision thermal models (A and B) shown in Figures 6A
727 and 6B, respectively. Depth evolution of the Albo-Cenomanian deposits (grey) is
728 distinguished from the Meso-Cenozoic successions (green) and water (blue). Depth of the

729 180°C isotherms produced by different thinning factors (β_A and β_B) meets the position of the
730 studied samples relatively soon after breakup of the crust (models A and A') or mantle (model
731 B) at 110 Ma. The isotherm is kept at a constant depth after onset of tectonic inversion. C:
732 Crust, S: SCLM, A: Asthenosphere.

Table A1.

Thermal and mechanical parameters considered for each type of rock in the model

	Sedimentary cover		Basement		
	Deposits	Triassic salt	Continental crust	SCLM	Asthenosphere
Thermal conductivity k W/(m.K)	2.25	6.5	2.25	3.3	3.3
Heat capacity Cp m ² /(m.s ²)	900	840	900	750	750
Heat production H μW/m ³	0.9	0	0.6	0.009	0.009
Density ρ kg/m ³	2500	2170	2800	3300	3300

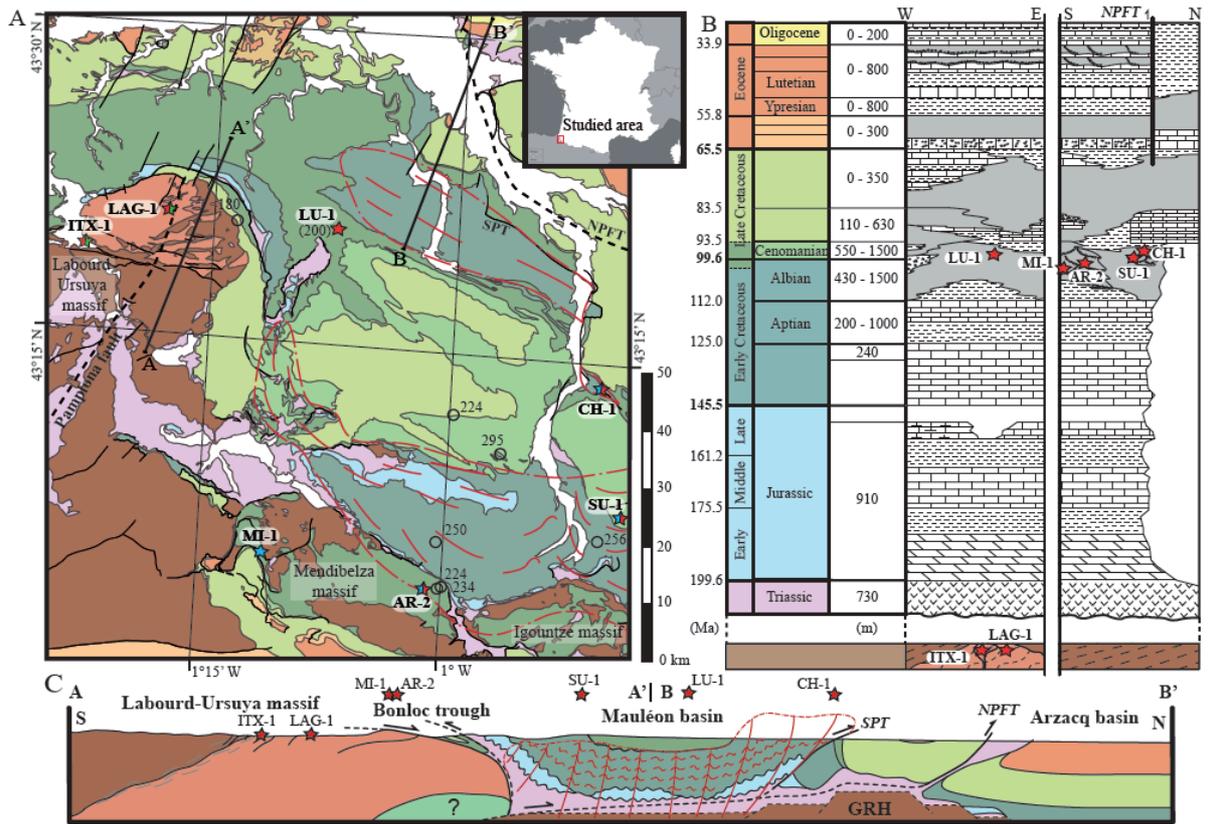


Figure 1

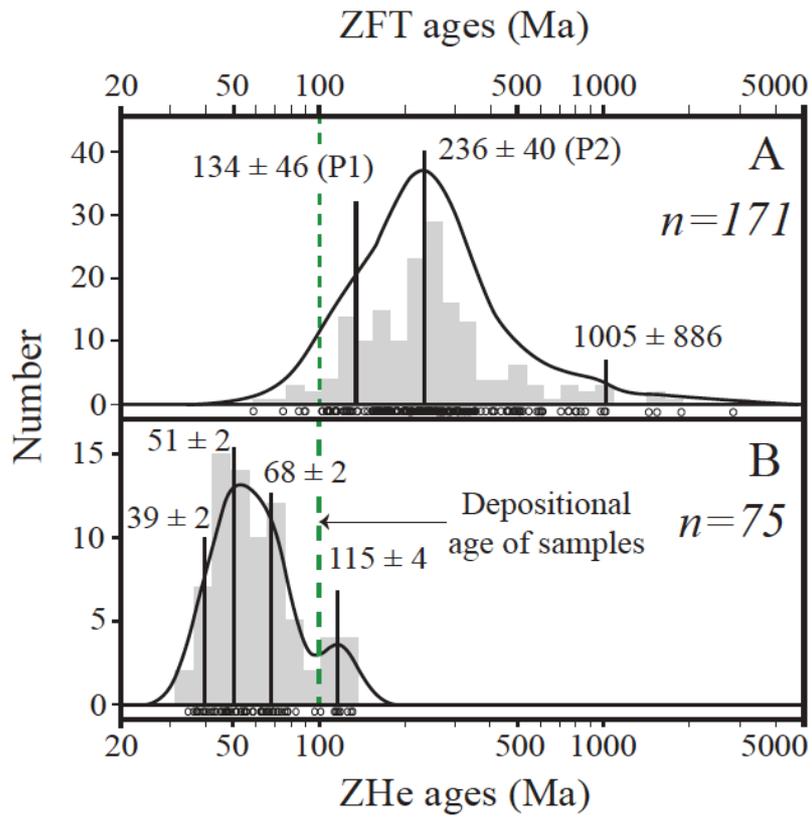


Figure 2

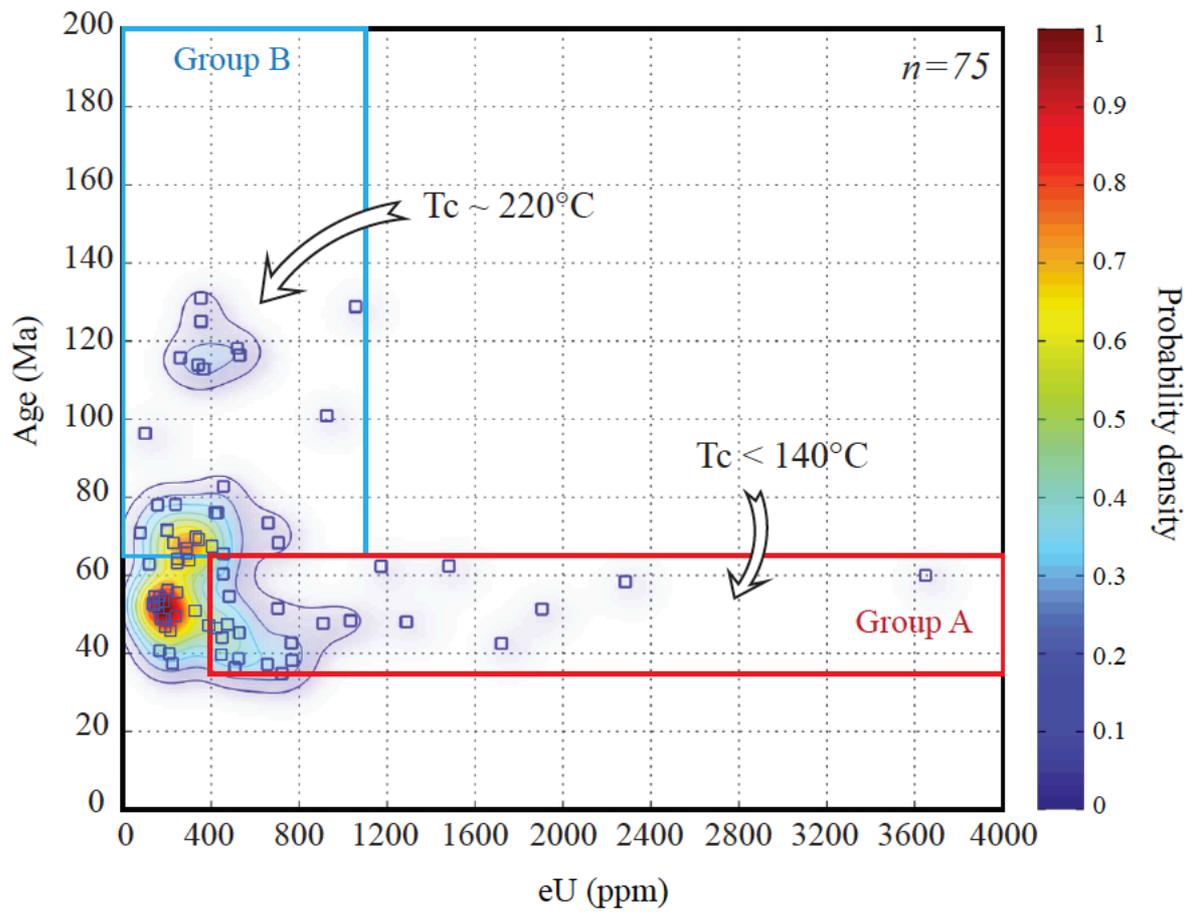


Figure 3

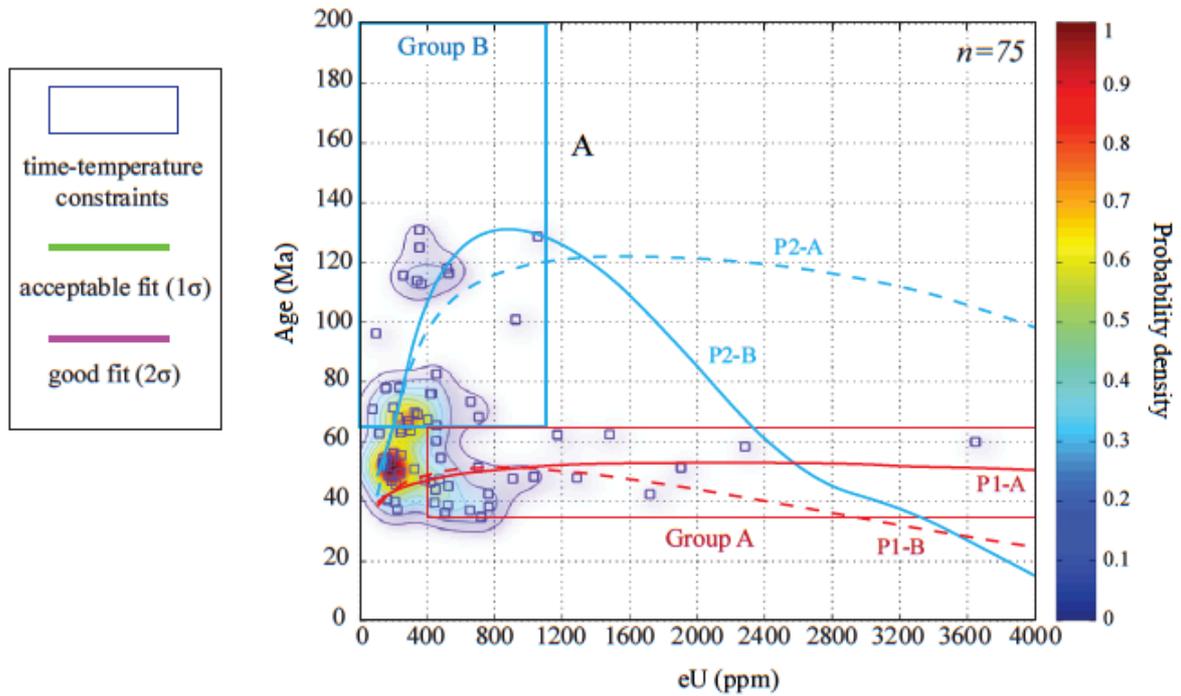
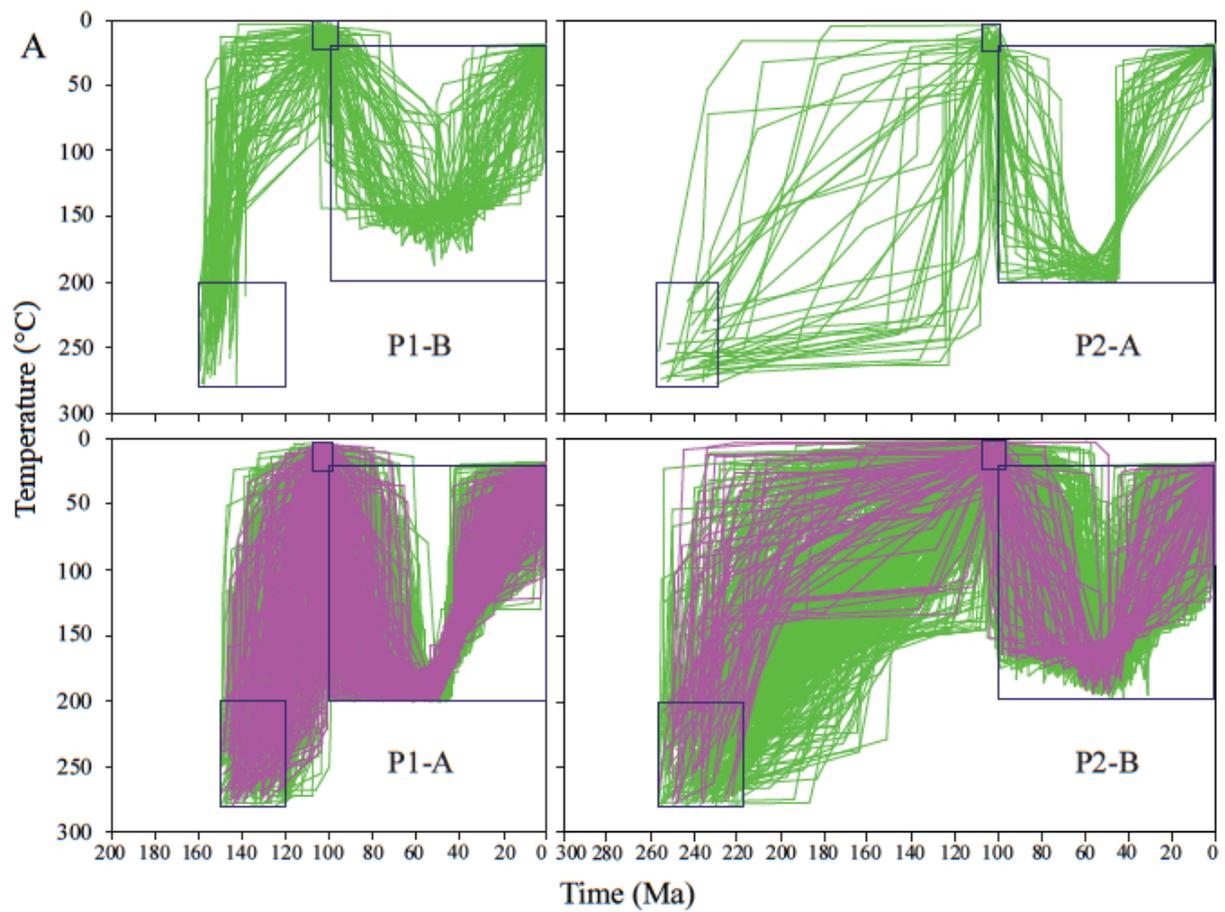


