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1 **Exhumation history of the Variscan orogen in western Iberia**
2 **as inferred from new K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ data on granites**
3 **from Portugal**

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9

10 **Abstract**

11 Exhumation of the roots of collapsing orogens is a key process in the evolution of mountain
12 belts, which is critical for the understanding of orogenic cycles. From new structural analyses, K-Ar
13 and $^{40}\text{Ar}/^{39}\text{Ar}$ on muscovite, biotite and K-feldspar, and available U-Pb ages, we constrain the
14 cooling and exhumation history of granitic batholiths from the main structural zones of the Variscan
15 orogen in Iberia. We show that: (1) the oldest Ar dated granites (ca. 335 Ma on biotite) record
16 exhumation of gneisses crystallized much earlier (up to ca. 530-510 Ma U-Pb age); (2) granitoids
17 crystallized at ca. 330 Ma record mostly ductile stretching (L-tectonites striking around N-S),
18 consistent with N-S extension (Variscan intra-orogenic collapse); (3) most granites crystallized
19 between ca. 320 and 305 Ma record ductile stretching consistent with deformation along NW-SE to
20 ENE-WSW shear zones during the late-Variscan compression (C3); (4) granites crystallized after
21 305 Ma record mostly isotropic strain (no visible ductile foliation or lineation), consistent with their
22 emplacement during/after the final collapse of the Variscan orogeny (E2); (5) comparison between
23 Ar ages on micas and zircon U-Pb ages shows two contrasting situations: (i) similar ages, reflecting
24 crystallization and fast cooling, which can be explained by relatively shallow intrusion and/or fast
25 tectonic uplift; (ii) significant difference between U-Pb and Ar ages, supporting crystallization of
26 deeper seated intrusions and their subsequent uplift for several Ma to tens of Ma; (6)
27 thermochronological modelling on K-feldspar supports significant tectonic exhumation in the core
28 of the chain between ca. 315 and 285 Ma, followed by late-stage passive denudation; 7) Late
29 Carboniferous to early Permian tectonic exhumation was accompanied by significant formation of
30 ore-deposits. Tin deposits occur in pegmatites and/or quartz veins associated with muscovite-rich
31 syn-C3 granites and those of tungsten in quartz veins, breccia pipes and skarns associated with the
32 E2 biotite-rich granites.

33 **Keywords:** *Variscan orogeny, western Iberia; $^{40}\text{Ar}/^{39}\text{Ar}$ and K-Ar dating; $^{40}\text{Ar}/^{39}\text{Ar}$*
34 *thermochronology, Multi-diffusion domain approach (MDD), collapse stage; long-term exhumation*

35 **1. Introduction**

36 The formation of mountain belts involves episodes of granitization at various scales
37 and depths, which contribute to the incremental growth and evolution of the Earth's
38 continental crust (e.g., Hawkesworth et al., 2019). At some point during, between or after the
39 main building stages, orogens may become gravitationally unstable and may experience
40 collapse (e.g., Burg et al., 1994; Rey et al., 2001), denudation, and isostatic rebound, leading
41 to the exhumation of their granitic roots. During the last 20 years, the links between orogenic
42 collapse and crustal exhumation have been extensively investigated through geological,
43 geochronological, and modelling approaches (Vanderhaeghe and Teyssier, 2001; Jadamec et
44 al., 2007; Gerbault et al., 2018; Chardon et al., 2020 and references therein). Some studies
45 have tested the effects of variable viscosity and density of the deep crust during lithospheric
46 extension and identified contrasted modes of exhumation with uplift rates reaching km/Myr
47 (e.g., Korchinsky et al., 2018). Others have studied the influence of in-situ radiogenic heating,
48 shearing, and partial melt generation and migration, which may modify the thermal and
49 mechanical properties of the crust, favouring rapid exhumation, even during ongoing
50 compression (e.g., Gerbault et al., 2018). Nevertheless, the timing and rates of exhumation
51 associated with either collapse tectonics or late denudation by erosion remain, in many cases,
52 insufficiently documented, despite being critical to the understanding of orogenic cycles and
53 ore-deposit formation.

54 Since the early nineteen forties, geochronological methods based on the natural decay
55 of radioactive elements have been extensively developed to solve a number of problems in
56 Earth Sciences (e.g., Nier et al., 1941; Holmes, 1946; Houtermans, 1946; Roth and Poty,
57 1985). In the case of old felsic rocks such as granites, the Rb-Sr and U-Pb methods have been
58 widely used to constrain the age of magma production and crystallization at depth. U-Pb on
59 zircon is an especially powerful approach, as the refractory behaviour of this mineral hampers

60 resetting of the chronometer at high temperature, up to ca. 900 °C (e.g., Lee et al., 1997;
61 Cherniak and Watson, 2000). While monazite may be sensitive to partial U-Pb age resetting
62 in the presence of hot fluids (e.g., Tartese et al., 2011), magmatic zircons generally keep the
63 memory of various stages of crystallization. These stages can sometimes be resolved at the
64 scale of a single mineral grain through laser-ablation and high-sensitivity ion-probe in-situ
65 measurements of various crystal growths (e.g. Ireland et al., 2008). However, due to its high
66 closure temperature, zircon U-Pb dating does not allow the reconstruction of the cooling
67 history below ca. 900 °C. Therefore, U-Pb on zircon has been used extensively to constrain
68 successive stages of granite crystallization worldwide. However, it is not suitable to interpret
69 cooling associated with the late-stage collapse and exhumation of mountain belts such as the
70 Variscan orogen in western Iberia.

71 One way to constrain exhumation rates is to reconstruct the thermal history of
72 batholiths exhumed to the surface by late orogenic isostatic rebound and denudation and
73 compare the differential behaviour of the various portions of an orogen by geochronological
74 and thermochronological investigation. $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronological approaches based on
75 contrasted diffusion of argon in K-bearing minerals with variable retentivity has been
76 increasingly applied to resolve complex cooling histories (McDougall and Harrison, 1999,
77 and references therein). Variable cooling and exhumation rates have been reported for several
78 orogenic belts, e.g., the Himalayan in Southeastern Tibet (Quidelleur et al., 1997; Harrison et
79 al., 2000), the Pyrenean part of the Hercynian chain (Maurel et al., 2004), the
80 Mesoproterozoic Albany-Fraser Orogen in Western Australia (Scibiorski et al., 2015), or the
81 Eburnean Orogen in the West African Craton (Sagna et al., 2021). In most cases, these studies
82 showed fast cooling and exhumation rates associated with tectonic compression, and much
83 lower cooling rates during late denudation by erosion processes. Such $^{40}\text{Ar}/^{39}\text{Ar}$
84 thermochronological approaches therefore provide important insights into the thermal

85 evolution of the continental crust. Furthermore, they may give valuable information on how,
86 where and when ore deposits may form at depth during the typical development of an orogen,
87 which can also have significant economic impact.

88 Here we present a K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological and thermochronological
89 study on granitic batholiths from the two main structural zones of the Variscan Orogen in
90 western Iberia, the Ossa–Morena and Central Iberian zones, and one sample from a
91 parautochthonous complex in the Galicia–Trás-os-Montes Zone (Fig. 1). The combined
92 analysis of minerals with contrasted Ar retentivity and closing temperatures, coupled with
93 available U-Pb data, allows us to reconstruct the thermal and vertical displacement history of
94 the roots of the Variscan Orogen in Iberia during its late stages.

95

96 **2. Geological background**

97 The Variscan belt was produced by the convergence and collision between Laurussia
98 and Gondwana, after the closure of the Rheic Ocean. The belt is characterized by several
99 geotectonic zones with specific paleogeographic, tectonic, metamorphic and magmatic
100 characteristics (Matte, 1991; Quesada, 1991; Franke, 2000; Simancas et al., 2005; Rosas et al.,
101 2008; Murphy et al., 2016). Subduction with exhumation and retrogression to amphibolite
102 facies of earlier granulite and eclogite rocks took place at ca. 390 Ma (Dallmeyer et al., 1991).
103 It was followed by continental collision at around 370 Ma (Moita et al., 2005a, b). Three main
104 phases of contractional deformation (C1 to C3; Fig. 2) and two extensional events (E1 and
105 E2; Fig. 2) under given metamorphic conditions (M1 to M3) and igneous processes (I1 to I3)
106 are considered responsible for the structures presently observed (e.g. Martínez Catalán et al.,
107 2014).

108 The Central Iberian, the Galicia–Trás-os-Montes and the Ossa-Morena zones may

109 correspond to the most internal parts of the Variscan Belt in the Iberian Peninsula (Julivert et
110 al. 1974; Farias et al. 1987). Metamorphic conditions recorded in these zones can be
111 summarized as follows (Ribeiro et al, 2019): the allochthonous massifs of the Galicia-Trás-
112 os-Montes Zone show MP-HT and HP-HT in the Upper Allochthon, HP-LT in the Middle
113 Allochthon, and MP-MT in the Lower Allochthon; the Ossa-Morena Zone records LP-HT in
114 the Parautochthonous, MP-LT and HP-MT in the Allochthon, and HP-HT close to the contact
115 with the Central Iberian Zone, whereas LP-LT metamorphic conditions have been recorded in
116 the South Portuguese Zone. Finally, domal areas of the Central Iberian Zone show P-T paths
117 from LP-LT to MP-HT. These metamorphic conditions reflect the three main stages of the
118 Variscan cycle: (1) LP-HT associated with the Cambrian-Early Ordovician rift stage; (2) HP-
119 LT associated with subduction; and (3) syn-to-post-collisional, MP and LP/HT in the
120 hinterland and foreland fold-and-thrust belts. Thrusting of high-grade rocks to lower crustal
121 levels produced retrogression to lower amphibolite and greenschist facies. Syn- and post-
122 orogenic collapses were also responsible for retrogression to greenschist facies, especially in
123 the core of the Variscan chain where post-orogenic isostatic rebound and denudation brought
124 deep-seated rocks to the surface.

125 The Central Iberian, the Galicia-Trás-os-Montes and the Ossa-Morena zones are
126 delimited by major shear zones that have played an essential role in the late development and
127 exhumation of the Variscan Orogen in Iberia. The Coimbra-Badajoz-Cordova shear zone is a
128 WNW-ESE mega-shear, which represents an aborted continental rift during Lower Paleozoic
129 times that evolved to a sinistral Variscan flower structure in the upper Paleozoic (Ribeiro et
130 al., 1990). It represents the limit between the Central Iberian and Ossa-Morena zones. The
131 pressures and temperatures reached during the earlier Cadomian Orogeny (Upper Proterozoic)
132 and during the M1 and M2 Variscan phases (Upper Paleozoic) indicate a poly metamorphic
133 evolution of the Badajoz-Cordoba shear zone (Ábalos et al., 1991). The Juzbado-Penalva do

134 Castelo Shear Zone is an ENE-WSW sinistral shear zone (Iglesias and Ribeiro, 1981), which
135 brought into contact the Figueira de Castelo Rodrigo–Lumbrals Anatectic Complex (S-type
136 granites and migmatites) with low-grade metamorphic units from the autochthonous of the
137 Central Iberian Zone (Ferreira et al. 2019). Exhumation rates associated with non-purely
138 horizontal strike-slip along the Juzbado–Penalva do Castelo Shear Zone has been recently
139 estimated at ca. 800 m/Ma between ca. 315 and 305 Ma (Ferreira et al., 2019).

140 During the initial steps of decompression in NW Iberia, large volumes of mafic to
141 felsic magmas were generated and emplaced into Neoproterozoic to Palaeozoic sequences of
142 meta-sedimentary and meta-igneous rocks of variable metamorphic grades. Plutonic bodies,
143 based on their geological, petrographic and geochemical characteristics, have been divided
144 into two main groups (e.g., Capdevila et al., 1973; Bea et al., 1999; Ferreira et al. 1987; Dias
145 et al., 2002). A first group consists of two-mica peraluminous granites, considered S-type and
146 resulting from the crystallization of wet peraluminous magmas, whose origin at a mid-crustal
147 level is related to orogenic metamorphism. These granites formed in the core of thermal
148 domes nearly coincident with late Variscan large-scale antiforms. The second group consists
149 of biotite-rich granites (I-Type), which represent the products of complex lower-crustal
150 melting with possible mixture with mantle-derived magmas (e.g. Dias et al., 1998; Orejana et
151 al., 2009; Villaseca et al., 2011). The biotite-rich granites are distributed parallel to ductile
152 shear zones trending NNW-SSE, ENE-WSW and NNE-SSW (Ferreira et al., 1987; Dias et al.,
153 2002; Valle Aguado et al. 2005).

154 From zircon and monazite U-Pb ages obtained on samples from the Central Iberian
155 and Galicia-Trás-os-Montes zones, the two types of granites have been further divided into
156 several age subgroups. The S-type granites have been divided into three sub-groups: 331-321
157 Ma, 317-315 Ma and 311-310 Ma. The biotite-rich I-Type granites have been divided into
158 four subgroups: late-E1 (320–315 Ma), syn-C3 (310-305 Ma), and late-C3 and E2 (305-280

159 Ma) (Dias et al., 1998; Dias et al., 2002; Valle-Aguado, 2005; Neiva et al., 2012; Pereira et al.,
160 2018 and references therein). Considering I-Type magmatism from the Ossa-Morena Zone
161 represented by synkinematic gabbros, granodiorites and tonalites, and older granodiorites
162 from Galicia, we can distinguish an older, generalized episode of I-Type magmatism (early
163 I1) from ca 370 to 335 Ma (Moita et al., 2005a,b; Jesus et al., 2007; Pin et al., 2008; Pereira et
164 al., 2015; Gutiérrez-Alonso et al., 2018).

