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The Mid-Variscan Allochthon: Keys from correlation, partial retrodeformation and plate-tectonic reconstruction to unlock the geometry of a non-cylindrical belt

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1 **The Mid-Variscan Allochthon: Keys from correlation, partial**
2 **retrodeformation and plate-tectonic reconstruction to unlock the geometry**
3 **of a non-cylindrical belt**

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28 14 **Abstract** The non-cylindrical character of the Variscan Belt is established via a description of
29 15 its most salient geological characteristics and correlations among the main and relatively
30 16 minor massifs from Bohemia to Morocco on the basis of a bibliographical synthesis. Using
31 17 the firmly established tectonostratigraphic domains, the continuity of Variscan zones is
32 18 discussed, as well as the western termination of the one that occupies most of the central part
33 19 of the belt. This zone is characterized by its allochthony, and breaks the continuity of the
34 20 zonation, that is, the cylindricity of the belt. A proposal is made to call this zone the Mid-
35 21 Variscan Allochthon and its wedge shape in map view is explained performing a step by step
36 22 restoration of late Variscan deformation. Once the geometry of the belt is interpreted by a
37 23 succession of regionally significant deformation events, a kinematic reconstruction is
38 24 proposed. It is based on recent developments on the knowledge of Paleozoic plate tectonics
39 25 and intended to show that the deformation model is compatible with current
40 26 paleogeographical developments.

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45 28 **Keywords:** Variscan Belt, Zone correlation, Mid-Variscan suture, Late Variscan restoration,
46 29 North Gondwana, Paleozoic

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114 **1. Introduction**

115 The Variscan Belt is the basement which formed along the Devonian, Carboniferous
116 and early Permian in central and western Europe. Its limits are the Tornquist line or
117 Teisseyre-Tornquist fault zone in the NE, separating the belt from the East European Craton,
118 and the southern part of the Iberian Peninsula. However, the Moroccan Meseta and the Anti-
119 Atlas are usually considered also part of it, and are included here because they are important
120 for our reconstruction. The name derives from Curiae Variscorum, the Latin name of Hof, in
121 Bavaria, Germany, which Tacitus described as the court of the Varisci (or Narisci), and which
122 roughly coincides with the Vogtland region in Saxony and part of the adjacent Bavaria.

123 The first attempt to extend local understanding about the Variscan Belt to the scale of
124 several massifs dates from the late 19th century, when Eduard Suess (1888) compared the
125 central European massifs of the Sudetes, Erzgebirge, Black Forest, Vosges and Massif
126 Central. His work found a continuation during the first third of the 20th century with the work
127 of his son, Franz Eduard Suess (1912, 1926) and of Franz Kossmat (1927), who have shown
128 the allochthonous character of large parts of the Bohemian Massif and established the
129 subdivision of the Variscides that is still in use today.

130 During the second half of the 20th century, the theory of Plate Tectonics prompted the
131 understanding of building of mountain chains by offering an explanation for their formation
132 mechanisms and evolution. The Variscan Belt was not an exception, and received outstanding
133 attention. Large-scale correlations, kinematic and evolutionary models were proposed at the
134 end of the 20th century and beginning of the 21st. These included the movements of large
135 plates and their interactions with smaller plates, generally formed at least in part by
136 continental crust, the so-called terranes. For the global understanding of the belt, significant
137 contributions include those of Philippe Matte (1986, 1991, 2001, 2002), Wolfgang Franke
138 (1989, 2000) and Gérard Stampfli and his collaborators (Stampfli et al., 2002, 2013; Stampfli
139 and Kozur, 2006).

140 However, despite the efforts devoted to the study of the Variscan Belt, and the
141 thousands of published works, the belt is reluctant to be understood as a whole. For instance,
142 there is no agreement on how many oceans were involved, which is to say how many sutures
143 exist. The number of oceans varies between one (Rheic) and five: Rheic, Saxothuringian,
144 Moldanubian, Rhenohercynian and Paleotethys (other names have been used for some of
145 these and eventually different ones). The same is true for the number of peri-Gondwanan
146 terranes separated by these oceans and presently included in the Variscides, which varies
147 between one (Kroner and Romer, 2013) and at least four, even nine if those grouped in
148 superterrane are considered (Stampfli et al., 2013).

149 The present work aims at contributing to a better understanding of the Variscan Belt. It
150 forms part of recent efforts for reconstructing the evolution of the Variscan Belt from the
151 early Paleozoic until the end of Carboniferous, which have been fueled in part by the analysis
152 and tracing of detrital zircon age populations (Franke and Dulce, 2017; Pérez-Cáceres et al.,
153 2017; Casas and Murphy, 2018; Stephan et al., 2019).

154 The paper starts by introducing a synoptic map (Fig. 1) showing the proposed
155 correlation of the tectonostratigraphic zones across the entire belt that will be used later for
156 interpretation. After a first section explaining how the map was constructed, the following
157 sections are devoted to a description of the Variscan massifs, avoiding as much as possible
158 unit names and geographic locations. Abbreviations are limited in the text to important
159 geological entities, while they are widely used in the figures. The Iberian Peninsula is
160 described first because the Iberian Massif offers a good representation of the Paleozoic
161 sedimentary record, the different styles of the Variscan structures and types of plutonism, and

162 serves as a bridge for correlation among the rest of European massifs and northern Africa. A
163 section is devoted to tackle such a correlation.

164 The main target of the paper is however to clarify the role of the units related to the
165 suture that occupies a central position in the Variscan Belt and can be followed from the NW
166 Iberian Massif to the Bohemian Massif. After the regional-scale reviews, it will be shown
167 how these units form part of a wider distinctive tectonic assemblage here referred to as the
168 Mid-Variscan Allochthon, the surrounding zones being considered as its relative Autochthon.
169 Then, a partial retrodeformation of the Allochthon and Autochthon is proposed in a section
170 which explains their current geometry in map view.

171 There is another zone including one or more sutures, known as the Rhenohercynian
172 Zone. Wrapping around the NW part of the belt, it is usually understood as the Rhenic or the
173 Rhenohercynian suture, which are two distinct entities that can however coexist in the same
174 zone. Finally, all tectonostratigraphic zones will be integrated in a comprehensive plate-
175 tectonics reconstruction. The latter is based on published models of Paleozoic kinematics as
176 far as the global framework and main plates are concerned, and on our own interpretation of
177 the Variscan evolution acknowledging the key role of the Mid-Variscan Allochthon.

178 The Variscan Belt is part of a much large collisional orogen that continued toward the
179 west by NW Africa and the Appalachians in North America, and to the east by the Balkans,
180 Carpathians and Caucasus, northern Africa and central southern Asia. The orogeny receives
181 different names depending on the geographical setting, the most popular being Alleghanian in
182 North America, and Variscan in Europe and Asia. These parts of the orogen fall outside what
183 we consider the Variscan Belt in a strict sense, and are not described neither included in our
184 model. The same occurs with the Variscan terranes in southern Europe and northern Africa
185 that were affected by the Alpine orogeny and whose late Variscan position is uncertain,
186 although some references to them are included in the text.