Figure 4

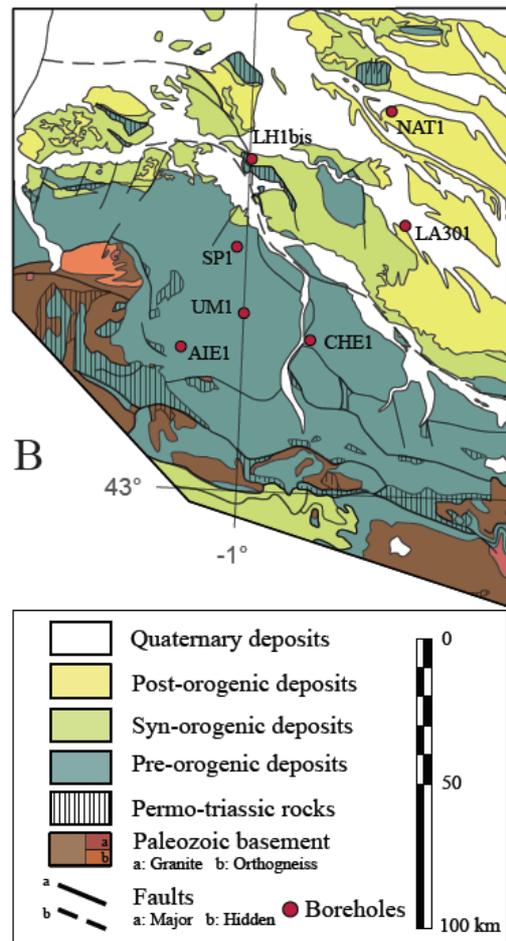
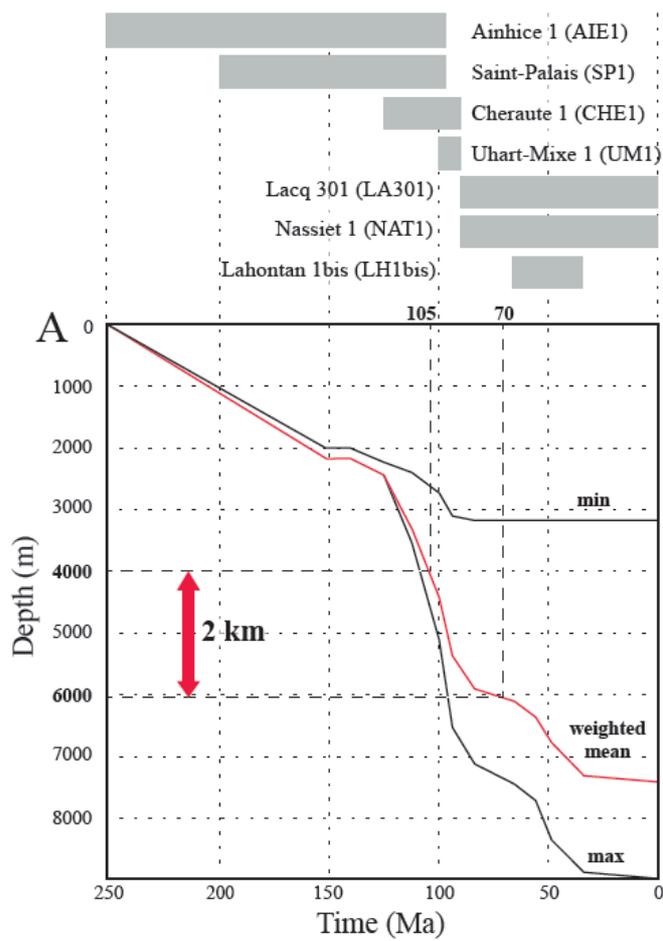