165 Several metallogenetic events associated with granites have been recognized (e.g.,
166 Neiva, 2002; Mateus and Noronha, 2010). Mostly Sn pegmatites and quartz veins with
167 cassiterite precipitated during the early stage. They are related to S-type granites and are
168 generally associated with quartz, muscovite and arsenopyrite. W (Sn) in quartz-vein deposits
169 are related to late-C3 biotite granites, and W (Mo) quartz veins and stockworks are related to
170 the E2 youngest biotite granites. Fluid inclusion and isotope data show that the biotite-rich
171 Variscan granites (310 to 280 Ma) may also play a different role in the W ore forming
172 processes, because no typical magmatic signature is found in mineralizing fluids, except in
173 the earliest stages responsible for the greisen formation. The heat flow regime supports a late
174 extensive hydrothermal activity throughout the entire crust, involving distinct fluid sources in
175 successively lower P-T conditions along a continuum that provided long-lived hydrothermal
176 systems (Mateus and Noronha, 2010; Noronha, 2017, and references therein).

177 The age of W and Sn mineralizations is still a matter of debate, due to very scarce
178 geochronological data, which might be biased by late hydrothermal events and alteration
179 processes. In the northeast Galicia-Trás-os-Montes Zone and in the northern part of the
180 Central Iberian zone, deformed Sn-pegmatites associated with cassiterite deposits in the exo-
181 contact of a syn-C3 granite were dated at 307 ± 3 Ma by Rb-Sr (Priem and den Tex, 1984),
182 whereas recent U-Pb ages on cassiterite range from 331 ± 5.6 to 310 ± 6 Ma (Zhang et al.,
183 2019). U-Pb ages on minerals of the columbite-tantalite group from pegmatites in the exo-

184 contact of a late-C3 biotite-muscovite granite yielded ages of 301 ± 4 and 316 ± 9 Ma (Lima
185 et al., 2013). $^{40}\text{Ar}/^{39}\text{Ar}$ ages on muscovite from similar non-deformed veins of pegmatite and
186 quartz including lithium minerals and fine cassiterite near the Portugal-Spain border recorded
187 two age groups around 305 and 295 Ma (Vieira et al., 2011). Further to the northwest, the
188 hydrothermal systems of W (Mo) at Borralha record post-magmatic cooling associated with
189 the youngest E2 biotite granites in the Galicia-Trás-os-Montes Zone. An $^{40}\text{Ar}/^{39}\text{Ar}$ age of
190 286.8 ± 1.2 Ma on muscovite probably reflects secondary formation from magmatic biotite or
191 feldspar alteration, while an age of 280.9 ± 1.2 Ma in sericite may date the age of greisen
192 alteration ($T \sim 300$ °C). Ages of 281.3 ± 1.2 and 277.3 ± 1.3 Ma were obtained on muscovite
193 (selvage) crystallized adjacent to molybdenite rich veins (Bobos et al., 2019), in very good
194 agreement with the previous zircon U–Pb age of 280 ± 5 Ma for the Gerês E2 biotite granite.
195 A younger K–Ar age of 273 ± 1.1 Ma was obtained on feldspar resulting from hydrothermal
196 alteration of the Gerês granite (Jaques et al., 2016).

197 At Panasqueira (southern part of the Central Iberian Zone) the largest W (Sn) deposit
198 in the Iberian Peninsula is spatially associated with a greisen cupola. $^{40}\text{Ar}/^{39}\text{Ar}$ ages on
199 muscovite from quartz veins and greisen constrain fluid circulation between 296.3 ± 0.8 and
200 291.6 ± 0.8 Ma (Snee et al., 1988). The earliest hydrothermal event (tourmalinization) was
201 recently dated at 305.2 ± 5.7 Ma by U–Pb on rutile (Carocci et al., 2019) and at 303.0 ± 3.3
202 and 301 ± 4.2 Ma by U–Pb on cassiterite (Zhang et al., 2019). In addition, a U–Pb age of 296.3
203 ± 4.2 Ma was obtained on apatite from the Panasqueira greisen, which is similar to a U–Pb
204 age of 294.5 ± 5.3 Ma obtained on apatite from the mineralized veins (Launay, 2018). All
205 these data are seemingly contradictory. This may be the result of the complex and long history
206 typical of a large deposit that implies successive reopening of mineralized veins.

207

208 3. Methods

209 3.1. Sampling strategy and description

210 Forty-five granitic outcrops in Central and Northern Portugal were studied and
211 sampled, in order to measure the visible strain and assess the timing of pluton emplacement,
212 the age of the main deformation episodes, and the exhumation of the roots of the Variscan
213 orogen in western Iberia. Deformation markers (faults, foliation, mineral lineation) were
214 measured as trends, providing a qualitative estimation of strain and its dominant orientation,
215 as reported in Table 1. Sampling initiated in the south near the city of Évora and progressed
216 northwards along a SSE-NNW transect across the Ossa-Morena and Central Iberian zones,
217 ending in the Galicia-Trás-os-Montes Zone. Only twelve out of the forty-five samples
218 collected were dated. In each zone, representative samples were selected for argon
219 geochronological and thermochronological studies based on their freshness and location in the
220 orogen, as well as type and degree of strain (Table 1). The analysed samples are reported on a
221 synthetic map showing the main Variscan and pre-Variscan chrono-stratigraphic units (Fig. 1).
222 For each sample, a detailed description of the geological unit such as given in the Geological
223 Map of Portugal scaled at 1/500 000 (LNEG, 1992) is also provided (Table 1, Supplement 1).

224 The five samples collected in the Ossa-Morena Zone (Fig.1) include pre-, syn- and
225 post-orogenic granitoids (*sl*), ranging from highly deformed to un-deformed. Near Évora, we
226 sampled a pervasively foliated, biotite-rich granitoid (sample 107D) referred to as
227 “Proterozoic rocks partially migmatized during the Variscan Orogeny” (Table 1). Slightly
228 further East, near Vendinha, we collected a less-deformed granite (sample 107G), described
229 as a “syn-orogenic tonalite” (Fig. 1, Table 1). On the geological map (LNEG, 1992), both
230 outcrops are part of an elliptical large body roughly elongated NW-SE, which may have (re-)
231 crystallized during a main deformation stage of the Variscan orogeny. Further north, sample

232 107J, collected near Pavia village, is from a weakly deformed two-micas porphyroid granite.
233 It is mapped as cutting Variscan tonalites and should therefore be younger; however, the
234 contact is unclear and seems to be made up of small faults. In any case, this granite is also
235 mapped as “orogenic” in the geological map (LNEG, 1992, Table 1). Further north, near
236 Campo Maior, another “orogenic” plutonic body mapped as anorthosite gabbro (though
237 containing biotite) was also collected (sample 107P). Finally, an undeformed porphyroid
238 granite cutting previous units, referred to as late-Variscan in the geological map (LNEG,
239 1992), was collected close to the northern boundary of the Ossa-Morena Zone near the
240 Monforte village (sample 107K).

241 The six studied Central Iberian Zone granites also range from deformed to
242 undeformed. Sample 107T is from an undeformed granite collected near the Alpalhão village,
243 close to the boundary between the Central Iberian and Ossa-Morena zones (Fig. 1). Sample
244 107X is from an isolated biotite-bearing quartz-diorite batholith near the Fundão town,
245 considered either as a “syn-tectonic” granite (LNEG, 1992), or a pre-Variscan Ordovician
246 granite and tonalite from zircon U-Pb ages around 478 ± 2 Ma reported more recently by
247 Antunes et al. (2012). A few tens of kilometres further to the NE, near the border with Spain,
248 we sampled a biotite granite with minor muscovite (sample 107AC), which is cut by E-W to
249 NE-SW normal faults (Table 1), thus pointing to a late tectonic (regional?) event with a N-S
250 to NW-SE extensional component. We note that the NE direction is also marked by numerous
251 felsic dykes in the area (Table 1). A few km to the north, deformed granites and gneisses are
252 also elongated along the ENE-WSW direction, but they are mapped as emplaced during early
253 deformation events. We sampled a lenticular gneiss with minor muscovite (sample 107AH2),
254 and a so-called “syn-tectonic granite” (sample 107AJ). The latter is affected by two distinct
255 vertical foliation planes (S-C mylonite), one oriented N135 (S-type) and the other N090 to
256 N100 (C-type), with a 15-N100 stretching lineation and shear criteria indicating left-lateral

257 kinematics (Table 1). In contrast, many structures with a dominant NW-SE orientation occur
258 further west. These include a NW-SE elongated deformed granite (sample 107AN) emplaced
259 during E1 stage, which we collected at Lamego town. The granite is cut by a major NW-SE
260 fault zone that can be followed on the geological map for several tens of kilometres (Fig. 1).

261 Finally, on the opposite block of this fault, we collected an undeformed porphyroid
262 monzonite granite (sample 107AS) near Guimarães city. This sample does not belong to the
263 Central Iberian Zone *stricto sensu*, it is a granite intruded into the Galicia-Trás-os-Montes
264 Zone.

265

266 3.2. *K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating*

267 Careful examination of thin sections (Supplementary Fig. 1) was conducted to
268 constrain the main petrographic characteristics of the samples and ensure that they did not
269 undergo significant weathering, which is critical for subsequent reliable argon dating. All the
270 samples are granitoids s.l. (granodiorites, fine-grained granites, porphyroid granites, aplites
271 dykes, orthogneisses), with structures and textures ranging from equant, isotropic and
272 undeformed to the naked eye, to strongly foliated (S-tectonite) or conspicuously lineated (L-
273 tectonite). The mineralogy is dominated by K-feldspar \pm quartz \pm mica (biotite, or muscovite,
274 or both), and includes some accessory minerals such as apatite, titanite, oxides and
275 occasionally tourmaline.

276 After crushing and sieving to a grain size in the interval 250 - 500 μm , the various K-
277 bearing mineral phases were extracted by means of magnetic separator and heavy liquids.
278 Whenever possible, either two or three distinct mineral phases were selected for a given
279 sample. For two-mica granites (107J, 107T and 107AC), especially, muscovite, biotite and K-

280 feldspar were extracted. Such separation is essential to check for internal consistency within
281 the sample and evaluate whether the ages measured may reflect crystallization age or cooling,
282 as these various phases have different Ar retentivity with respect to temperature. As such,
283 they can be used to carry out a thermo-chronological study aimed at constraining the path of
284 the samples through the different isotherms, and thus discuss the thermal history in terms of
285 exhumation and/or potential partial resetting during thermal events.

286 The K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ analyses were carried out at the GEOPS Laboratory in Orsay
287 (France). Details on the K-Ar analytical procedure and its applicability can be found
288 elsewhere (Gillot et al., 2006). For each mineral phase, K was measured at least twice by
289 flame absorption photometry, and compared with standards MDO-G (Gillot et al., 1992) and
290 BCR2 (Rackzek et al., 2001). The K-value is the arithmetic average of independent
291 measurements and is considered final when the relative standard deviation (1σ) is lower than
292 1%. The $^{40}\text{Ar}/^{36}\text{Ar}$ isotopic composition was measured on a 180° sector multi-collector mass-
293 spectrometer similar to the one described in Cassagnol and Gillot (1982) and Gillot and
294 Cornette (1986). At least two independent Ar measurements were completed on both biotite
295 and muscovite and the final K-Ar age is calculated as the weighted-mean of individual
296 determinations consistent at the 1σ level. One Ar determination was conducted in K-feldspar
297 of the samples for preliminary purposes, as K-Ar ages on such phases generally provide
298 inconclusive information regarding granite crystallization age at depth due to their poor Ar
299 retentivity at high temperature (e.g. Sagna et al., 2017). However, K-feldspar can be used to
300 achieve a thermochronological analysis through step-heating $^{40}\text{Ar}/^{39}\text{Ar}$ (see next section, and
301 Lovera et al., 1997; Sagna et al., 2021).

302 The $^{40}\text{Ar}/^{39}\text{Ar}$ analyses were performed using the multi-collector mass-spectrometer
303 and procedure described by Coulié et al. (2004). Aliquots of the samples (30 to 80 mg) were
304 first irradiated along with fluence monitors during 60 h in the Corvallis Triga reactor

305 (Radiation Center, Oregon state University, USA), using the CLICIT facility. Gas extraction
306 of the samples was completed by step heating with a high frequency furnace. A Nd-YAG
307 laser (1064 nm) was used to fuse 20 mg aliquots of the Heidelberg biotite international
308 standard HD-B1 (Fuhrmann et al., 1987), in order to determine the J factors based on the age
309 of 24.18 Ma for HD-B1 (Schwartz and Trieloff, 2007). The mass discrimination was
310 corrected using a linear law applied to analyses of air aliquots measured with the same ^{40}Ar
311 signal intensity as that measured for a typical step. The K isotopic ratio and ^{40}K decay
312 constants of Steiger and Jäger (1977) were used. It has been demonstrated that $^{40}\text{Ar}/^{39}\text{Ar}$ ages
313 with 0.1% of analytical relative precision can be reached with our instrument (e.g. Ricci et al.,
314 2013). However, as $^{40}\text{Ar}/^{39}\text{Ar}$ dating is relative to the flux monitor age, this represents the
315 main limiting factor for the accuracy of this technique (Kuiper et al., 2008). It should be noted
316 that differences of about 1% exist for the admitted ages of commonly used fluence monitors
317 such as Fish Canyon Tuff Sanidine and HDB-1. These differences can be explained by inter-
318 laboratory biases, or by the presence of previously undetected ^{39}Ar recoil (Phillips et al.,
319 2017). Therefore, the $^{40}\text{Ar}/^{39}\text{Ar}$ total age uncertainties from the present study include a
320 systematic error of 1% for the J-Factor determination.