187

188 **2. Geological map of the Variscan Belt**

189 Among early maps sketching the geology of several Variscan massifs and correlating
190 their zones, the one published by Kossmat (1927; an English translation is provided by
191 Meinhold, 2017) deserves special mention. It followed original tectonic subdivision of the
192 central European Variscides established by Suess (1926) and shows the tectonostratigraphic
193 zonation of the central European Variscan Belt. Correlations with the Armorican and Iberian

194 massifs had been proposed by Suess (1888), Kossmat (1921) and Stille (1924), and where
195 supported shortly after by Lotze (1929). The allochthonous character of several Variscan
196 zones was first recognized by Suess (1912, 1926) and Kossmat (1927) in the Bohemian
197 Massif. In the French Massif Central, Variscan nappes were first described by Demay
198 (1942), although allochthonistic interpretations were only accepted since the seventies
199 (Mattauer and Etchecopar, 1976; Briand, 1978). In Iberia, it was first proposed by Carlé
200 (1945) and became progressively accepted following the work of Ries and Shackleton (1971).

201 A correlation of thrust faults and allochthonous nappes between the southern
202 Armorican Massif and NW Iberia was proposed by Matte and Burg (1981), and a map
203 correlating the allochthonous units of the Bohemian Massif with equivalents of the
204 Schwarzwald and Vosges, French Massif Central, South Armorican Domain and the NW
205 Iberian Allochthon was included in Tollmann (1982). Sketches showing the Variscan
206 zonation in many papers bear close similarities and are probably based on Tollmann's map,
207 which is the case of the one presented here, an improved version of maps previously
208 published by the first author and co-workers (Martínez Catalán, 2011; Martínez Catalán et al.,
209 2007, 2020, Ballèvre et al., 2014).

210 Surprisingly, few attempts have been made to explain the weird shape of the
211 allochthonous nappes in these maps. This shape is that of a wedge whose western edge is
212 rather narrow. A model explaining that geometry by a tongue-shaped allochthon derived from
213 southern Baltica, possibly an island arc, emplaced obliquely on the northern margin of
214 Gondwana and deformed by dextral wrench motion was proposed by Martínez Catalán
215 (1990). In later reconstructions (Martínez Catalán et al., 2007; Martínez Catalán, 2011), the
216 island arc hypothesis was abandoned, the original curvature of the allochthon was smoothed
217 and the role of transcurrence was emphasized and put in relation with dextral transpression
218 along a wide shear zone developed between Gondwana and Laurussia. An elaborated version
219 of the latter model is discussed in this paper.

220 The reconstruction shown in Figure 1 represents the Variscan Belt at the early
221 Permian. It postdates ductile and brittle transcurrence, essentially dextral, assumed to have
222 ended by *c.* 290 Ma, and is inspired by the map of present distribution of terranes in Europe
223 (Oczlon, 2006), but the paleoposition of the continental realms is largely based on the
224 reconstruction of Stampfli and Kozur (2006) for 300 Ma (Gzhelian), with several
225 modifications. Early Permian deformation has been accounted for if significant but neglected
226 if subordinate. Ireland has been approached to Great Britain *c.* 40 km to suppress Mesozoic-

227 Cenozoic extension in the Celtic Sea (Dyment et al., 1990). Subsequently, the British Isles
228 have been rotated 8° counterclockwise to compensate for crustal thinning, rifting, and basic
229 igneous activity, probably related with development of oceanic crust in Permian times, and
230 mostly identified in the western part of the English Channel (Smith and Curry, 1975). The
231 position of Morocco in relation with Iberia has been taken from Matthews et al. (2016), and
232 both are separated by a large dextral fault common in plate reconstructions (Le Pichon et al.,
233 1977; Stampfli and Kozur, 2006) and called here the South Iberia Fault (SIF). It is parallel to
234 other late-Variscan faults disrupting the Variscan zonation and previous structures, the more
235 important of which are the South Meseta Fault (SMF), the North Pyrenean Fault (NPF), the
236 Bray Fault (BrF) and the Elbe Fault Zone (EFZ).

237 The position of Iberia in relation to France is shown prior to the opening of the Bay of
238 Biscay, approximately following Le Pichon et al. (1977), and implying a post-Triassic
239 clockwise rotation of *c.* 35° of the Iberian microplate to account for differences in
240 paleodeclination to both sides of the North Pyrenean Fault (NPF; Van der Voo, 1969; Schott,
241 1981; Boillot et al., 1984). Rotation in the southern part of the fault allowed the opening of
242 the Bay of Biscay (Choukroune et al., 1973a, b) while the northern, straight segment acted as
243 a left-lateral fault with a translation of *c.* 400 km between 110 and 75 Ma (Albian-Campanian;
244 Boillot et al., 1984). Modern reconstructions, as that of Nirrengarten et al. (2018), differ
245 somewhat but not essentially from that used here. What matters for the purpose of this
246 contribution is not the precise reconstruction after the latest Variscan movements, but the
247 relative positions of Iberia and the Armorican massif before these movements, which can be
248 constrained with stratigraphy, structures and magnetic anomalies at both sides of the Bay of
249 Biscay.

250 In France, the Armorican, Central and Vosges massifs have been drawn respecting
251 their present-day relative position. Correlation of domains and faults between the two first
252 massifs is in part based on Baptiste (2016). For the Maures-Estérel-Corsica-Sardinia Block
253 (MECS), Edel et al. (2014) calculated a 90° clockwise rotation during the late Carboniferous
254 and early Permian (305-280 Ma). These authors favor a position of the MECS at low latitude
255 in the southern hemisphere, with the Corsica-Sardinia Block oriented W-E in the latest
256 Carboniferous. This was followed by a 90°-100° clockwise rotation in the early Permian,
257 when the block acquired a N-S attitude, while additional 60° clockwise rotation would have
258 occurred during the Triassic.

259 This reconstruction differs from the one which considers a NE-SW orientation for the
1 260 Corsica-Sardinia Block and the absence of rotations prior to the anticlockwise rotation to its
2 261 present N-S position in the Miocene (Dewey et al., 1989; Rosenbaum et al., 2002). The NE-
3 262 SW orientation is contemplated by Rossi et al. (2009), who place this block close to the
4 263 French Massif Central. But for them, the connection between the Massif Central and the
5 264 MECS occurs via a tight orocline in the Bohemian Massif following an extensive Stephanian
6 265 dextral shear zone represented by the external massifs of the Alps, a model based on Matte
7 266 (2001), which is no longer supported by the reconstruction of Edel et al. (2014, 2018). Thus,
8 267 we consider an orientation for the MECS as in Edel et al. (2014, 2018), but occupying a
9 268 somewhat southern position to account for continuity of structures with those of the French
10 269 Massif Central (Fig. 1).

20 270 The Bohemian Massif has not been modified from its present position in central
21 271 Europe, and is maintained in its orientation relative to the French Massif Central and the
22 272 Rhenish Massif.

26 273 More details for the reconstruction are given in the following sections. They describe
27 274 the main outcrops of the Variscan basement in central and western Europe, the Variscan Belt
28 275 in a strict sense, and in Morocco. The description is intended to highlight the aspects
29 276 meaningful for correlation among the massifs and the evolution of the belt as a whole. The
30 277 sections corresponding to the Iberian Peninsula, the French massifs, the Bohemian Massif and
31 278 Morocco include description of the established tectonostratigraphic zonation and other details
32 279 pertinent to the reconstruction of the Variscan Belt.

42 281 **3. Iberian Peninsula**

44 282 The largest outcrop of the Variscan basement is the Iberian Massif, occupying the
45 283 western half of the peninsula. Other are the Pyrenean Axial Zone, Catalonia Coast Ranges,
46 284 Basque massifs and occurrences in the Iberian Chain (Fig. 2a). For outcrops occurring in the
47 285 Betic Range, establishing the boundary with the Mesozoic cover is not always possible, and it
48 286 is difficult to correlate these outcrops with the zones established in the Iberian Massif. They
49 287 are not included in the map, but are briefly described together with rocks of the Rif Domain in
50 288 northern Morocco.