Figure 5

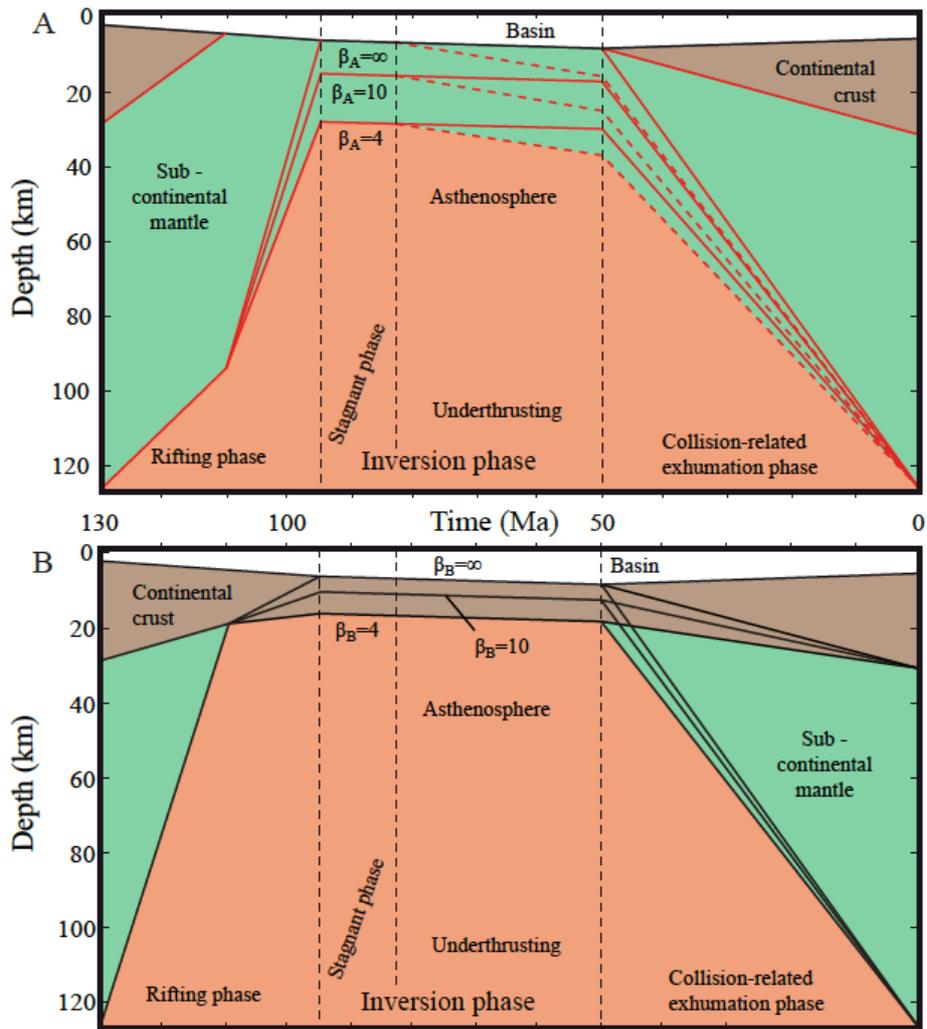


Figure 6

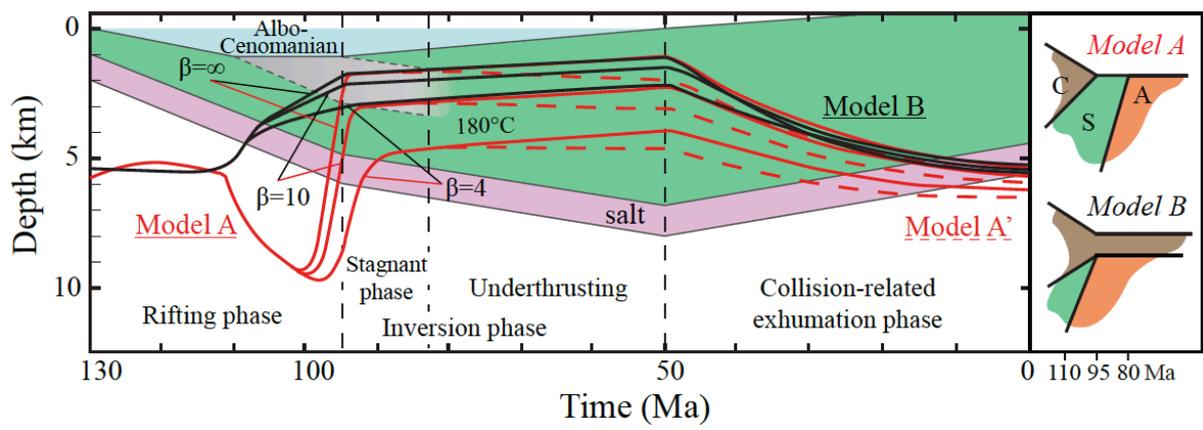


Figure 7