321

322 3.3. $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology

323 The multi-diffusion domain approach (MDD) was used here to reconstruct the thermal
324 history recorded in K-feldspars of key samples (107J, T, X, AC and AS). The MDD approach
325 relies on the hypothesis that diffusion of Ar occurs similarly in nature as in the laboratory and
326 that diffusion of Ar in K-feldspar can be modelled by a discrete distribution of non-interacting
327 domains (Lovera et al., 1989; Lovera, 1992). Using the laboratory $^{40}\text{Ar}/^{39}\text{Ar}$ step-heating
328 measurements, diffusion parameters (i.e. activation energy and frequency factor) can be

329 recovered from the Arrhenius diagram built using the degassing systematics of reactor-
330 produced $^{39}\text{Ar}_K$ (Lovera et al., 1989). In the case of our samples, the first degassing steps did
331 not allow a precise determination of the activation energy (see Supplementary Figures 1 and
332 2) and hence the value of $46 \pm 6 \text{ kcal.mol}^{-1}$ typically used for K-feldspar was imposed
333 (Lovera et al., 1997). Monotonic cooling histories can be computed using inverse modelling,
334 hence without an a-priori thermal history. On the other hand, a forward modelling approach
335 with an input thermal history is necessary when post emplacement re-heating is considered. In
336 our case, forward modelling was achieved using constraints obtained from biotite and
337 muscovite (closure temperatures of $300 \text{ }^\circ\text{C}$ and $350 \text{ }^\circ\text{C}$, respectively; McDougall and Harrison,
338 1999), and available U-Pb ages on zircons ($T > 650 \text{ }^\circ\text{C}$). The cooling history that best
339 reproduces the degassing age spectra is shown for each sample (Supplementary Figure 3).
340 Finally, we tested the potential effects of a thermal event, by imposing moderate re-heating
341 (RH) below a temperature of $300 \text{ }^\circ\text{C}$ (closure temperature of biotite) at various epochs (200
342 Ma, 240 Ma, and 260 Ma, Supplementary Fig. 4) for our best-characterized sample (107AC).

343

344 **4. Results**

345 4.1. K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating

346 The new K-Ar ages (Table 2) range from 264 ± 4 to 335 ± 3 Ma. For samples with 3
347 mineral phases analysed (J, T, AC), K-Ar ages on biotite and muscovite are mutually
348 indistinguishable within the range of uncertainties. Preliminary ages on K-feldspar for these
349 samples are, however, systematically younger by more than 20 Ma, and up to 30 Ma (Table 2).
350 Such discrepancy is significant, as it represents a relative offset of ca 10%. Sample 107X, for
351 which only two phases were available, shows a similar behaviour, with K-Ar ages of 288 ± 4
352 and 264 ± 4 Ma obtained on biotite and K-feldspar, respectively.

353 $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra obtained during incremental step-heating on biotite and
354 muscovite produced relatively well-defined plateaus (Fig. 3), with more than 50% of ^{39}Ar gas
355 released and a minimum of three steps, all steps included in the plateau being consistent at the
356 2σ level (e.g., Fleck et al., 1977; McDougall and Harrison, 1999). We note, however, that
357 biotites from samples 107J and 107X do not meet these criteria, due to the presence of steps
358 with older apparent ages at relatively high temperatures, which may reflect the contribution of
359 inclusions of incompletely reset accessory minerals, e.g. small oxide grains. Such minerals,
360 along with apatite, frequently occur in the samples (as observed in thin section). However,
361 such effect appears limited, as revealed by the overall flat shape of the spectra. Similar outlier
362 steps also occur for biotites from samples 107T, 107AC and 107AN, but in these cases they
363 do not preclude the definition of a plateau. The effect is negligible as plateau ages and
364 integrated ages for these samples are fully consistent. In other words, removing the outliers
365 would define a more robust plateau, i.e. calculated from a higher ^{39}Ar content, but with
366 negligible effect on the total gas age. The spectra on muscovite are even flatter and provide
367 similar plateau and integrated ages (Fig. 3). They also yield comparable results with the
368 $^{40}\text{Ar}/^{39}\text{Ar}$ obtained on biotite of the same samples and are fully compatible with the K-Ar ages
369 (Fig. 4).

370 In contrast, the spectra obtained during incremental heating of K-feldspar (Fig. 5)
371 from the various samples show a clear stairway shape, typical of either a long gradual cooling
372 history or thermal disturbance (re-heating; McDougall and Harrison, 1999).

373

374 4.2. $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology (MDD modelling)

375 Inverse modelling (monotonic cooling) and forward modelling give similar results for
376 the feldspars of samples 107AC and 107AS (Fig. 6 and Supplementary Figs. 2 and 3). This

377 confirms the relevance of the value of 46 Kcal/mol used for activation energy (Lovera et al.,
378 1997) for all our samples. The cooling histories (300-250 Ma) imposed for forward MDD
379 modelling that best fit $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra for feldspars from our samples 107J, 107X,
380 107AC, and 107AS (see Supplementary Figure 4) are shown on Figure 7. While sample AS
381 records a relatively early and fast temperature drop after 300 Ma, samples J, X and AC show
382 a much longer cooling history. The latter two samples especially, which belong to the Central
383 Iberian Zone, record temperatures exceeding 200 °C until ca. 250 Ma.

384 Results of our test regarding a possible re-heating event are shown on Supplementary
385 Figures 5. The first two tests (re-heating at 200 Ma and 240 Ma) fail to reproduce any of the
386 apparent ages of the degassing spectra. Conversely, a re-heating of about 275 °C at 260 Ma
387 (blue line in Supplementary Figure 5C) fits most of the spectra relatively well, but is not
388 substantially better than the simplest hypothesis of monotonic cooling since 300 Ma.

389

390 **5. Discussion**

391 Overall, the new $^{40}\text{Ar}/^{39}\text{Ar}$ ages (Table 3 and Figs. 5-6) confirm the validity of the new
392 K-Ar ages and show that none of our samples have undergone significant thermal disturbance
393 above the closure temperature of biotite (i.e. ca. 300 °C). By extrapolation, this also supports
394 the validity of the K-Ar ages on biotite for samples not analysed with $^{40}\text{Ar}/^{39}\text{Ar}$, especially
395 107D and 107G (see also comparison with available U-Pb ages in Supplement 1). In contrast,
396 the abnormally young K-Ar ages systematically obtained on K-feldspars from our samples,
397 and the stairway shape of the degassing $^{40}\text{Ar}/^{39}\text{Ar}$ spectra (Fig. 5), both reflect incomplete Ar
398 retention. Our MDD modelling test for re-heating does not support significant re-opening of
399 the Ar system due to re-heating (Supplementary Fig. 5), ruling out significant contribution
400 from large, generalized thermal events, such as the plume-related emplacement of the Central

401 Atlantic Magmatic Province (CAMP) ca. 200 Ma ago (e.g., Marzoli et al., 1999; Nomade et
402 al., 2007). This is consistent with recent apatite fission-track data obtained on two of our
403 samples (107T and AC), which show no detectable temperature increase beyond 60-110 °C
404 during the last 250 Ma (Barbarand et al., 2021). Therefore, the thermal histories recorded by
405 the various mineral phases of our various samples (muscovite, biotite, and K-feldspar) most
406 likely reflect monotonic cooling at various rates in relation to granite emplacement and
407 exhumation.

408

409 *5.1. Syn- and post-deformation episodes of granite production: where and when?*

410 Comparison of the new K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages on biotite and muscovite with
411 available U-Pb data on zircon inferred to represent true crystallization age of the granites is
412 given for each sample in Supplement 1 and shown on Fig. 8. In summary, the new K-Ar and
413 $^{40}\text{Ar}/^{39}\text{Ar}$ ages on micas are either similar to or slightly lower than existing U-Pb ages on
414 zircon. In the former case, the new ages are consistent with U-Pb crystallization ages, despite
415 the great difference in closure temperature: up to ca. 900 °C for U-Pb in zircon (Lee et al.,
416 1997), compared to ca. 350 or ca. 300 °C for Ar in muscovite or biotite, respectively
417 (McDougall and Harrison, 1999). We thus infer rapid cooling. Rapid cooling can take place in
418 two situations: granite emplacement at relatively shallow crustal level ($T < 300$ °C), or granite
419 emplacement during very fast uplift. A U-Pb age significantly older than the Ar age indicates
420 a different thermal history, which can be explained by crystallization at greater depth ($T >$
421 closure temperature of muscovite), followed by relatively slow cooling.

422 Plotting the new mica K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages on a structural map of Portugal (Fig. 9)
423 allows several main patterns to be distinguished, which are summarized as follows:

424 - At the scale of the whole investigated area, the new ages decrease globally from S to N, i.e.

425 from 335 ± 3 Ma (sample 107D) to 290 ± 4 Ma (107AJ). The youngest age (sample 107X, ca.
426 285 Ma), however, departs from the general trend.- Two main age groups occur according to
427 the two main tectonic zones. Samples in the Central Iberian Zone are constrained between
428 305 and 285 Ma, whereas samples from the Ossa-Morena and Galicia-Trás-os-Montes zones
429 are dated in the range 335-305 Ma. Regarding the Galicia- Trás-os-Montes Zone, however, a
430 sampling bias is possible, as only one sample (107AS) has been analysed here, whereas U-Pb
431 ages as young as 280 ± 5 Ma have been reported for E2 biotite granites in this area (Mendes
432 and Dias, 2004). Older granites in Galicia and northern Portugal have been dated around 347-
433 310 Ma (Martínez Catalán et al., 2014; Gutiérrez-Alonso et al., 2018), which is comparable to
434 the age of synkinematic gabbros, granodiorites, and tonalites in the Ossa-Morena Zone.
435 However, the oldest granites from Galicia and northern Portugal are associated with Variscan
436 shear zones, whereas deformed plutonic bodies in the southern Ossa-Morena Zone show
437 tholeiitic to calc-alkaline affinities associated with subduction processes (e.g., Pereira et al.,
438 2015).

439 - Within each of the zones investigated here, the oldest K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages are obtained
440 on deformed granites. In the Ossa-Morena Zone, the rocks with greater deformation occur
441 along the NW direction and are dated in the range 335-310 Ma, whereas undeformed granites
442 cross-cutting previous units (samples 107K and 107P) yield K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages closer to
443 305 Ma. In the Central Iberian Zone (and Galicia-Trás-os-Montes Zone, sample 107AS) ages
444 around 305-300 Ma are obtained on samples along major tectonic structures with a similar
445 NW-SE trend. In addition, a group of rocks with significant superimposed late deformation
446 along the NE-SW to E-W direction in the Central Iberian Zone (107X, 107AC, 107AH2) is
447 here dated around 295-285 Ma.

448 - Strongly deformed samples from the Ossa-Morena Zone (107D and 107G) show similar U-
449 Pb and K-Ar ages, supporting very rapid cooling after (syn-tectonic) crystallization. For these

450 samples, steep pervasive foliation and mineral lineation striking N160 and N170 (samples
451 107D and 107G, respectively) suggest an important event of ductile stretching ca. 335-325
452 Ma ago.

453 - Undeformed samples from both the Ossa-Morena and the Galicia-Trás-os-Montes zones
454 also show very similar U-Pb, K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages, but the lack of obvious syn-kinematic
455 (ductile) crystallization for these samples supports granite emplacement and rapid cooling at a
456 shallow level around 305 Ma.

457 - In contrast, an offset from a few Ma up to 30 Ma is observed between U-Pb and the new K-
458 Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages for our deformed samples from the Central Iberian Zone and also for
459 the sample 107J from the Ossa-Morena Zone. This suggests that the crystallization of these
460 granites constrained by U-Pb ages occurred relatively deep. Then, their subsequent cooling
461 occurred over millions of years at a slower rate, as constrained by the $^{40}\text{Ar}/^{39}\text{Ar}$ ages. This
462 may be due to the position of the sampled granites closer to the axis of the chain, where
463 important deformation and thickening of the crust occurred, leaving the corresponding rocks
464 at depth for a relatively long period of time before their exhumation by either tectonic
465 processes or late erosion and isostasy.