58 289 The Iberian Massif is divided into six zones (Lotze, 1945; Julivert et al., 1972, 1980;
59 290 Ribeiro, 1974; Farias et al., 1987), of which five run parallel to each other delineating the

291 hinge and southern limb of the Ibero-Armorican Arc: Cantabrian, West Asturian-Leonese,
292 Central Iberian, Ossa-Morena and South Portuguese zones (Fig. 2a). The sixth is the Galicia-
293 Trás-os-Montes Zone, an allochthonous ensemble emplaced above the Central Iberian Zone
294 during the Variscan collision, and shown with more detail in Figure 2b and c. The following
295 sections describe their tectonostratigraphic patterns and the Variscan magmatism, and the
296 same is done for the smaller Iberian inliers. Figure 3 sketches the structure of NW, central and
297 SW parts of the Iberian Massif.

298 For a general review of the geology of Iberian Massif or the Iberian Peninsula, the
299 reader is referred to the books edited by Dallmeyer and Martínez-García (1990), Gibbons and
300 Moreno (2002), Vera (2004) and Quesada and Oliveira (2019).

3.1. Cantabrian Zone (CZ)

303 This zone is unique in the Variscan Belt, cropping out in northern Spain, at the core of
304 the Ibero-Armorican Arc (IAA) and possibly in the Basque Massifs of the Pyrenees and the
305 Balearic island of Menorca (Figs. 1 and 2a). It forms a non- or little metamorphic foreland
306 thrust belt developed in the late stages of Variscan deformation. The CZ includes a rather
307 complete Paleozoic preorogenic sequence representing a shallow-water stable continental
308 platform resting unconformably above a thick terrigenous Ediacaran succession. It occupied a
309 proximal segment of the northern passive margin of Gondwana, and recorded tholeiitic and
310 alkaline magmatism related to Cambro-Ordovician rifting (Corretgé and Suárez, 1990).
311 Moreover, it shows relatively ice-proximal Upper Ordovician glacigenics (Gutiérrez-Marco et
312 al., 2010). The synorogenic sequences include terrigenous alternances, carbonates and coal
313 (Truyols et al., 1990; Pérez-Estaún et al., 2004).

314 The CZ is an example of thin-skinned tectonics (Pérez-Estaún et al., 1988, 1994).
315 Deformation occurred during the middle and upper Carboniferous (325-300 Ma), and
316 consisted on east-directed thrusting (Bastida et al., 1984; Pérez-Estaún et al., 1988; Alonso et
317 al., 2009). A subsequent event formed the Asturian Arc (the core of the IAA), an orocline that
318 folded the thrust faults about a vertical axis (Gutiérrez-Alonso et al., 2004; Weil et al., 2012).
319 The formation of the Asturian Arc occurred between the latest Carboniferous and the earliest
320 Permian (310-295 Ma; Weil et al., 2010; Pastor-Galán et al., 2011). Thrusting took place
321 under diagenetic conditions, but a late extensional episode, produced anchizonal and localized
322 epizonal metamorphism at the core of the arc (García-López et al., 1999; Aller et al., 2005;

323 Valín et al., 2016; Blanco-Ferrera et al., 2017), and some small granitic bodies intruded in the
324 metamorphic areas at 292 Ma (Valverde-Vaquero et al., 1999).

325

326 3.2. *West Asturian-Leonese Zone (WALZ)*

327 This zone is a slate belt derived from somewhat more distal parts of the Gondwana
328 margin than the CZ, and is also represented only in Spain. Its stratigraphy is similar to that of
329 the CZ, including a thick terrigenous Ediacaran and a complete, shallow marine Paleozoic
330 succession, generally thick and continuous from the early Cambrian to the Lower Devonian
331 (Pérez-Estaún et al., 1990).

332 The WALZ was affected by recumbent folding and thrusting (Matte, 1968; Bastida et
333 al., 1986; Pérez-Estaún et al., 1991; Martínez Catalán et al., 2003) and by medium pressure
334 Barrovian metamorphism. Crustal thickening was followed by a high-temperature (T) and
335 low-pressure (P) event coeval with extensional collapse which also produced detachments and
336 gneiss domes (Figs. 2a and 3a; Capdevila, 1969; Arenas and Martínez Catalán, 2003;
337 Martínez Catalán et al., 2014).

338 Regional cleavage axial planar to recumbent folds is dated at *c.* 335 Ma to the east of
339 the WALZ and *c.* 360 Ma in the adjacent and more internal Central Iberian Zone (Dallmeyer
340 et al., 1997), indicating diachronism and propagation toward the external domain of the belt.
341 The same can be deduced for thrusting, dated at *c.* 343 Ma in the basal thrust of the
342 allochthonous Galicia-Trás-os-Montes Zone, and 321 Ma at the limit between the WALZ and
343 the CZ. Extensional detachments and doming are late Carboniferous, dated around 315-300
344 Ma.

345

346 3.3. *Central Iberian Zone (CIZ)*

347 Also a slate belt, the CIZ derived from more distal parts of the northern Gondwana
348 platform than the WALZ. Its stratigraphy is similar to that of the CZ and WALZ, although
349 differences in the Ediacaran succession, volcanic component and tectonic style allows defining
350 three domains.

351

352 3.3.1. *Ollo de Sapo and Schist-Greywacke Domains (OSD, SGD)*

1
2 353 The terrigenous Ediacaran succession of the CIZ is known as the Schist-Greywacke
3
4 354 Complex (Teixeira, 1955) or the Alcludian (Tamain, 1971). It reflects sedimentation in a
5
6 355 Cadomian retro-arc basin representing the transition from a Cadomian arc in the periphery of
7
8 356 the West African Craton to Cambrian rifting marking the start of peri-Gondwanan terrane
9
10 357 dispersion (Linnemann et al., 2008). An Ediacaran glaciation (*c.* 565 Ma) has been identified
11 358 by glaciomarine deposits in the southern part of the CIZ (Linnemann et al., 2018).

12
13 359 Late Cambrian-Lower Ordovician voluminous felsic magmatism characterizes most of
14 360 the CIZ, including felsic volcanics dated at 490-475 Ma, and intrusive granitoids of 500-465
15 361 Ma (Valverde-Vaquero and Dunning, 2000; Bea et al., 2006; Zeck et al., 2007a; Solá et al.,
16 362 2008; Montero et al., 2009a; Díez Montes et al., 2010; Rubio Ordóñez et al., 2012; Talavera
17 363 et al., 2013). Volcanic and subvolcanic rocks are particularly voluminous at the NE border of
18 364 the CIZ, where the felsic Ollo de Sapo Fm. crops out at the core of the 570 km long Ollo de
19 365 Sapo Antiform (Fig. 2a, b). The rocks in the antiform were deformed first by recumbent and
20 366 then by upright folds. Due to the occurrence of the Ollo de Sapo Fm. and to deformation
21 367 characteristics, this NE band has been described either as the Ollo de Sapo Domain (OSD;
22 368 Martínez Catalán et al., 2004a) or the Domain of recumbent folds (DRF, Fig. 3a; Díez Balda
23 369 et al., 1990). Most of the rest of the CIZ is described by the same authors respectively as the
24 370 Schist-Greywacke Domain (SGD) and Domain of upright folds (DUF, Fig. 3b).