466

467 *5.2. Exhumation rates and processes*

468 The long-lasting thermal history recorded by some of our samples (several Ma
469 difference between U-Pb and Ar ages) is unlikely to result solely from in-situ cooling at a
470 fixed vertical position after granite emplacement/crystallization. It has been shown that a
471 batch of liquid granitic magma of a few cubic kilometres emplaced at depths around 6-8 km
472 (e.g., Vigneresse, 1999) cools down to around 300 °C on timescales of only a few tens of
473 thousands of years (e.g., Nabelek et al., 2012). Such period is far shorter than the uncertainty

474 of both U-Pb, K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages. On the other hand, a long-lasting thermal evolution
475 exceeding a few Ma more realistically records the successive positions of the granitic bodies
476 at various depths. Therefore, monotonic cooling rates can be reasonably interpreted in terms
477 of exhumation rates. For consistency with earlier works on batholiths from Portugal, we will
478 use an average geothermal gradient of 42 °C/km (Pereira et al., 2017) throughout the period
479 of interest. From the reconstructed thermal history of our 4 samples (107J, X, AC and AS, Fig.
480 7), we can draw a synthetic cartoon showing the position of the granites along a main N-S
481 cross-section passing through the Ossa-Morena, Central Iberian and Galicia-Trás-os-Montes
482 zones at decreasing ages (Fig. 10):

483 - At 300 Ma, our samples from the Ossa-Morena (107J) and Galicia-Trás-os-Montes (107AS)
484 zones were already located at shallow depths ($T < 300$ °C, $d \approx 7$ km), whereas samples from
485 the Central Iberian Zone (107X and 107AC) had just crystallized at a deeper structural level
486 ($d > 15$ km).

487 - Between 300 and 285 Ma, sample 107J experienced relatively modest cooling (2 °C/Ma)
488 and slight uplift (ca. 1 km), supporting a rather stable character of the central part of the Ossa-
489 Morena Zone, which was structured and tectonically exhumed by intra-orogenic collapse well
490 before (no evidence for significant deformation after 325 Ma). Over the same period,
491 feldspars of our three other samples (107X, AC and AS) show relatively rapid cooling (12-
492 25 °C/Ma), corresponding to relatively fast vertical exhumation (300-600 m/Ma), which
493 supports a significant influence of tectonics. In the Central Iberian Zone, rates of tectonic
494 exhumation have been recently estimated from U-Pb ages obtained on zircon and apatite in
495 deformed granites sampled just to the North of an ENE-WSW trending main sinistral shear
496 zone (Juzbado-Penalva do Castelo Shear Zone), which was active during the late-Variscan C3
497 contractional stage (Ferreira et al., 2019). From the different closure temperatures of zircon
498 and apatite (ca. 830 °C and 480 °C, respectively), the authors calculated cooling rates ranging

499 from 0 to 35 °C/Ma, which, for a geothermal gradient of 42 °C/km (Pereira et al., 2017), gives
500 an exhumation rate of up to 800 m/Ma between ca. 315 and 305 Ma. Our deformed gneissic
501 sample 107AH2, also located north of the Juzbado-Penalva do Castelo Shear Zone, yields a
502 K-Ar biotite age of 294 ± 4 Ma, for an estimated U-Pb zircon crystallization age of 313-317
503 Ma (Table 1 and Supplement 1). Considering closing temperatures of 830 °C (Ferreira et al.,
504 2019) and 300 °C for zircon and biotite, respectively, sample 107AH2 cooled at a rate of 25
505 °C/Ma over the period 315-295 Ma, corresponding to a vertical exhumation rate of 500-800
506 m/Ma. These values are similar to the rates obtained from MDD modelling on feldspars of
507 deformed samples 107X and 107AC, between 300 and 285 Ma. Altogether, these data point
508 to relatively fast cooling and exhumation associated with significant tectonic deformation in
509 the Central Iberian Zone before 285 Ma. It should be noted that samples 107AH2 and 107AC
510 are located north and south of the Juzbado-Penalva do Castelo Shear Zone, respectively (cf.
511 Fig. 9). This confirms the dominant trans-current movement in the Juzbado-Penalva do
512 Castelo Shear Zone, although with a possible small vertical movement as suggested by
513 Ferreira et al. (2019).

514 - Between 285 and 275 Ma, samples 107J and 107X experienced modest cooling and
515 exhumation (5 to 8 °C/Ma, corresponding to 1 to 2 km uplift), likely due to denudation and
516 isostatic rebound, whereas samples 107AC and 107AS do not show any significant sign of
517 cooling/uplift. This period coincides with the emplacement of late-stage (post-orogenic)
518 granites and mineralization events, which marked the end of the Variscan cycle.

519 - Between 275 and 250 Ma, most of our samples appear stable and remain close to the surface
520 ($d = 3-5$ km), except for sample 107AC, which shows slight cooling (4 °C/Ma) and uplift
521 from ca. 7 km to 4 km depth. This time lapse may mark the progressive onset of the alpine
522 cycle, but no obvious imprint is recorded by the feldspars of the four tested samples.

523

524 *5.3. Mineralization processes*

525 A full characterization of ore forming processes is beyond the scope of the present
526 paper. However, our new data, combined with previous studies (Mateus and Noronha, 2010;
527 Noronha, 2017), provide a general and comprehensive view on the timing of main
528 mineralization events and their genetic link with granite emplacement and exhumation.

529 Fluid inclusion studies in W-Sn hydrothermal vein deposits suggest that most of the
530 fluids that migrated in the upper crust at the end of the Variscan orogeny were mixed with
531 surface waters during the decrease in the P-T conditions (Boiron et al., 1996). In mineralized
532 systems, aqueous-carbonic fluids with trapping conditions of P (50 to 150 MPa) and T (300 to
533 520 °C) associated with cassiterite and wolframite are contemporaneous with a thermal peak
534 related to granite emplacement (Noronha et al., 1999; 2013). A continuous input of superficial
535 waters in the system and the decreasing of the P-T conditions yields a dilution of fluids with a
536 rise in density. The fluid-related sulphide deposition ($30 < P < 100$ MPa and $T < 370$ °C) are
537 aqueous with low salinity and assumed to result from the mixing of modified fluids with
538 meteoric fluids. Some more saline and colder fluids (brines) have been recognized in quartz
539 veins related to hydrothermal alterations associated with late-Variscan Pb mineralizations
540 (Marques de Sá et al., 2019). This cycle suggests that fluids are related to granite
541 crystallization relatively deep and that their subsequent cooling occurred at a slow rate, over
542 millions of years, justifying an important role of magmatism in the ore forming processes
543 supplying heat to drive the mineralizing system. The late exhumation can justify the input of
544 surficial fluids into the system. This situation is compatible with the case in which the U-Pb
545 age is significantly older than the Ar age, which can be explained by crystallization at greater
546 depth ($T >$ closure temperature of muscovite) followed by relatively slow cooling. The

547 emplacement of the granitic rocks and the uplift and general decompression of the Variscan
548 units were the main driving forces for fluid migration, as magmatic fluids are relatively scarce
549 in the mineralizing systems.

550 In contrast, undeformed granites with similar K-Ar, $^{40}\text{Ar}/^{39}\text{Ar}$ mica ages and U-Pb
551 zircon ages record rapid cooling. In such case, granite emplacement occurred at relatively
552 shallow crustal levels, and was responsible for contact metamorphism and skarn type breccia
553 pipes deposits (metasomatic deposits).

554

555 *5.4. Broader implications*

556 Our data complement existing studies on the Variscan Orogen and others, and provide
557 significant new insights into the complex vertical motions and exhumation mechanisms in
558 evolving orogens.

559 Several episodes of exhumation, sometimes diachronic, have been previously
560 proposed for the main portions of the Variscan Orogen in central and western Europe:

561 (1) Early tectonic exhumation associated with subduction of the hyper-extended
562 margins of Gondwana. This episode, which lasted from middle to late Devonian (see recent
563 synthesis in Vanderhaegen et al., 2020), was not investigated in the present study.

564 (2) Early to mid-Carboniferous exhumation during/after continental collision. This
565 episode is generally attributed to significant thickening of the crust and subsequent collapse of
566 the chain. It has been recorded by metamorphic and syn-orogenic magmatic rocks from
567 several Variscan massifs: in Brittany (Gumiaux et al., 2004), in the French Massif Central
568 (e.g., Gébelin et al., 2009; Chardon et al., 2020), in the western part of the Maures-Tanneron
569 Massif (Morillon et al., 2000; Gerbault et al., 2018), in the Vosges Massif (Boutin et al.,

570 1995), in the Bohemian Massif and the western Carpathians (e.g., Bues et al., 2002; Franek et
571 al., 2011; Moussalam et al., 2012; Vacek et al., 2019), and in Iberia (e.g., Rosas et al., 2008;
572 Pereira et al., 2017; Gutiérrez-Alonso, 2018; Ferreira et al., 2019; this study). Proposed
573 exhumation mechanisms include large-scale crustal folding and thrusting, transcurrent motion
574 along shear zones, upward flow of buoyant material, and horizontal spreading along crustal
575 detachments. Our new data on deformed samples from the Ossa-Morena Zone record ductile
576 stretching consistent with N-S extension and very fast uplift, here interpreted as a brief
577 episode of Variscan intra-orogenic collapse around 335-330 Ma (E1 in Fig. 2). In contrast,
578 most granites crystallized between ca. 325 and 305 Ma record ductile stretching and
579 significant uplift consistent with deformation along NW-SE to ENE-WSW shear zones during
580 late-Variscan compression (C3). This supports a significant stress change around 330-325 Ma,
581 at least in this part of Iberia, and suggests that contrasted styles of tectonic deformation
582 (extensional vs transpressional) can equally promote significant rapid uplift/exhumation of the
583 lower crust within only a few Ma. These characteristics may additionally constitute a unique
584 witness of a (diachronic?) change in deformation style from extension-dominated to
585 transpression-dominated across the orogen.

586 (3) Late Carboniferous to early Permian tectonic exhumation associated with the final
587 gravitational collapse of the Variscan Orogen (e.g., Burg et al., 1995, Marques et al., 2002).
588 This stage involved significant crustal thinning through transtensional and extensional
589 deformation along low-angle detachments. Related decompression triggered significant
590 partial melting of the lower crust (anatexis), presently witnessed by migmatite domes and
591 late-orogenic anatectic granites, often intruded by younger post-orogenic granites. These
592 various types of rocks are presently exposed in many places of the Variscan Orogen, e.g., in
593 the South Armorican Domain (Turillot et al., 2011; Ballouard et al., 2015), in the French
594 Massif Central (Chardon et al., 2020; Vanderhaeghe et al., 2020), in Italy (Ayuso et al., 1994),

595 in the Pyrenees (Maurel et al., 2004), and in Iberia (references above; this study). In most
596 instances, available U-Pb, Rb-Sr and $^{40}\text{Ar}/^{39}\text{Ar}$ ages on amphiboles and micas suggest
597 significant exhumation until at least 290 Ma (Turillot et al., 2011; Chardon et al., 2020).
598 Nevertheless, $^{40}\text{Ar}/^{39}\text{Ar}$ data on feldspars are very scarce (e.g. Maurel et al., 2004), limiting
599 our knowledge of subsequent shallow exhumation dynamics at the full scale of the orogen.
600 Our new MDD data on samples 107X and 107 AC from the Central Iberian Zone have
601 recorded significant cooling and exhumation between ca. 300 and 285 Ma. While the period
602 300-295 Ma coincides with late sinistral movement along main shear zones in the Central
603 Iberian Zone (see previous point), exhumation in the range 295-285 Ma may more
604 realistically be related to (the onset of?) transtension and extension processes associated with
605 the final collapse of the chain (E3 in Fig. 2). Such possibility is supported by the presence of
606 dyke swarms with sharp contacts (chilled margins) and brittle faults with normal-dextral
607 kinematics affecting our samples 107 X and 107AC, respectively (see Table 1).

608 (4) Late exhumation driven by either Alpine tectonics (*sl*), or denudation by erosion
609 processes. The thermal influence of the Alpine cycle related to Triassic early extension and
610 opening and closure of the Tethys has been recorded for instance in the central Serbo-
611 Macedonian Massif in the Carpathians (e.g., Antic et al., 2017). Regarding Iberia, recent
612 Apatite Fission Track (AFT) data on siliciclastic sediments filling Mesozoic basins along the
613 western margin of Iberia support significant uplift of the Newfoundland-Iberia rift shoulder
614 following the opening of the Atlantic at ca. 150 Ma (Barbarand et al., 2021). However, the
615 granites studied here are located further to the East of the margin. In fact, AFT data on our
616 samples 107T and 107AC show continuous cooling since granite emplacement. Our new
617 MDD data on K-feldspars of the four tested samples from the various structural zones support
618 very slow cooling rates (only a few $^{\circ}\text{C}/\text{Ma}$) between 275 and 250 Ma. This is consistent with
619 average cooling rates of 1 $^{\circ}\text{C}/\text{km}$ recorded by K-feldspar from the weakly deformed Mont-

620 Louis granite in the Eastern Pyrenees, though the latter rate has been estimated on a much
621 larger temporal window between 295 and 60 Ma (Maurel et al., 2004). The relatively low
622 cooling rates here obtained for the Central Iberian Zone are interpreted as evidence for slow
623 exhumation driven by passive denudation and isostatic re-adjustment immediately after the
624 collapse of the orogen.