25
26 371 The main difference between the CIZ and the CZ and WALZ is the presence of the
27 372 Lower Ordovician Toledanian unconformity, responsible for the partial or total absence of
28 373 Cambrian deposits. In the central and southern parts of the CIZ, the Ordovician starts with red
29 374 beds including conglomerates, sandstones, silts and pelites, indicating transgression above a
30 375 paleorelief or/and sedimentation in (half)grabens. Above, the Armorican Quartzite is almost
31 376 always present, followed by more pelitic successions and culminating by a discontinuous
32 377 level of limestone and Hirnantian diamictites, often missing due to a glacio-eustatic
33 378 disconformity (Robardet, 1981; Martínez Poyatos et al., 2004; Gutiérrez-Marco et al., 2019).

34
35 379 The synorogenic sequence, late Tournaisian to Westphalian, consists of paralic to
36 380 flysch-like deposits, the more important of which is Los Pedroches Culm in the south of the
37 381 SGD, up to 6000 m thick and occupying the core of a late-Variscan syncline which coincides
38 382 with its depocentre (Martínez Poyatos, 2002).

383 The CIZ is structurally characterized by the overthrusting of the Galicia-Trás-os-
384 Montes Zone (the Allochthon) on its NW part (Fig. 3a), and by two folding events, one
385 preceding and the other following thrusting. The early folds are overturned or recumbent in
386 the area adjacent to the WALZ in the NW, and in the relatively deep zones cropping out in the
387 Central System (DRF) (Matte, 1968; Macaya et al., 1991; Díez Montes et al., 2010). Axial
388 planar cleavage accompanies these folds. In the DUF, the early folds have an upright attitude
389 (Fig. 3b), but they are not recognized in wide areas and may not develop cleavage. Where
390 present, the early folds were either refolded or flattened by the late folds (Fig. 2a), but the
391 style and size are quite similar for both fold systems. The interference is clear in the SE part
392 of the DUF, and represents the southern part of the hinge zone of a curved structure (Aerden,
393 2004) called here the Central Iberian Arc (CIA) after Martínez Catalán (2012) and Martínez
394 Catalán et al. (2015). Although not fully accepted (*e.g.* Dias et al., 2016), the arc has been
395 acknowledged in several old and recent publications (Staub, 1926, 1927; Llopis-Lladó, 1966;
396 Shaw et al., 2012; Johnson et al., 2013; Simancas et al., 2013; Pastor-Galán et al., 2016).
397 According to Martínez Catalán (2012), the CIA developed contemporaneously with the late
398 folds, which are rather continuous and delineate the southern flank of the IAA (Fig. 2a).
399 Interestingly, in the DRF the late folds are bent by the IAA, suggesting that both arcs are not
400 coeval and the CIA predates the IAA. The first folds predate and the late folds deform early
401 and middle Carboniferous synorogenic deposits, which often alternate with a bimodal
402 volcanic suite linked to an extensional event that can also be identified in the whole SW part
403 of the Iberian Massif (Martínez Poyatos, 2002; Martínez Poyatos et al., 2012).

404 The early folds were dated at 360-335 Ma in NW Iberia, where the emplacement of
405 the Allochthon occurred there around 345-340 Ma. Subsequently, the thickened crust
406 underwent extensional collapse starting at *c.* 335 Ma, whereas late folds are dated at 315-305
407 Ma (Dallmeyer et al., 1997; Valle Aguado et al., 2005; Valverde-Vaquero et al. 2007a;
408 Martínez Catalán et al., 2007, 2009; Rubio Pascual et al., 2013a). Ductile shear zones
409 associated to the late folds have been dated at 310-305 Ma (Gutiérrez-Alonso et al., 2015).

410 Metamorphism is similar to that of the WALZ. Medium and high grade rocks are
411 exposed surrounding the NW Iberian Allochthon and in the Alpine horst forming the Central
412 System. A Barrovian event related with crustal thickening, dated around 350 Ma was
413 followed by a low-P/high-T Buchan episode during subsequent extension at 335-325 Ma, late
414 folding, and renewed extension between 315-300 Ma (Escuder Viruete et al., 1994; Díez
415 Balda et al., 1995; Rubio Pascual et al., 2013a, 2016). Extensional detachments and gneiss

416 domes developed where crustal thickening had been large, in the area covered by the
417 Allochthon and in the Central System (Fig. 2a).

418 No sutures occur among the CZ, WALZ and CIZ, which represent a preserved,
419 coherent segment of the northern Gondwanan platform (Gutiérrez-Marco et al., 1999a;
420 Robardet, 2002, 2003), some aspects being comparable to other outer segments that escaped
421 the Variscan deformation (*e.g.* southern Turkey; Ghienne et al., 2010). The Hirnantian
422 glaciation is identified in the CIZ and CZ (Martínez Catalán et al., 1992; Gutiérrez-Marco et
423 al., 2010; Subias et al., 2015). The boundaries among these three zones are presently Variscan
424 structures such as early folds and thrust or detachment faults (Fig. 3a), thought to represent
425 inverted normal faults of the Gondwana platform which conditioned variations of thickness
426 and facies in the preorogenic sequences (Pérez-Estaún et al., 1990; Martínez Catalán et al.,
427 1992).

3.3.2. Obejo-Valsequillo Domain (OVD)

430 In the southern part of the CIZ, the Obejo-Valsequillo Domain (Apalategui et al.,
431 1985) has a Paleozoic sequence comparable to that of the SGD while the underlying
432 Ediacaran is the same as in the Ossa-Morena Zone. Arc-related intrusives in the Ediacaran
433 have yielded 580-570 Ma (Bandrés et al., 2004; Talavera et al., 2008), and low-grade
434 metamorphism was dated at *c.* 550 Ma (Blatrix and Burg, 1981).

435 Martín Parra et al. (2006) locate the northern limit of the OVD at an extensional shear
436 zone to the north of the Pedroches Carboniferous Culm basin and the elongated Pedroches
437 batholith, but the Paleozoic sequence immediately south of the batholith is rather similar to
438 that of the SGD. The proposed boundary is the northernmost of a series of sinistral wrench
439 faults south of the batholith (Fig. 2a), where the sequence changes and the Lower Devonian
440 directly overlies the early Ordovician.

441 Structurally, the OVD is characterized by thrust faults, NW-SE upright late folds, and
442 high-angle, left-lateral and reverse faults (Fig. 3b, c), all affecting synorogenic Culm and
443 paralic deposits. These are late Variscan structures, but the southern part, the OVD includes
444 an early Variscan large recumbent syncline-anticline pair with NE vergence, probably Upper
445 Devonian in age, which predates the Culm (Azor et al., 1994; Martínez Poyatos, 2002;
446 Martínez Poyatos et al., 1995, 1998, 2012). Thrusting took place during the middle
447 Carboniferous and late folding and faulting, both related to sinistral wrenching, are

448 Westphalian in age (c. 315-300 Ma). Variscan metamorphism ranges between very low grade
449 to lower greenschist facies (Martínez Poyatos et al., 2001).