625 To summarize, our data recorded the successive stages of exhumation associated with
626 the main steps of evolution of the Variscan belt following early continental collision. The
627 combined use of new structural data, K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages on micas, and K-feldspars
628 thermochronology, along with available U-Pb ages support a major role of the various
629 tectonic processes over a cumulative period of about 50 Myr, followed by much slower
630 evolution driven by passive denudation. Such a scenario is consistent with cooling histories
631 depicted by similar geochronological and $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronological studies on other
632 orogens. For the Cretaceous Gangdese batholith emplaced in Southeastern Tibet during
633 Tethys subduction, mean rates of about 25°C/Myr have been recorded from about 900 to
634 300 °C (Quidelleur et al., 1997), in relation to the high slip rate of 7 mm/yr modelled for the
635 Gangdese Thrust (Harrison et al., 2000). Such fast tectonic exhumation rates have also been
636 inferred for older orogens. During the Mesoproterozoic Albany-Fraser Orogen (Western
637 Australia), rapid exhumation driven by transpressional tectonic activity induced high cooling
638 rate, up to 30 °C/Myr from 585 to 365 °C (Scibiorski et al., 2015). For the Eburnean Orogen
639 in the West African Craton, Paleoproterozoic granites record relatively high cooling rates up
640 to 20 °C/Myr from 900 to 300 °C (Sagna et al., 2021). In contrast, much lower cooling rates
641 of only a few °C/Myr have been inferred for the Grenville Orogen over the 500-350 °C
642 (Reynolds et al., 1995) and 500-150 °C (Cosca et al., 1991) temperature ranges, suggesting
643 that exhumation was there mainly controlled by post-orogenic erosional unroofing.

644

645 **6. Conclusions**

646 Our combined K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological and thermochronological study of
647 granites from various portions of the Variscan orogen in western Iberia shows that:

- 648 - Granites crystallized at ca. 335 Ma in the Ossa-Morena Zone have recorded mostly
649 ductile stretching (L-tectonites striking around N-S), consistent with N-S extension
650 associated with Variscan intra-orogenic collapse. The small age difference between U-
651 Pb, K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages on these granites supports relatively fast cooling
652 associated with rapid tectonic exhumation.
- 653 - Deformed granites in the Central Iberian Zone show the highest offset between U-Pb
654 and K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages. This supports crystallization at deep level in the core of
655 the chain, followed by a long-cooling history associated with significant exhumation
656 by tectonics processes: fast tectonic exhumation between ca 315 Ma and 285 Ma, due
657 to (1) non purely-horizontal left-lateral movement along shear zones during the end of
658 C3, and (2) late extension during the final collapse of the Variscan orogen (E2).
- 659 - Youngest dated granites (crystallization <305 Ma) in both the Ossa-Morena and the
660 Central Iberian zones have recorded mostly isotropic strain (no visible ductile foliation
661 or lineation) and show no significant offset between U/Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ ages. This
662 supports shallow granite emplacement, during E2, and subsequent low-rate denudation
663 and isostatic rebound.
- 664 - Our new K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ data, along with available U-Pb, support a northward
665 propagation of tectonic exhumation from the Ossa-Morena Zone to the Central Iberian
666 Zone.
- 667 - Late Carboniferous to early Permian tectonic exhumation was accompanied by
668 significant ore-deposits formation with tin deposits occurring in pegmatites and/or
669 quartz veins associated with the muscovite-rich syn-C3 granites and those of tungsten

670 occurring in quartz veins, breccia pipes and skarns associated with the E2 biotite-rich
671 granites.

672

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683 The authors declare that they have no known competing financial interests or personal

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687 **References**

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1017 Table 1: Main characteristics of the investigated outcrops and samples. OMZ – Ossa-Morena
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1060

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1062

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1068 Figure 7: Comparison of the 4 thermal histories obtained with MDD forward modelling on K-
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1071

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1077 Shear Zone; JPCSZ: Juzbado-Penalva do Castelo Shear Zone ; CBCSZ - Coimbra-Badajoz-
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1080 Figure 10: Synthetic cartoon showing the successive vertical positions of samples J, X, AC
1081 and AS between 300 and 250 Ma, as deduced from MDD modelling. Sample 107AS is a
1082 granite intruded into the nappes of the GTMZ (after thrusting). F1: main thrust separating the
1083 South Portuguese Zone (SPZ) and the Ossa-Morena Zone (OMZ); F2: Coimbra-Badajoz-
1084 Cordoba Shear Zone (CBCSZ) separating the Ossa-Morena Zone and the Central Iberian
1085 Zone; F3: Juzbado-Penalva do Castelo Shear Zone (JPCSZ). Solid lines indicate inferred
1086 active fault movements at a given time; dotted lines indicate inactive fault movements. In
1087 each structural zone, colored arrows, show the amount of uplift between successive stages.

Sample	Longitude (°)	Latitude (°)	Zone	Symbol – description on the geological map	Amount of strain	Strain type – kinematics <i>Field data</i>	U-Pb (Ma)
107D	38.46046	-7.75290	OMZ	pe _{mg} – Proterozoic series migmatized Variscan orogeny	+++	Pervasive steep foliation strike N160	<i>577 ± 10 (1)</i> 341 ± 2 (2)
107G	38.48016	-7.64957	OMZ	t – orogenic tonalite	++	L-tectonite with lineation strike N170 Late sinistral strike-slip E-W fault	329 ± 1 (3)
107J	38.84073	-7.89542	OMZ	γ – orogenic biotite-rich granite	+	Weak foliation strike N120	323 ± 1 (3)
107K	39.08173	-7.41841	OMZ	γ ₃ – late-orogenic biotite-rich porphyroid granite	no	Isotropic	305 ± 6 (4) 301 ± 5 (5)
107P	38.98935	-7.14817	OMZ	T1 – orogenic anorthosite gabbro	no	Isotropic	305 ± 6 (4)
107T	39.41024	-7.62334	CIZ	γ ₃ – late-orogenic porphyroid monzonite granite	no	Isotropic	305 ± 6 (4)
107X	40.17919	-7.48563	CIZ	γII 2a – syn-tectonic quartzdiorite and biotite granite	no	Heavily fractured	<i>478 ± 0.5 (6)</i> 310 ± 4 (7)
107AC	40.76379	-6.85296	CIZ	Monzonite granite with sparse feldspar megacrysts; muscovite and biotite granite	no	Fractured	303 ± 7 (7)
107AH2	40.84370	-7.18323	CIZ	γ1T – orogenic augen gneiss	+++	Pervasive foliation, sub-vertical E-W ductile shear zone, sinistral kinematics	313 - 317 Ma (8,9)
107AJ	40.99665	-7.47315	CIZ	γI ₃ – syn-F3 orogenic 2-mica granite	++	Ductile, sub-vertical foliation N135 (S), N100 sub-vertical (C), lineation 15°, N100, left-lateral kinematics	319 ± 3 (10)
107AN	41.08813	-7.79960	CIZ	γIII – early orogenic biotite granodiorite	++	Ductile, foliation N070, 55°SE Dykes N015, 25°E cut by faults N45, 55°SE striation 45°, N160, normal dextral kinematics	319 ± 4 (11)
107AS	41.50952	-8.25861	GTMZ	γII 2b (γIIIc?) – syn-orogenic porphyroid granite/granodiorite	+	Faults N060 sub-vertical, sinistral strike-slip Faults N020, sub-vertical, dextral strike-slip	309 ± 1 (11)

1

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Table 2. New K-Ar results on the various mineral phases of our samples. Lat.: latitude, Long.: longitude; KFd: K-Feldspar (only one Ar determination), biot: biotite, musc: muscovite. The mean age is obtained by weighing by the amount of radiogenic argon ($^{40}\text{Ar}^*$), except in the case of 107D-biot, 107G-biot, and 107J-musc; for the latter, all three individual determinations overlap within uncertainties, and the mean age (*) is calculated by weighing by the inverse of variance. Measured ages in normal font, and mean ages in bold font. Uncertainties quoted at the 1σ level.

Sample	Longitude	Latitude	K%	$^{40}\text{Ar}^*$ (%)	$^{40}\text{Ar}^*$ (10^{15} at.g $^{-1}$)	Age (Ma)	Unc. (Ma)
107D-biot	-7.75290	38.46046	7.598	98.7	2.918	334.6	4.7
				99.2	2.943	337.2	4.8
				97.9	2.903	333.1	4.7
					mean*	335.0	2.7
107G-biot	-7.64957	38.48016	7.643	98.4	2.927	333.8	4.7
				98.6	2.850	325.8	4.6
				96.5	2.889	329.8	4.7
					mean*	329.7	2.7
107J-biot	-7.89542	38.84073	7.471	97.3	2.722	318.9	4.5
				95.7	2.619	307.8	4.4
				99.3	2.663	312.5	4.4
					mean	313.0	4.4
107J-musc	-7.89542	38.84073	8.722	97.2	3.095	311.3	4.4
				98.9	3.122	313.8	4.4
				98.7	3.102	311.9	4.4
					mean*	312.3	2.6
107K-biot	-7.41841	39.08173	6.808	98.9	2.356	304.2	4.3
				96.9	2.363	305/0	4.3
					mean	304.6	4.3
107J-KFd	-7.89542	38.84073	10.050	94.1	3.273	287.6	4.1
107P-biot	-7.14817	38.98935	7.725	98.9	2.685	305.4	4.3
				98.5	2.706	307.6	4.4
					mean	306.5	4.3
107T-KFd	-7.62334	39.41024	10.890	96.3	3.319	270.5	3.8
107T-biot	-7.62334	39.41024	7.494	97.8	2.549	299.4	4.2
				97.5	2.509	295.1	4.2
					mean	297.3	4.2

107T-musc	-7.62334	39.41024	7.928	98.0	2.714	301.2	4.3
				98.1	2.732	303.0	4.3
					mean	302.1	4.3
107X-KFd	-7.48563	40.17919	11.308	88.0	3.355	263.8	3.7
107X-biot	-7.48563	40.17919	7.692	98.9	2.492	286.3	4.1
				98.6	2.533	290.6	4.1
					mean	288.4	4.1
107AC-KFd	-6.85296	40.76379	8.819	96.1	2.683	270.0	3.8
107AC-biot	-6.85296	40.76379	7.335	98.0	2.491	299.0	4.2
				92.2	2.464	296.0	4.2
					mean	297.4	4.2
107AC-musc	-6.85296	40.76379	8.266	97.7	2.811	299.3	4.2
				95.7	2.723	290.7	4.1
				98.5	2.738	292.1	4.1
					mean	294.0	4.2
107AH2-biot	-7.18323	40.84370	7.561	97.8	2.482	289.8	4.1
				98.3	2.560	298.1	4.2
					mean	294.0	4.2
107AJ-biot	-7.47315	40.99665	7.521	99.1	2.491	292.1	4.1
				98.6	2.451	287.8	4.1
					mean	289.9	4.1
107AN-biot	-7.79960	41.08813	7.371	92.2	2.553	304.4	4.3
				98.9	2.606	310.3	4.4
				98.8	2.502	298.8	4.2
					mean	304.5	4.3
107AS-biot	-8.25861	41.50952	7.640	93.4	2.691	309.1	4.4
				97.9	2.671	307.1	4.4
					mean	308.1	4.4

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Sample	Plateau age (Ma)	n/N	(% ^{39}Ar)	MSWD	Integrated age (Ma)	K-Ar age (Ma)
107J-biot	-	-	-	-	309.2 ± 2.9	313.0 ± 4.4
107J-musc	308.7 ± 0.2 (2.8)	8/13	83.8	2.1	308.8 ± 3.0	312.2 ± 2.6
107T-biot	297.1 ± 0.2 (2.7)	4/11	52.5	1.2	295.7 ± 2.8	297.3 ± 4.2
107T-musc	297.1 ± 0.2 (2.8)	6/14	70.0	1.9	297.2 ± 2.8	302.1 ± 4.3
107AC-biot	296.4 ± 0.3 (2.7)	5/13	51.8	1.5	294.1 ± 2.8	297.4 ± 4.2
107AC-musc	294.5 ± 0.1 (2.7)	11/15	92.1	2.3	294.9 ± 2.8	294.0 ± 2.4
107X-biot	-	-	-	-	283.7 ± 2.8	288.4 ± 4.1
107AN-biot	300.4 ± 0.2 (2.8)	5/11	51.8	3.0	298.6 ± 2.8	304.5 ± 4.3
107AS - biot	-	-	-	-	304.2 ± 2.9	308.1 ± 4.4
107J - KFd	-	-	-	-	281.0 ± 2.7	287.6 ± 4.1 *
107T - KFd	-	-	-	-	272.6 ± 2.7	270.5 ± 3.8 *
107X - KFd	-	-	-	-	265.6 ± 2.6	263.8 ± 3.7 *
107AC - KFd	-	-	-	-	261.0 ± 2.6	270.0 ± 3.8 *
107AC - KFd-duplicate	-	-	-	-	261.3 ± 2.6	270.0 ± 3.8 *
107AS - KFd	-	-	-	-	292.3 ± 2.6	

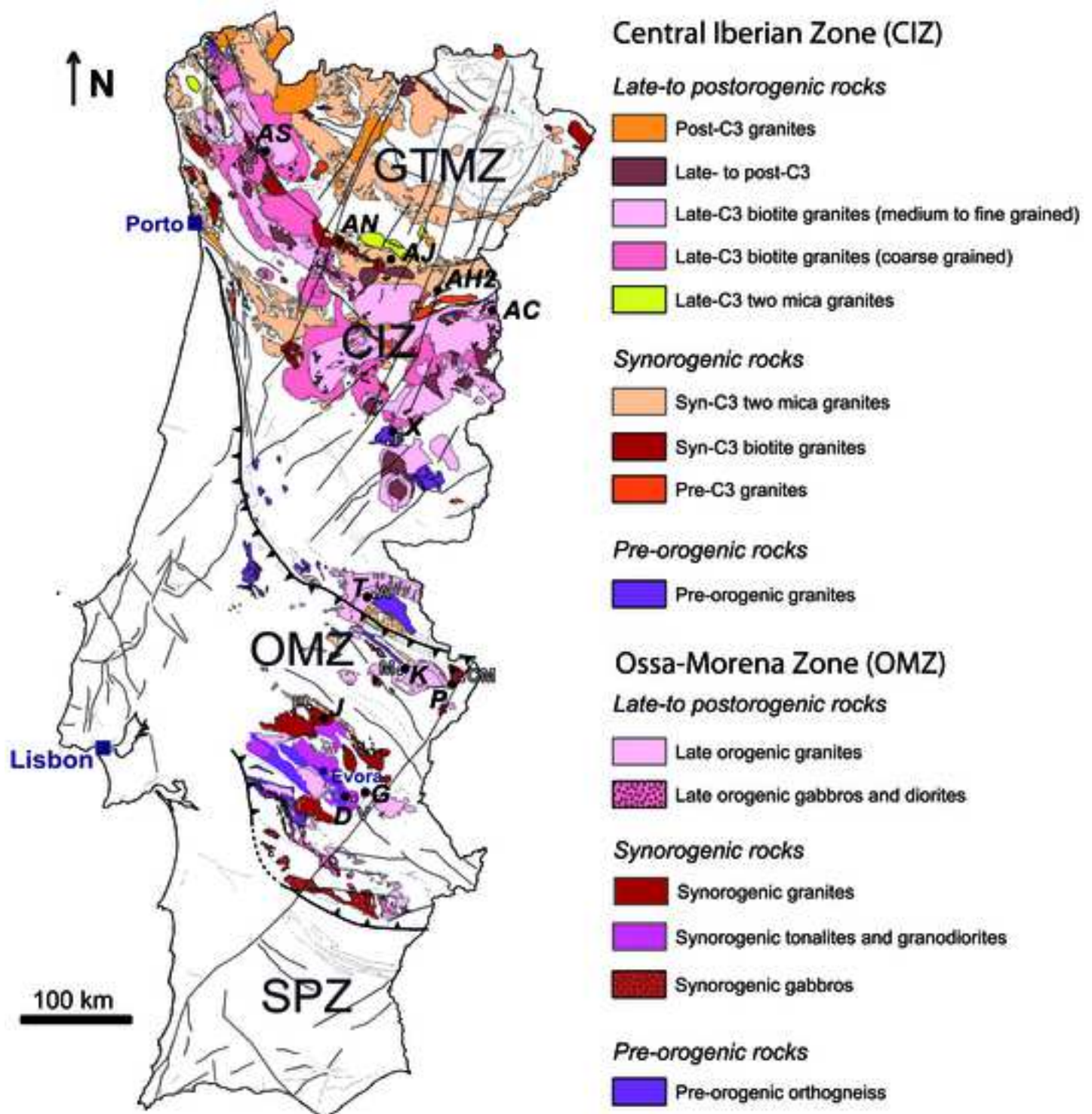


Figure 1: Simplified geological map of Portugal showing the various Variscan and pre-Variscan chrono-stratigraphic units in the main tectonic zones. The position of our dated samples is reported with black circles. Sample acronyms indicated with bold italic font. SPZ is South Portuguese Zone, and GTMZ is Galicia-Trás-os-Montes Zone. Main towns shown with dark blue solid squares; smaller towns and villages shown with white squares circled with black. A: Alpalhão, CM: Campo Maior, F: Fundão, G: Guimarães, M: Monforte, P: Pavia, V: Vendinha.

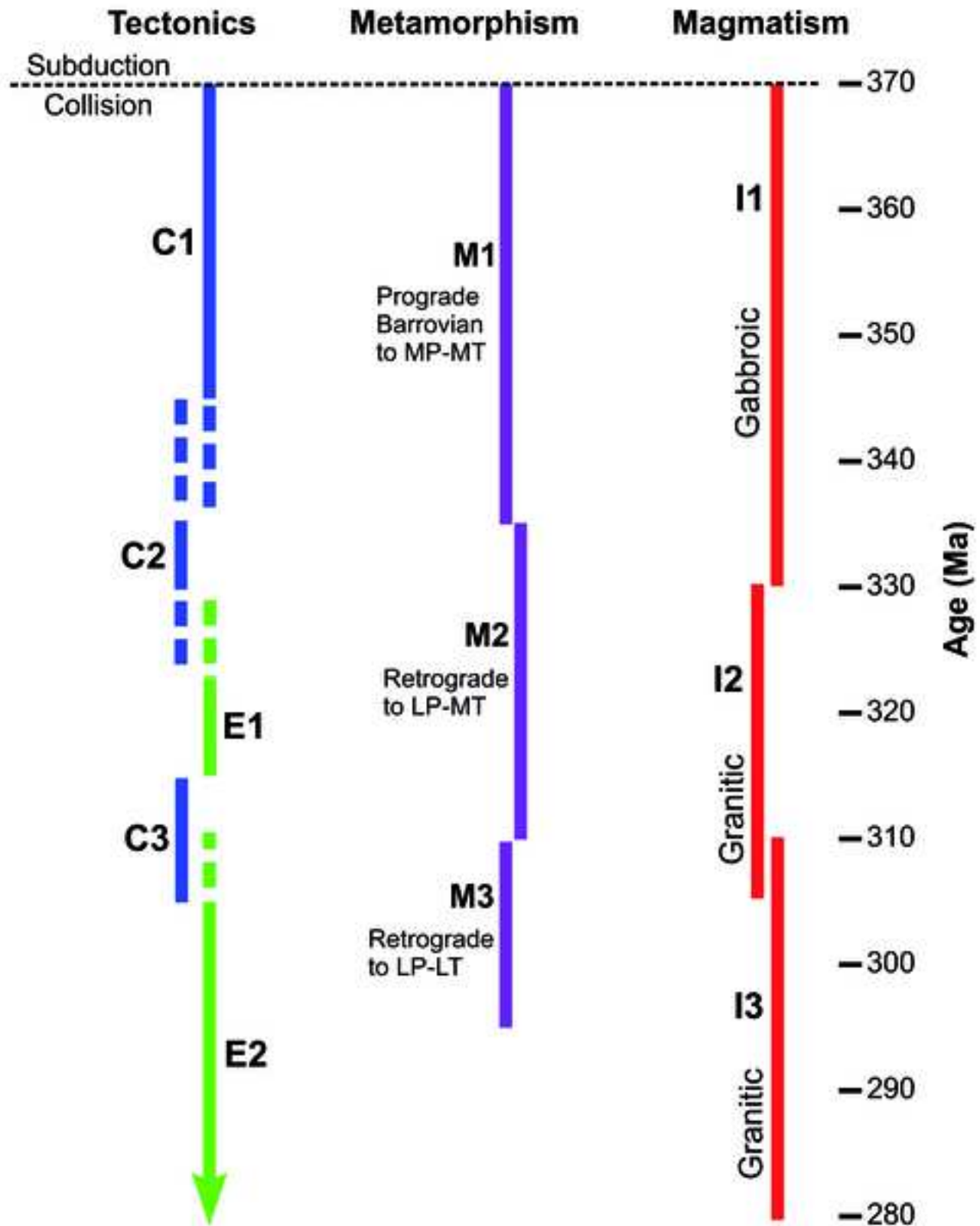


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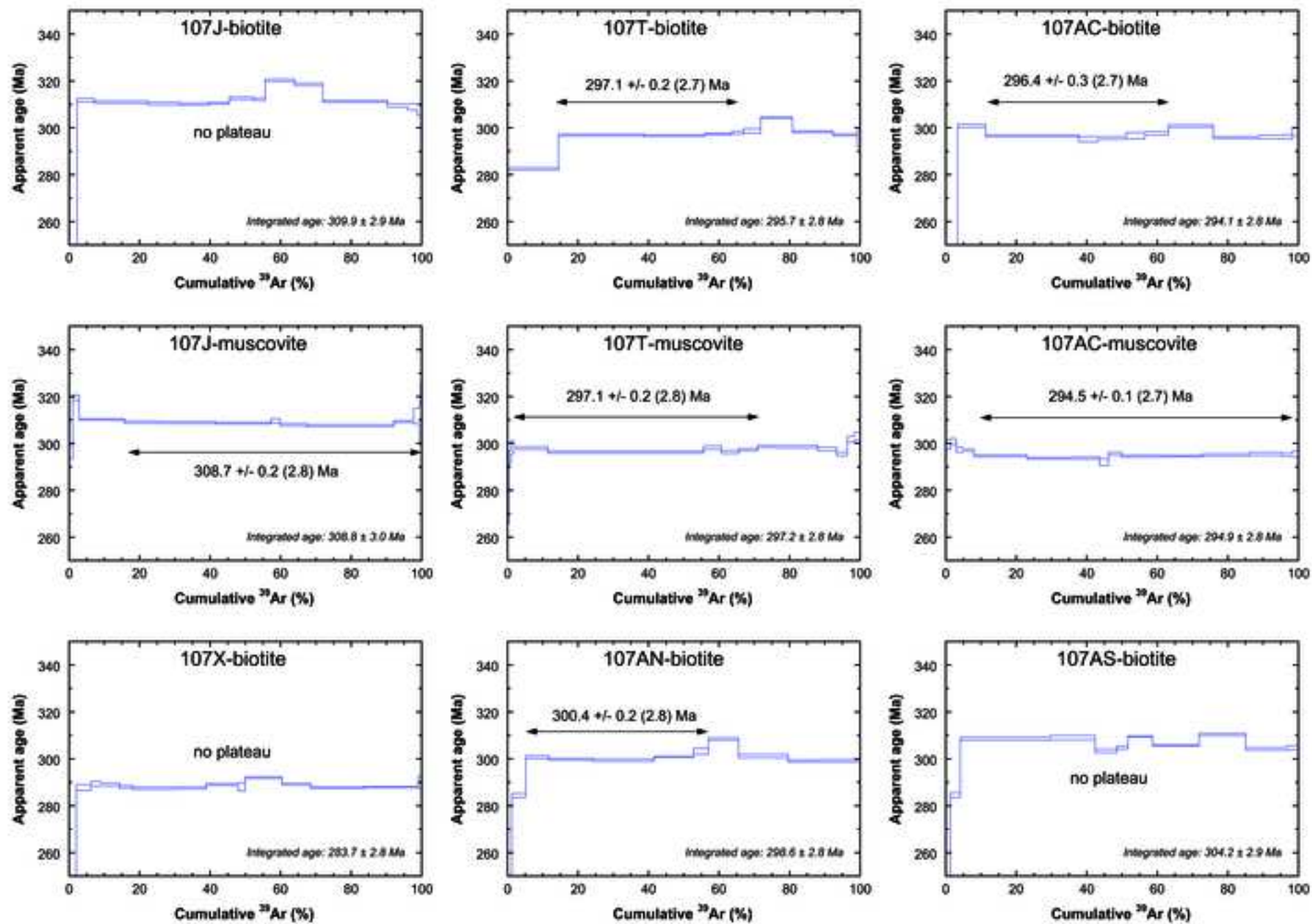


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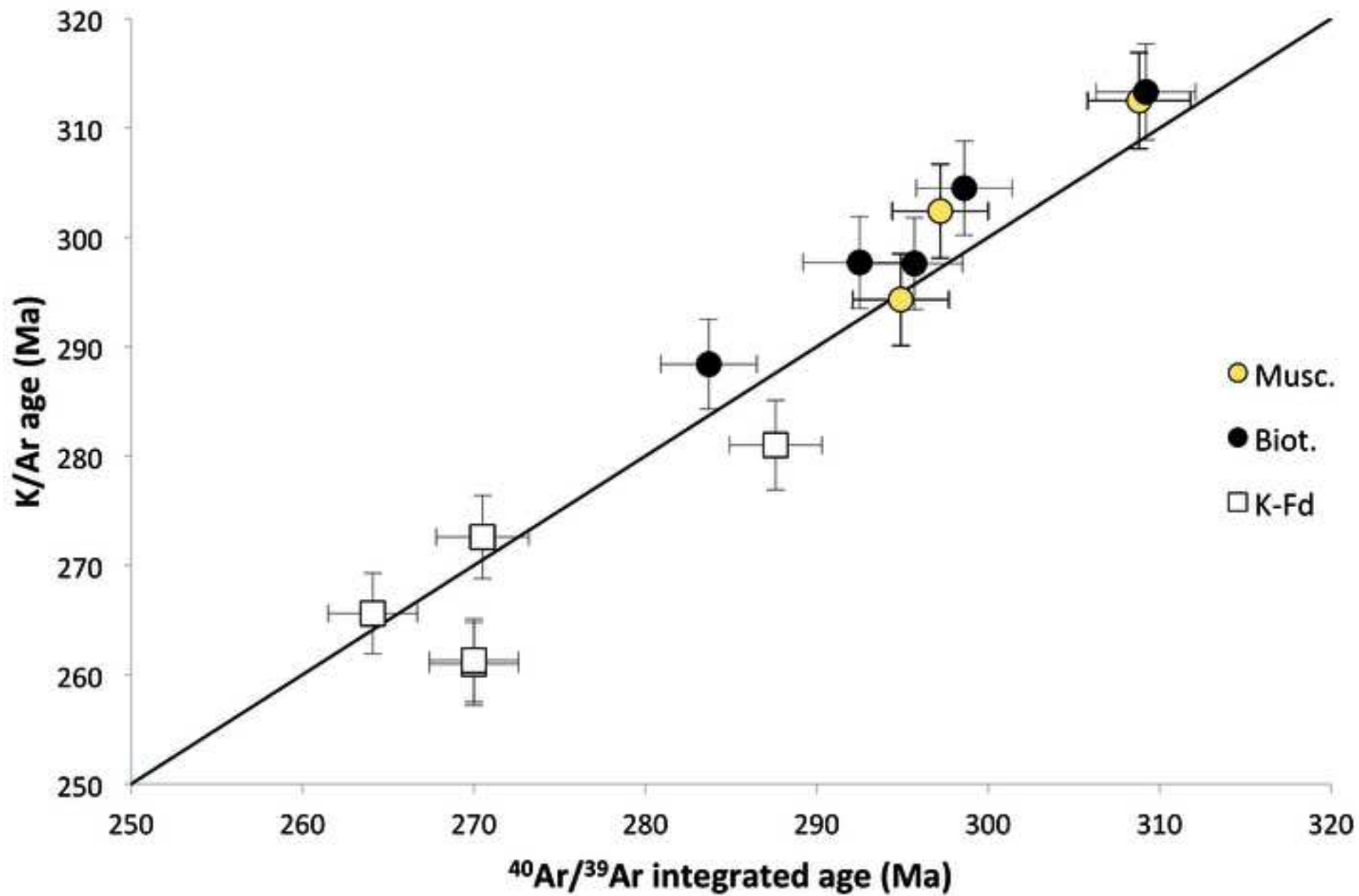