450

451 *3.4. Ossa-Morena Zone (OMZ)*

452 The OMZ is a slate belt comparable to the CIZ, although with differences in the
453 stratigraphy and structural evolution. At its NE and SW limits, three units depart from the
454 general zone definition and show high-grade metamorphism or even are not metasedimentary.
455 These units are described separately. Most of the OMZ was, as the CIZ, part of the northern
456 Gondwanan platform, although with differences in the faunal content and facies of the
457 Paleozoic, which are more distal in the OMZ (Robardet and Gutiérrez Marco, 1990). The
458 Lower Ordovician shallow-water Armorican Quartzite is absent, but the Hirnantian glatiation
459 is well represented. The Ediacaran is also different, and consists of a lower sequence, the
460 Black Series, with terrigenous sediments, carbonates, black quartzites, cherts, felsic volcanics
461 and amphibolites of possible arc derivation, and an upper unit with massive basic arc
462 volcanism and detrital alternations.

463 The protolith of an amphibolite was dated at c. 611 Ma (Schäfer, 1990), and
464 orthogneisses yielded 585 Ma (Schäfer et al., 1993a). These ages are comparable to those of
465 the OVD in the CIZ and in the eastern part of the WALZ (600 Ma, Fernández-Suárez et al.,
466 1998). Younger protolith ages vary between 560 and 500 Ma (Galindo et al., 1990; Ochsner,
467 1993; Ordóñez Casado, 1998; Salman and Montero, 1999; Montero et al., 2000).
468 Amphibolite-facies metamorphism is dated around 600 Ma in the Black Series while the
469 overlying basic volcanic unit was metamorphic before the Cambrian (560-550 Ma; Dallmeyer
470 and Quesada, 1992). The older ages chart the Cadomian arc while the upper Ediacaran ages
471 correspond to shortening and thickening related to arc-continent collision. The Cambrian ages
472 date the collapse of the Cadomian arc (Bandrés et al., 2004), and the transition to rifting
473 inside the Gondwana margin (Sánchez-García et al., 2008, 2010, 2014), which continued
474 during the Lower Ordovician with bimodal, sometimes alkaline magmatism (470-460 Ma;
475 García Casquero et al., 1985; Galindo, 1989; Ochsner, 1993).

476 The Variscan structure of the central part of the OMZ includes a first event of
477 recumbent folds with a NW-SE attitude and SW vergence followed by thrusting toward the
478 SW, extension and late upright NW-SE folds (Expósito et al., 2002, 2003). Early folding was
479 Devonian, and a synorogenic flysch rests unconformably above. Most is early Carboniferous

480 (Boogaard and Vázquez, 1981; Giese et al., 1994), but Lower Devonian fossils found at the
481 base may be inherited or indicate an older age (Piçarra, 1997; Piçarra et al., 1998; Pereira et
482 al., 1998).

483 The early folds were cut and displaced to the SSW by brittle-ductile thrust faults
484 probably developed in continuation of folding. Later, during the early Carboniferous, an
485 extensional episode produced normal faults and basins where detrital and carbonate rocks
486 were deposited together with volcanics. The late folds are related to Namurian-Westphalian
487 sinistral transpression, and were accompanied by reverse and left-lateral faults whose
488 development continued until the Stephanian (Simancas et al., 2001; Expósito et al., 2002).
489 Sinistral wrenching affected the whole OMZ (Expósito et al., 2002).

490 It has been proposed that the limit between the CIZ and OMZ represents a suture,
491 although its age has been considered Cadomian, then reworked as a Variscan shear zone
492 (Ribeiro et al., 1990; Quesada, 1991; Ábalos et al., 1991; López-Guijarro et al., 2008),
493 Cambro-Ordovician (Palacios et al., 2010), or Variscan (Brun and Burg, 1982, Matte, 1986;
494 Azor et al., 1994; Simancas et al., 2001, 2002). A difference of 15° in paleolatitude was
495 reported by Perroud et al. (1985) for the Upper Devonian on opposite sides of the CIZ/OMZ
496 boundary (Figs. 1 and 2a). Such a difference, with the OMZ occupying the southernmost
497 position, was questioned by Robardet (2002) based on paleobiographical constraints. He
498 considered the CIZ and OMZ as different paleogeographic domains and proposed a more
499 distal (western) position of the OMZ off NW Africa, but did not interpret the limit as a suture.

500

501 *3.4.1. Central Unit (CU)*

502 Supporting the Variscan suture hypothesis is the occurrence of eclogites in the so-
503 called Central Unit (Azor Pérez, 1997; Azor et al. 1994), a confusing name for a unit used to
504 define the northern boundary of the OMZ. This unit is a narrow band consisting on migmatitic
505 paragneisses, schists, Lower Ordovician orthogneisses and amphibolites with occasional
506 eclogitic relics. High-P metamorphism (1.45-1.9 GPa, 550 °C (Ábalos et al., 1991; López
507 Sánchez-Vizcaíno et al., 2003; Pereira et al., 2010), is dated at 377±19 Ma (Abati et al.,
508 2018). An oceanic signature has been inferred from some of the basic rocks (Gómez Pugnaire
509 et al., 2003), but the absence of ophiolites casts doubt on the suture, which in any case would
510 be cryptic, hidden by the strong left-lateral imprint of the Badajoz-Córdoba Shear Zone
511 (BCSZ) at the CIZ/OMZ boundary.

512 The structural significance of the CU is hindered by its coincidence with the BCSZ
1 513 and the fact that it is bounded by two faults (Fig. 3b, c), a normal detachment with left-lateral
2 514 component at its northern limit and a sinistral structure probably hiding an older thrust fault
3 515 responsible for the exhumation of the high-P rocks to the south (Azor et al., 1994; Simancas
4 516 et al., 2001). The seismic profile IBERSEIS (Fig. 3c), crossing the OMZ and the OVD, shows
5 517 NE-dipping reflections that coincide with these two faults (Simancas et al., 2003). The
6 518 seismic profile ALCUDIA, traversing the southern part of the CIZ (Fig. 3b), also images a
7 519 NE-dipping reflection coincident with the northern fault (Martínez Poyatos et al., 2012).

14 520 The CU generally shows the foliation dipping to the NE or subvertical, which
15 521 reinforced by the NE-dipping seismic reflections led to interpret a subduction of the OMZ
16 522 under the CIZ (Simancas et al., 2001, 2003). However, some geological transects across the
17 523 middle part of the CU unveil an antiformal structure (Apalategui Isasa and Higuera, 1980;
18 524 Arriola et al., 1981; Gonçalves, 1971). Ultramytonites at the hinge zone of the antiform show
19 525 a stretching lineation trending NW-SE to N-S, with top to the north kinematics, and
20 526 migmatitic paragneisses are quickly replaced upward or northward by schists. This points to a
21 527 N-dipping ductile extensional detachment related with the ascent of a gneiss dome that could
22 528 explain the exhumation of the high-P rocks. The dome was later stretched by left-lateral
23 529 shearing along the BCSZ, being presently very narrow (Fig. 2a).

34 530

35 531 *3.4.2. Évora-Aracena Metamorphic Belt (EAMB)*

36 532 The southern boundary of the OMZ is interpreted as a suture, partly occurring at the
37 533 EAMB, a low-P/high-T domain with a high-P ensemble located tectonically above, the
38 534 Cubito-Moura Unit (Rubio Pascual et al., 2013b). The Évora Massif in Portugal is wide (Fig.
39 535 2a), and has been interpreted as a gneiss dome where a ductile detachment with top to SE
40 536 kinematics separates gneisses and schists, below, from the overlying Cubito-Moura Unit, a
41 537 phyllonitic mélange including lenses of eclogite and blueschists (Araújo and Ribeiro, 1996;
42 538 Dias da Silva et al., 2018) which attest for a high-P metamorphic event dated at 370-360 Ma
43 539 (Moita et al., 2005). The high-P rocks are tectonic imbrications mixed with ophiolitic
44 540 fragments and emplaced toward the N-NE before early Variscan recumbent folding with west
45 541 and SW vergence occurred (Araújo and Ribeiro, 1996; Simancas et al., 2003). The fragments
46 542 include MORB-type, c. 480 Ma old metabasites in greenschist and amphibolite facies and
47 543 ultramafics, interpreted as remnants of the Rheic Ocean (Fonseca et al., 1999; Pedro et al.,
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544 2010). Low-P/high-T granulite-facies metamorphism of the EAMB took place at 343-330 Ma
545 (Dallmeyer et al., 1993; Castro et al., 1999), while magmatism associated to dome
546 development is dated at 343-319 Ma (Dias da Silva et al., 2018). Late upright folding
547 occurred in a left-lateral shearing environment bounded by the Southern Iberian Shear Zone
548 (SISZ) which separates the OMZ from the SPZ (Crespo-Blanc and Orozco, 1988, 1991).

549 In Spain, the Aracena Belt is a narrow, strongly sheared band, the upper part of which
550 includes high-P rocks in the Cubito-Moura Unit and also in the underlying Fuenteheredos
551 Group. Both form part of a NE-verging orogenic wedge with pieces of subducted continental
552 and oceanic crust, later emplaced on top of the OMZ (Rubio Pascual et al., 2013b). Part of
553 this wedge is preserved also in the South Portuguese Zone.

554 3.4.3. Beja-Acebuches Unit (BAU)

555 The BAU is a narrow mafic and ultramafic unit separating the OMZ and the SPZ,
556 where diorite and granodiorite were dated at 352 Ma (Pin et al., 2008; Pin and Rodríguez,
557 2009) and metabasites at 345-330 Ma (Azor et al., 2008, 2009). It has been interpreted as i) an
558 ophiolite (Andrade, 1985; Crespo-Blanc and Orozco, 1991), ii) linked to an arc related with
559 northward oceanic subduction (Santos et al., 1990; Quesada et al., 1994), iii) the product of
560 ocean ridge subduction (Castro et al., 1996a, 1996b), iv) reflecting slab break-off following
561 subduction and collision (Pin et al., 2008), and v) an ephemeral oceanic realm opened after
562 consumption of the Rheic lithosphere and possibly related to a mantle plume (Azor et al.,
563 2008).

564 According to its age, the BAU can be related to the extensional event affecting the
565 southern part of the CIZ, the OMZ and the SPZ between the uppermost Devonian and the
566 Viséan, which produced abundant magmatism and an elevated thermal gradient in the upper
567 crust. This magmatism is linked to the emplacement of the Iberian Reflective Body (IRB),
568 identified in the IBERSEIS seismic profile (Simancas et al., 2003), and was responsible for
569 the middle Carboniferous low-P/high-T metamorphism undergone by the EAMB (Rubio
570 Pascual et al., 2013b). The IRB intruded along a detachment rooted in the middle crust (Fig.
571 3c), and a possible link with local oceanization or a mantle plume has been suggested by
572 Simancas et al. (2006) and Azor et al. (2008). The latter was questioned by Pin et al. (2008)
573 based on the geochemical signature of the Beja igneous complex.
574

575 The BAU was affected by the middle Carboniferous, low-P granulite-facies
1 576 metamorphism of the EAMB, and is strongly sheared by the SISZ. High-T mylonites
2 577 developed in a context of ductile thrusting, and greenschist-facies mylonites developed later
3 578 related with sinistral wrenching (Díaz Azpiroz and Fernández, 2005; Díaz Azpiroz et al.,
4 579 2006; Pérez-Cáceres et al., 2016). Actually, the whole OMZ, together with the OVD of the
5 580 CIZ and the northern part of the South Portuguese Zone can be considered a wide and
6 581 heterogeneous shear zone, with two main zones where deformation concentrated, the BCSZ in
7 582 the NE and the SISZ in the SW.

17 584 *3.5. South Portuguese Zone (SPZ)*

19 585 This zone represents a Variscan foredeep basin characterized by thin-skinned thrust
20 586 tectonics, and is usually correlated with the Rhenohercynian Zone of Kossmat (1927) in the
21 587 Bohemian Massif (Fig. 1). It is separated from the OMZ by an accretionary prism, the Pulo do
22 588 Lobo Unit, and by the BAU (Figs. 2a and 3c; Silva et al., 1990; Onézime et al., 2002;
23 589 Nysaether et al., 2002; Simancas et al., 2003; Nance and Linnemann, 2008). The Pulo do
24 590 Lobo and the ofiolitic relicts of the Cubito-Moura Unit in the OMZ represent the Rheic
25 591 suture, as the age of the latter (480 Ma) is that of the opening of the Rheic Ocean (Franke et
26 592 al., 2017). So, the basement beneath the SPZ thrust belt should be that of Avalonia, which is
27 593 supported by zircon age populations (de la Rosa et al., 2002; Braid et al., 2011; Pereira et al.,
28 594 2014; Rodriguez et al., 2015; Pérez-Cáceres et al., 2016, 2020).

38 595 According to Rubio Pascual et al. (2013b), at *c.* 400 Ma the ocean was closed after
39 596 subduction toward the SW (in present coordinates). Previously, an arc was built above, and is
40 597 preserved in La Minilla, a unit of the SPZ with evidences of high-P metamorphism. The
41 598 Lower to Middle Devonian Pulo do Lobo Unit was probably, deposited in the back-arc, and
42 599 later subducted, as indicated by lawsonite pseudomorphs. These two units represent a NE-
43 600 verging accretionary wedge partly preserved in the SPZ (Onézime et al., 2002; Rubio Pascual
44 601 et al., 2013b) and partly in the Cubito-Moura Unit and the Fuenteheridos Group of the
45 602 EAMB. They derive from the SW margin of the OMZ and were thrust above it after
46 603 subduction.

55 604 Building and exhumation of the accretionary wedge occurred at the Lower-Middle
56 605 Devonian (Silva et al., 1990; Eden and Andrews, 1990), and Upper Devonian to early
57 606 Carboniferous synorogenic formations were deposited on top of high-P rocks. Terrigenous

607 formations were laid down and during the uppermost Devonian and Visean alternate with
608 submarine, mostly siliceous volcanics, including massive sulfide deposits forming the
609 Volcano-Sedimentary Complex during an extensional episode (Dunning et al., 2002; Rosa et
610 al., 2009). Extension and magmatism affected the SPZ and OMZ, indicating that the two
611 zones were accreted by the Upper Devonian (Simancas et al., 2003, 2005; Pereira et al., 2007a;
612 Pereira et al., 2006b, 2007b; Pin et al., 2008). If early Carboniferous extension created a
613 narrow ocean, this was the Rhenohercynian, its suture being represented by the BAU.

614 Renewed shortening during the Visean was recorded by the Culm, representing in the
615 SPZ a foredeep basin with sedimentation propagating toward the SW while a thin-skinned
616 thrust belt developed from the late Visean until the Moscovian (*c.* 330-310 Ma; Simancas et
617 al., 2004). Synkinematic low-P metamorphism affecting the Upper Devonian and
618 Carboniferous rocks increases to the north from diagenesis to anchimetamorphic and
619 greenschist facies, reaching andalusite, cordierite and sillimanite zones in the east, around the
620 northern Sierra de Sevilla batholith (Abad et al., 2001, 2004).