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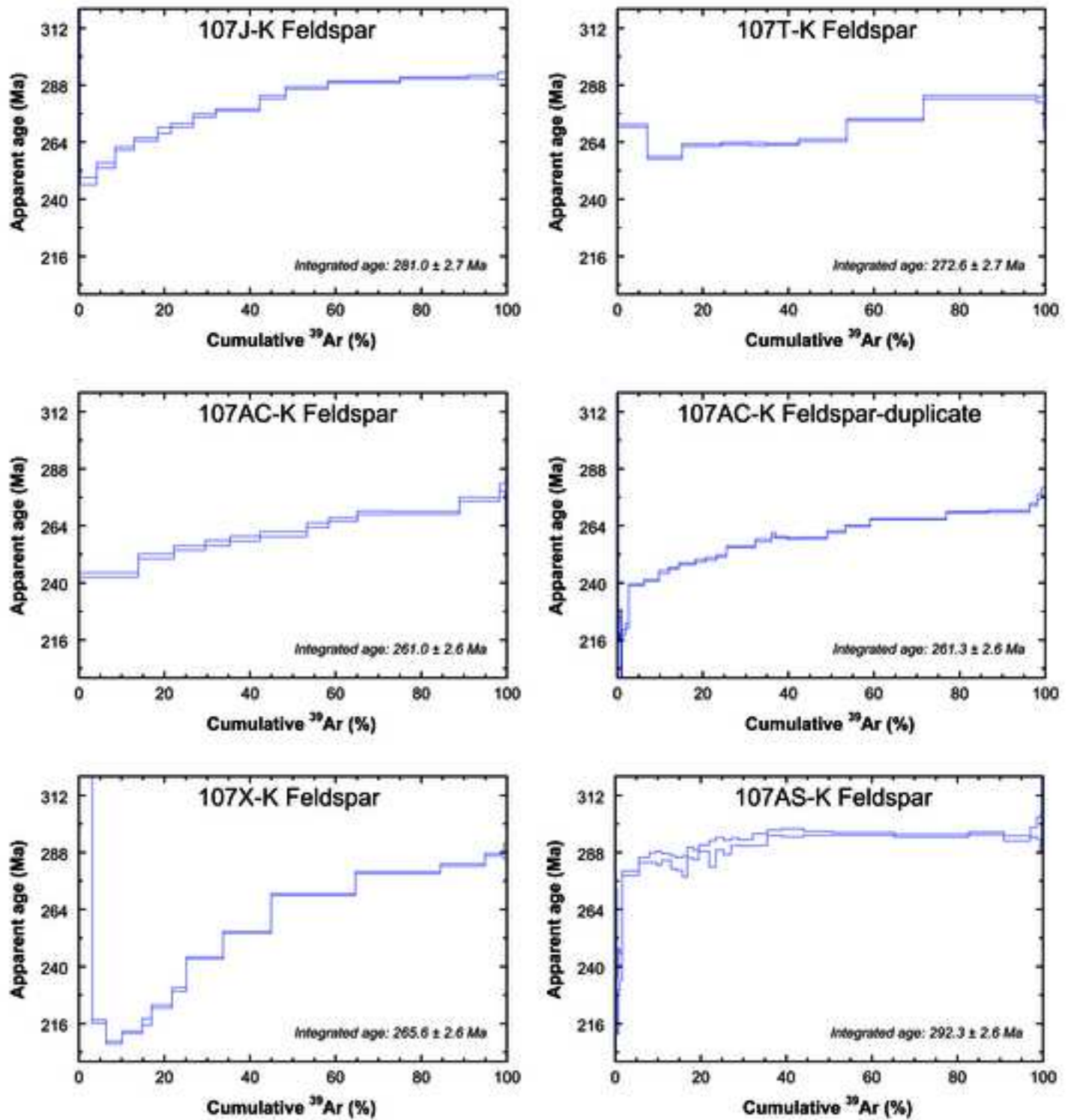


Figure 5: $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra on K-Feldspars of our samples.

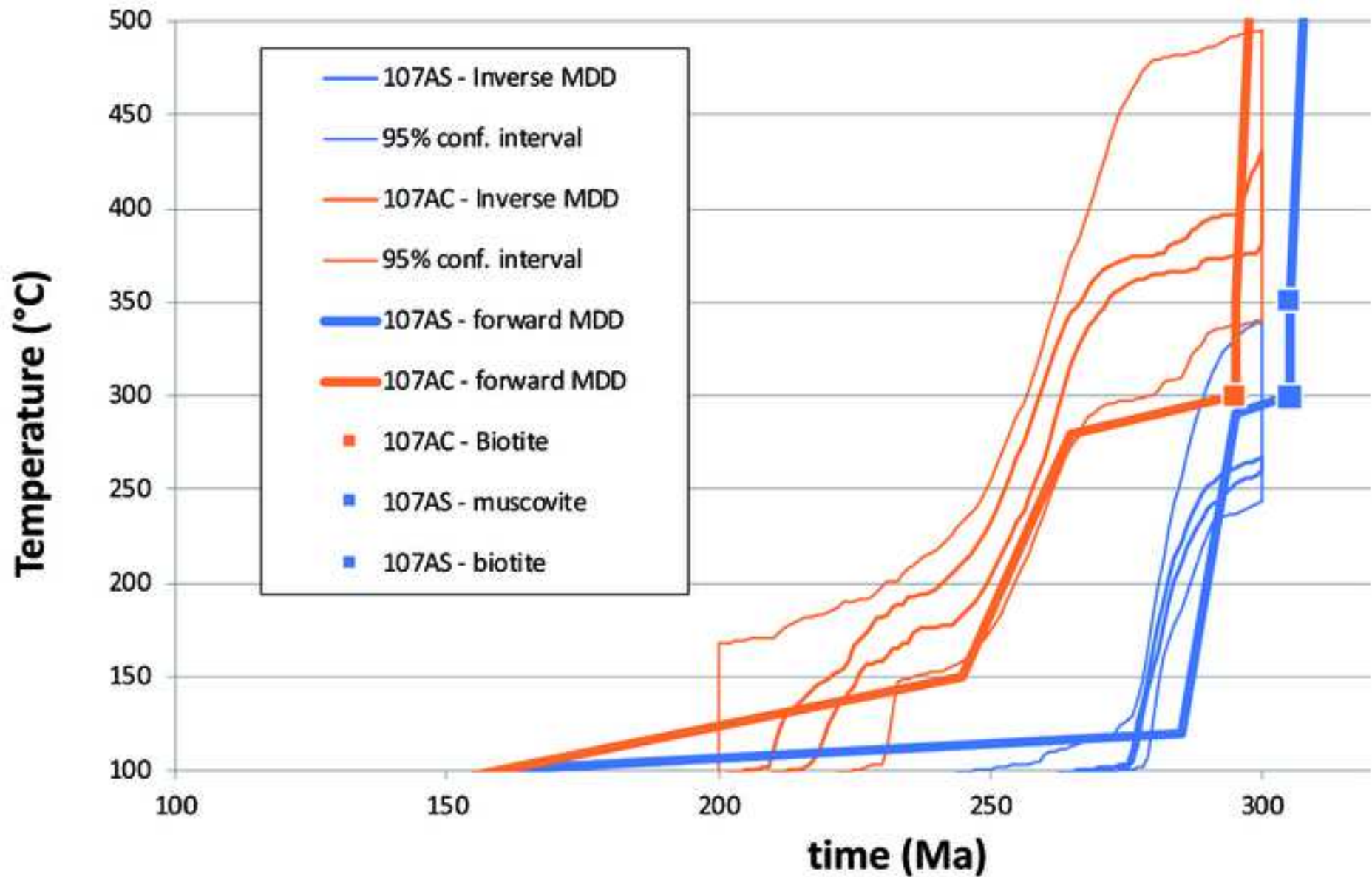


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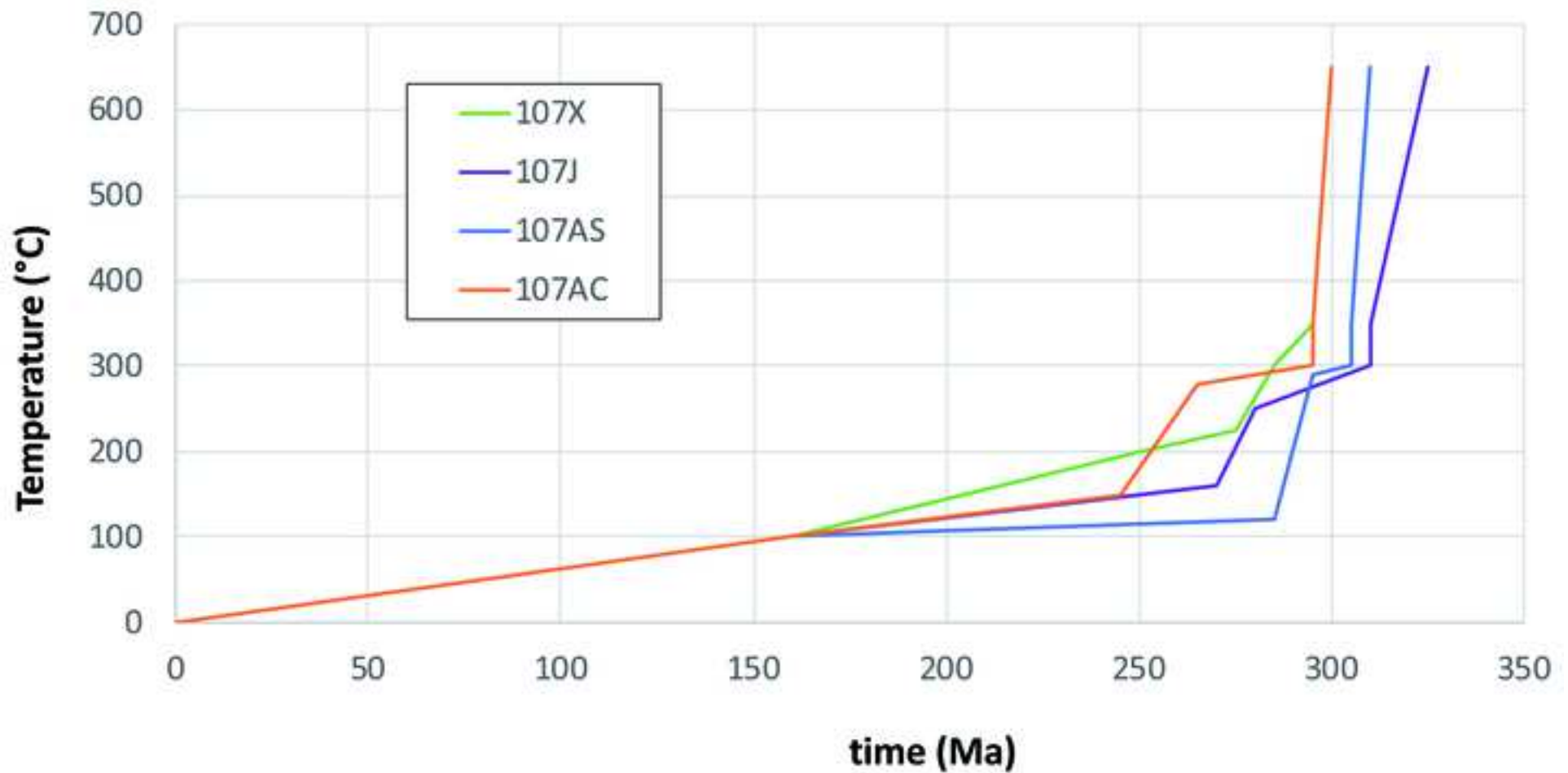


Figure 7: Comparison of the 4 thermal histories obtained with MDD forward modelling on K-Feldspars of our samples 107J, X, AC and AS (all using an imposed activation energy of 46 kcal/mol; Lovera et al., 1997).

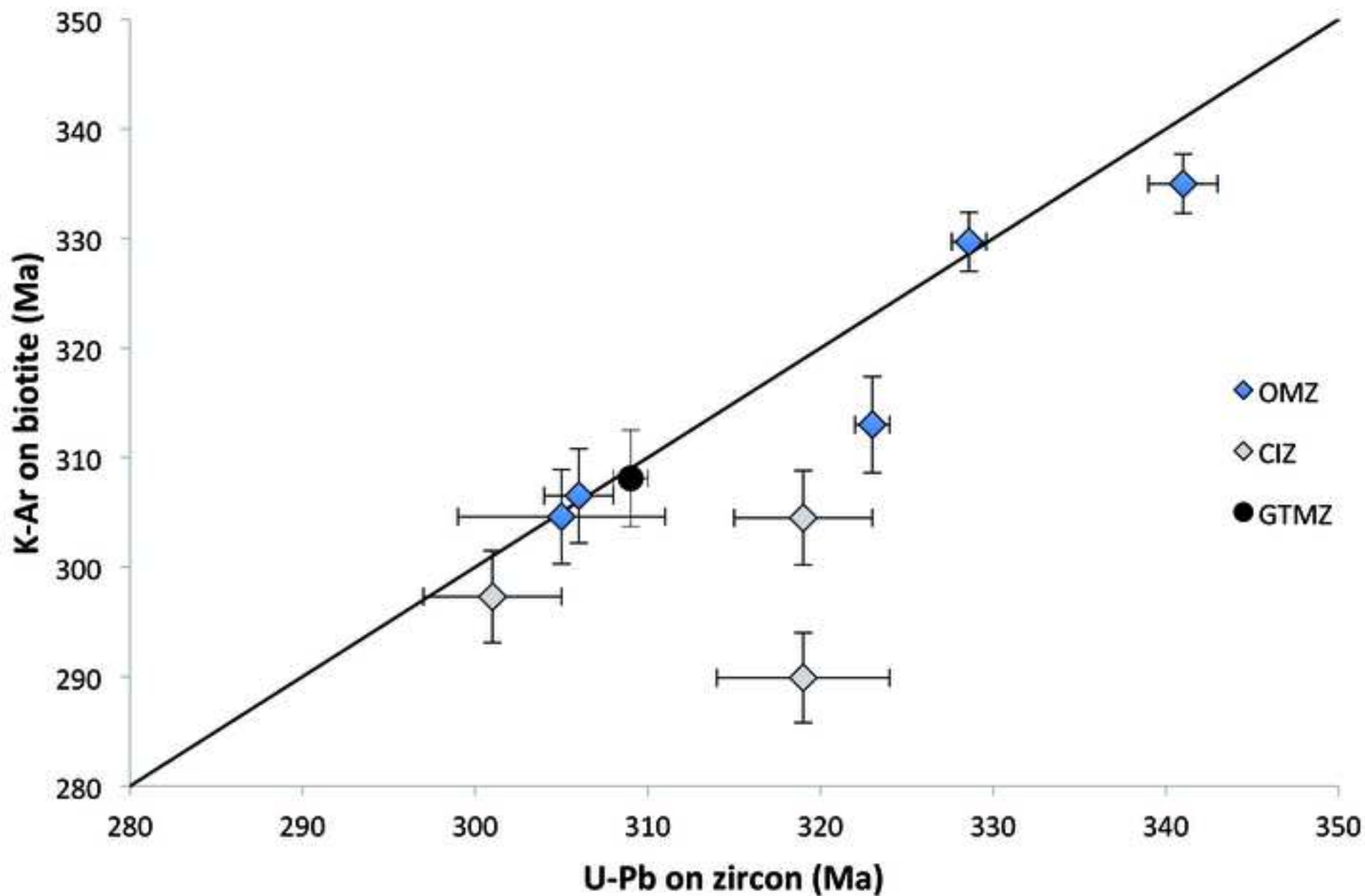


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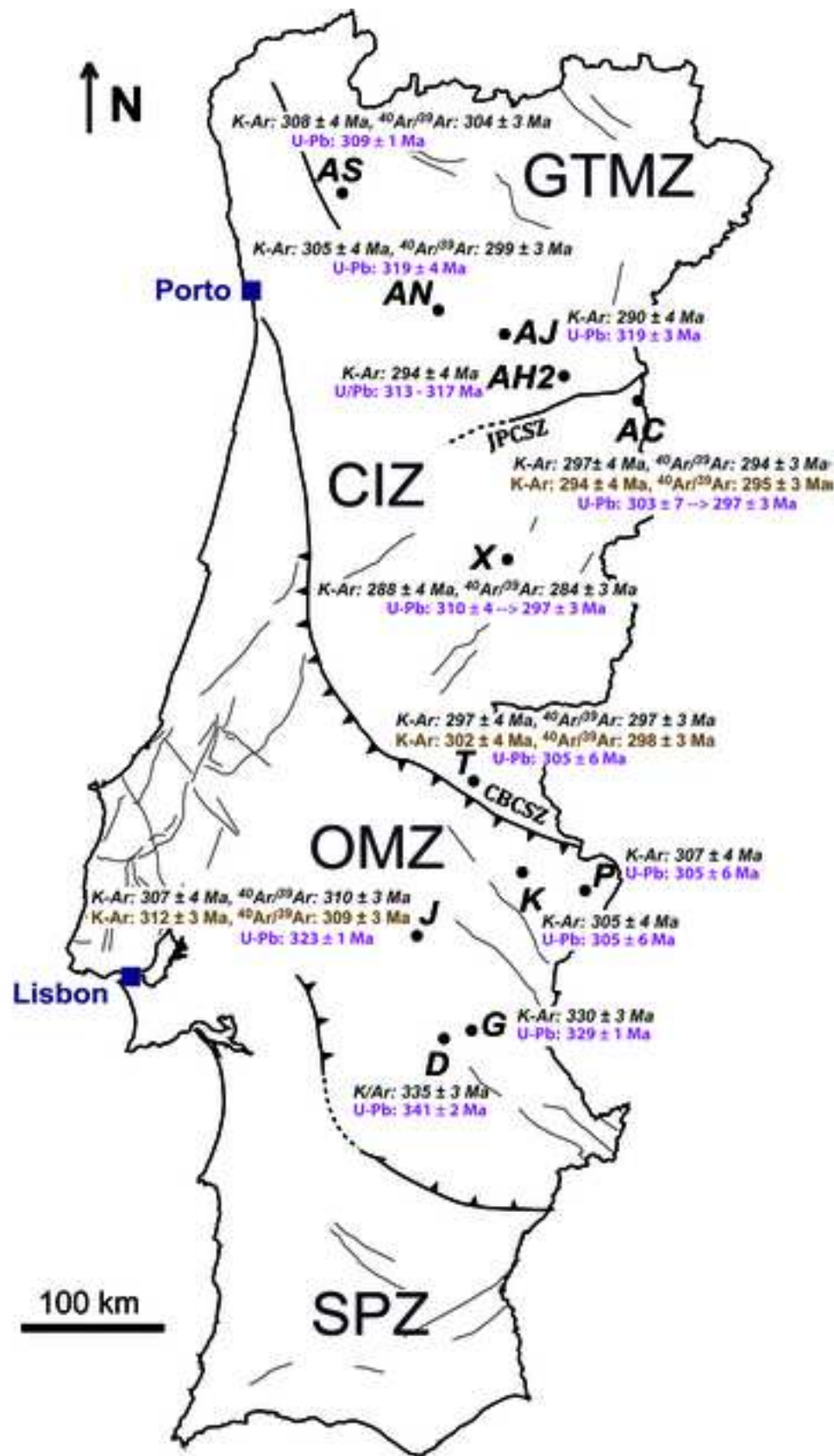


Figure 9: Comparison of our new K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages on micas and available U-Pb on zircons (see references in Table 1). The new ages acquired on biotite and muscovite are shown with black and brown characters, respectively. JPCSZ: Juzbado-Penalva do Castelo Shear Zone; JPCSZ: Juzbado-Penalva do Castelo Shear Zone ; CBCSZ - Coimbra-Badajoz-Cordoba Shear Zone.

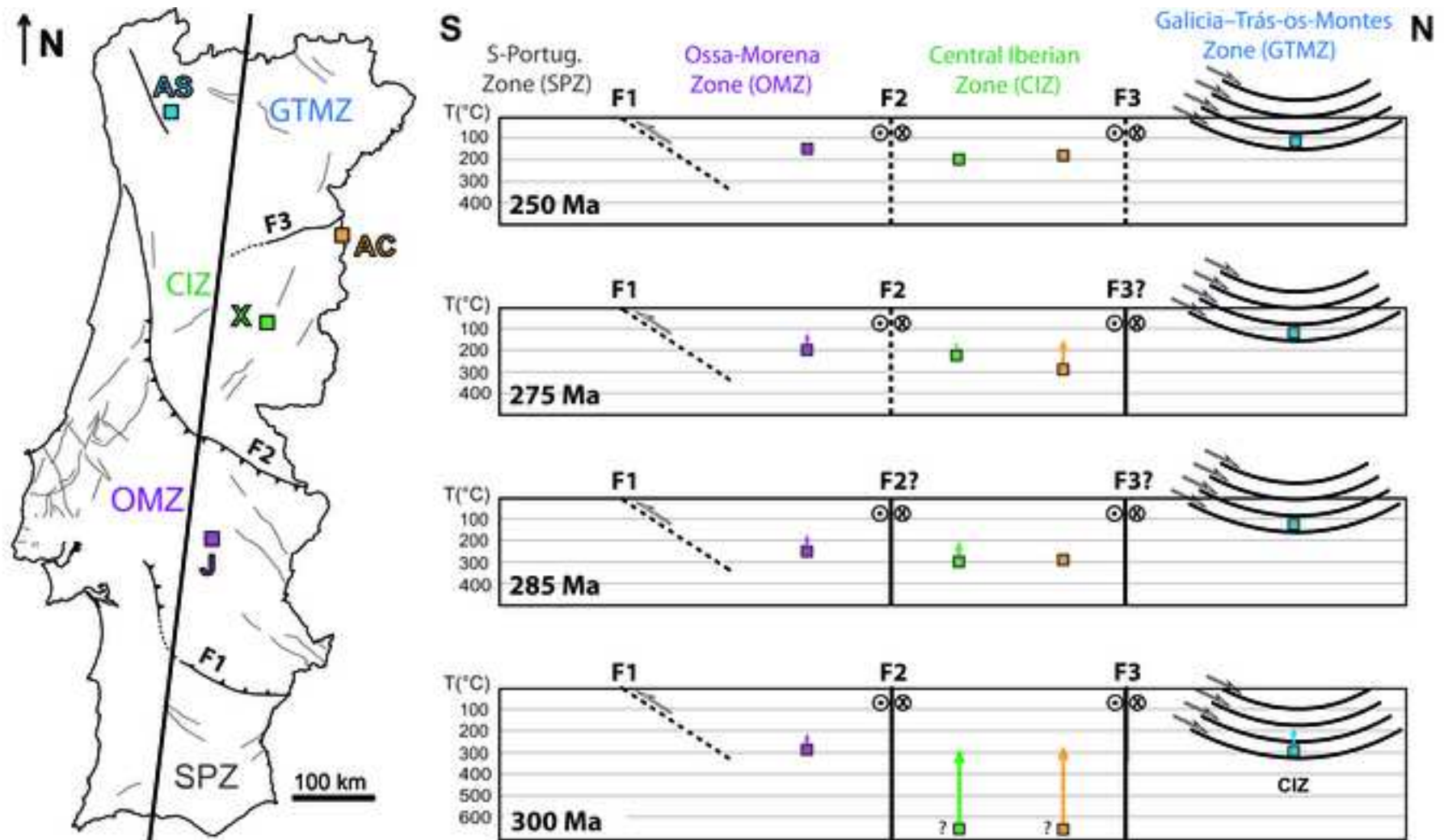


Figure 10: Synthetic cartoon showing the successive vertical positions of samples J, X, AC and AS between 300 and 250 Ma, as deduced from MDD modelling. Sample 107AS is a granite intruded into the nappes of the GTMZ (after thrusting). F1: main thrust separating the South Portuguese Zone (SPZ) and the Ossa-Morena Zone (OMZ); F2: Coimbra-Badajoz-Cordoba Shear Zone (CBCSZ) separating the Ossa-Morena Zone and the Central Iberian Zone; F3: Juzbado-Penalva do Castelo Shear Zone (JPCSZ). Solid lines indicate inferred active fault movements at a given time; dotted lines indicate inactive fault movements. In each structural zone, colored arrows, show the amount of uplift between successive stages.



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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.

Credit Authors statement

Anthony Hildenbrand (AH) and Fernando Ornelas Marques (FOM) designed the study. **AH** and **Xavier Quidelleur (XQ)** achieved the geochronological and thermochronological analyses. **Fernando Noronha (FN)** supervised the aspects regarding minearization processes. All authors (AH, FOM, XQ, FN) contributed to drafting and discussion.