622 *3.6. Galicia-Trás-os-Montes Zone (GTMZ)*

623 This zone represents the NW Iberian Allochthon (Fig. 2a), a nappe pile preserved in
624 five large complexes and consisting in three groups of stacked allochthonous units and an
625 underlying Parautochthon (Martínez Catalán et al., 2009; Ballèvre et al., 2014; Arenas et al.,
626 2016). The Upper Allochthon is a peri-Gondwanan terrane rifted at *c.* 500 Ma leaving behind
627 an oceanic domain, the Galicia-Brittany-Massif Central Ocean (Matte, 2001, 2007). The
628 Middle Allochthon includes units derived from this Cambro-Ordovician ocean which
629 recorded renewed activity during the Lower Devonian. The Lower Allochthon derives from
630 the outermost margin of north or NW Gondwana and the Parautochthon represents a more
631 proximal region of the same margin. The Allochthon was progressively assembled in an
632 accretionary prism during early Variscan convergence, their units being stacked during
633 subduction or underthrusting between 400-395 Ma (Upper Allochthon) and 375-365 Ma
634 (Lower Allochthon), before being thrust above the CIZ at *c.* 340 Ma having incorporated
635 the Parautochthon to its base.

636 Three of the allochthonous complexes occur in Galicia (NW Spain): Cabo Ortegal,
637 Órdenes and Malpica-Tui (Figs. 2b and 3a), and two in Trás-os-Montes (northern Portugal):
638 Bragança and Morais (Fig. 2c). They are circular, elliptical or elongated klippen preserved in

639 synforms and structural basins developed among gneiss domes, and more or less flattened
640 subsequently by the late set of upright folds. The following sections describe the
641 allochthonous groups from bottom to top of the nappe stack.

642

643 *3.6.1. Parautochthon*

644 This ensemble consists of early Carboniferous synorogenic deposits forming a rather
645 continuous carpet at the base of the Allochthon, tectonically overlain by lower Paleozoic
646 sediments and metavolcanics. Both are internally imbricated and the whole package is limited
647 by a basal detachment developed along a weak level of Silurian black shales. Deformation is
648 pervasive in both ensembles, and metamorphism is of the greenschist facies, but the
649 synorogenic sediments are characterized by lower temperature.

650 The synorogenic deposits include flysch-type, turbiditic deposits with detrital zircons
651 indicating an early Carboniferous maximum depositional age and derivation essentially from
652 the Allochthon (Dias da Silva et al. 2014a; Martínez Catalán et al., 2008, 2016). A different
653 type comprises alternations of slate with sandstone or limestone and include small to large
654 olistoliths of the preorogenic sequence (Dias da Silva et al. 2014b; González Clavijo et al.,
655 2016, 2021). The synorogenics were deposited in flexural basins in front of the allochthonous
656 orogenic wedge, being later incorporated into thrust sheets (Martínez Catalán et al., 2016;
657 Farias and Marcos, 2019).

658 The preorogenic sequence is comparable to that of the CIZ although more distal.
659 Greenish and purple (lie-de-vin) schists are common (Ribeiro, 1974), as well as
660 metavolcanics, bimodal but dominantly felsic interbedded with the upper Cambrian and
661 Ordovician sediments comparable to the Cambro-Ordovician of the Autochthon. The
662 youngest zircon population found in Galicia has 560 Ma (Díez Fernández et al., 2012a). In
663 Portugal, basic rocks are dated at 494 Ma, and felsic volcanics identical to the Ollo de Sapo
664 Fm. of the CIZ yield also the same age, 485 Ma (Dias da Silva et al., 2014c). A metaquartzite
665 provided a few *c.* 486 Ma detrital zircons (Dias da Silva et al., 2014a). Younger Ordovician to
666 Silurian volcanism includes alkaline felsic rocks dated between 475-440 Ma (Valverde-
667 Vaquero et al., 2005, 2007b), as well as basic tuffs, metabasalts and diabase sills (Ferragne,
668 1972; Gallastegui et al., 1988; Ancochea et al., 1988; Meireles, 2013).

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670 *3.6.2. Lower Allochthon*

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2 671 This group of units consists of a thick sequence with paragneisses (metagreywackes),
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4 672 schists, phyllites, carbonaceous schists, calc-silicate rocks and meta-quartzites intruded by
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6 673 late Cambrian-Early Ordovician igneous rocks. The youngest zircon population in Galicia has
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8 674 566 Ma, suggesting an Ediacaran age of the sequence. Age spectra indicate a paleoposition to
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10 675 the NE of the West African craton (Díez Fernández et al., 2010). In Portugal, the sequence
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12 676 includes early Paleozoic metasediments and a bimodal volcanic suite dated at 455 Ma (Dias
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14 677 da Silva et al., 2016).

15 678 Intrusive rocks consist of a granitic calc-alkaline suite formed by high-K granites,
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17 679 granodiorites, tonalites, an alkaline to peralkaline suite, and basic rocks (Floor, 1966;
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19 680 Rodríguez Aller, 2005; Montero et al., 2009b; Abati et al., 2010; Díez Fernández et al.,
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21 681 2012b). Rb-SR and U-Pb dating shows magmatic pulses at *c.* 490 Ma for granodiorites, 475
22
23 682 Ma for granites, and 475-470 Ma for alkaline/peralkaline intrusions (Van Calsteren et al.,
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25 683 1979; García Garzón et al., 1981; Santos Zalduegui et al., 1995; Díez Fernández et al.,
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27 684 2012b). Geochemical data point to a geodynamic setting in a back-arc or retro-arc
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29 685 environment (Fuenlabrada et al., 2012).

30 686 Metamorphism records an early Variscan high-P/low- to medium-T metamorphism
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32 687 identified by eclogites, jadeite-bearing orthogneisses and blueschists, and interpreted as the
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34 688 product of continental subduction (Munhá et al., 1984; Schermerhorn and Kotsch, 1984;
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36 689 Ribeiro, 1991; Gil Ibarra and Dallmeyer, 1991; Arenas et al., 1995). According to Ar-Ar
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38 690 and U-Pb data, subduction may have started before 370 Ma and ended at *c.* 365 Ma (Van
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40 691 Calsteren et al., 1979; Santos Zalduegui et al., 1995; Rodríguez et al., 2003; Abati et al.,
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42 692 2010). Emplacement by thrusting occurred after 346 Ma, as shown by U-Pb age of monazite
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44 693 in migmatites developed during exhumation (Abati and Dunning, 2002).

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48 695 *3.6.3. Middle Allochthon*

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50 696 This group includes true oceanic units and others interpreted as transitional between
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52 697 the extended continental margin of northern Gondwana and the Rheic oceanic crust. Ages of
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54 698 igneous protoliths are either Cambro-Ordovician or Lower Devonian, which leads to a
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56 699 subdivision in two groups, Cambro-Ordovician ophiolites and oceanic supracrustals, and
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58 700 Lower Devonian ophiolites. When both groups occur together, the older occupies a lower
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701 structural position, just above the Lower Allochthon, implying that it was located closer to the
1 702 extended margin of Gondwana (Martínez Catalán et al., 1996, 2002).

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4 703 The Cambro-Ordovician group includes three types of units: i) with a clear oceanic
5 704 affinity, ii) continental-oceanic transitional units and iii) the ophiolitic part of a tectonic
6 705 mélange. The first type is well represented by an unit consisting on metagabbroic
7 706 amphibolites, gabbros, pegmatoid gabbros, pyroxenites, ultramafic rocks and minor tonalites
8 707 and leucogabbros (Martínez et al., 2012; Arenas and Sánchez Martínez, 2015). Their
9 708 geochemistry is that of island-arc tholeiites, N-MORB and within-plate basalts, with trace
10 709 elements similar to those generated in plume-ridge interactions. Isotopic composition of
11 710 zircons points to a juvenile mafic protolith without interaction with crustal components. U-Pb
12 711 zircon ages yield two populations of 495 and 475 Ma dating respectively the gabbroic
13 712 protolith and granulite facies metamorphism transitional between low- and medium-P types,
14 713 possibly related to mid ocean ridge subduction. Other units of the first type are strongly
15 714 deformed and consists of a tectonically repeated succession of metabasites, metapelitic
16 715 phyllites, metagabbros and serpentinites with some layers of granitic orthogneisses and
17 716 metacherts. Basic rocks plot in the fields of island-arc tholeiites, supra-subduction zone
18 717 basalts or destructive plate-margin basalts. An orthogneiss yielded 497 ± 4 Ma, and their
19 718 immobile trace elements point to a volcanic arc or a supra-subduction zone granitoid (Arenas
20 719 et al., 2007a, b; Arenas and Sánchez Martínez, 2015). These units underwent a high-P event
21 720 followed by a mylonitic greenschist-facies foliation, axial planar to tight recumbent folds and
22 721 dated at 363-367 Ma (Dallmeyer et al., 1997).

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39 722 The second type of Cambro-Ordovician units is interpreted as the upper part of a
40 723 transitional crust between the attenuated Gondwana margin and an oceanic domain. They
41 724 consist of phyllites, schists, mafic igneous rocks with N-MORB affinity and alkaline
42 725 metavolcanics. Detrital zircons in metasediments yield a maximum depositional age of 512
43 726 Ma and suggest a position relatively distal in relation to a late Cambrian source (Díez
44 727 Fernández et al., 2010). These units were affected by blueschist-facies metamorphism, dated
45 728 at 363 ± 2 Ma, followed by amphibolite to greenschist facies during exhumation by thrusting
46 729 followed by extensional collapse dated at 337 ± 3 Ma (Pereira et al., 2003; Rodríguez Aller,
47 730 2005; López-Carmona et al., 2010, 2013, 2014).

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56 731 The third type occurs in the eastern of the Cabo Ortegal Complex, and is formed by an
57 732 ophiolitic mélange in the upper part and a mixture of ophiolitic and metasedimentary slices
58 733 (Arenas et al., 2009). The ophiolitic part includes two igneous series, one with calc-alkaline

734 affinity and the other with tholeiitic island-arc affinity. Both were generated in a mature
735 volcanic arc in the periphery of Gondwana between 527 and 485 Ma (protolith ages of
736 amphibolites and orthogneisses). The metasediments include phyllites, phyllonites, marbles,
737 quartzites and a conglomerate whose detrital zircons yielded a population of 630-464 Ma
738 constraining the end of magmatic arc activity to the Lower Ordovician. The mélange was
739 formed during the Upper Devonian in a Variscan subduction zone.

740 The Lower Devonian ophiolites are well represented in the Órdenes and Morais
741 Complexes. They include metaperidotites (harzburgites) overlain by metagabbros and diabase
742 to pegmatoid dikes (Díaz García et al., 1999a; Pin et al., 2002, 2006). The chemistry of the
743 basic rocks is that of island arc tholeiites or transitional between N-MORB and island-arc
744 tholeiites. A negative Nb anomaly suggests a supra-subduction zone setting (Sánchez
745 Martínez et al., 2007a). The protoliths are dated at 405-395 Ma, and initial highly positive
746 ϵNd values point to magmas being juvenile (Díaz García et al., 1999a; Pin et al., 2002, 2006).

747 More deformed, often mylonitic representatives consist of foliated amphibolites
748 derived from gabbroic protoliths and mylonitic greenstones alternating with pelitic schists and
749 inclusions of metagabbros and serpentinites (Ribeiro, 1974; Marques et al., 1991-1992;
750 Sánchez Martínez et al. 2007b; Arenas et al., 2014). The amphibolites derive from island-arc
751 tholeiites and may represent the plutonic section of an arc-related ophiolite. Its age is 400-395
752 Ma (Sánchez Martínez et al., 2007b, 2011; Arenas et al., 2014) but the age spectra include a
753 zircon population of 1155 ± 14 Ma, interpreted as extracted from the mantle during partial
754 melting or from a Mesoproterozoic basement. The mylonitic greenschists are transitional
755 between island-arc tholeiites and N-MORB, with a small negative Nb anomaly.

756 P-T conditions recorded by garnet amphibolites in a metamorphic sole of an
757 amphibolite-facies unit of the Órdenes Complex yielded 650 °C and 1.15 GPa, pointing to a
758 relatively hot subduction environment for ophiolite imbrication (Díaz García et al., 1999a).
759 The prograde foliation was dated at 390-380 Ma (Peucat et al., 1990; Dallmeyer et al., 1991,
760 1997), and greenschist-facies mylonitic fabric at 364 Ma (Dallmeyer et al., 1997).

761 762 *3.6.4. Upper Allochthon*

763 This ensemble includes two groups separated by a subtractive, low-dipping
764 detachment. They have a comparable lithological association but differ in the metamorphic
765 evolution: the structurally lower group underwent early Variscan high-T and high-P

766 metamorphism reaching eclogite facies. In the overlying group, the early Variscan event was
767 Barrovian but recorded a Cambro-Ordovician, granulite-facies intermediate-P event . Both
768 groups consist of a thick terrigenous sequence intruded by Cambro-Ordovician gabbros,
769 granitoids and diabase dikes. Ultramafics occur only in the lower group, where serpentized
770 ultrabasic rocks of one unit contribute, together with similar rocks of a Lower Devonian
771 ophiolite, to the magnetic anomaly of the eastern Órdenes Complex (Martínez Catalán et al.,
772 2012), used for correlation with the Armorican Massif (Fig. 1).

773 The metasediments of the lower group are high-grade paragneisses whose maximum
774 depositional age is *c.* 510 Ma (Albert et al., 2015). Low grade units occurring on top of the
775 upper group include metapelites and scarce quartzites and conglomerates. There, the youngest
776 detrital zircon age population (530-500 Ma), and a diabase dike (510 Ma) indicate a middle
777 Cambrian maximum depositional age (Fernández-Suárez et al., 2003; Díaz García et al.,
778 2010; Castiñeiras et al., 2010; Fuenlabrada et al., 2010). The intrusives are dated at 520-490
779 Ma in both groups (Dallmeyer and Tucker, 1993; Abati et al., 1999; Ordóñez Casado et al.,
780 2001; Fernández-Suárez et al., 2002, 2007). In the lower group, basic rocks are tholeiitic,
781 related to continental rifting (Gil Ibarguchi et al., 1990; Galán and Marcos, 1997), while in the
782 upper group they are tholeiites generated in an ensialic magmatic arc (Andonaegui et al.,
783 2002, 2016), which is confirmed by the chemistry of granitoids (Andonaegui et al., 2012).

784 Ultramafic rocks are spinel-bearing harzburgites, dunites and pyroxenites of oceanic
785 affinity (Girardeau and Gil Ibarguchi, 1991) or representing the mantle root of a magmatic
786 arc. Sm-Nd dating yielded a protolith age of *c.* 500 Ma, whereas growth of garnet in
787 pyroxenite dated at 390 Ma reflects high-P/high-T, early Variscan metamorphism (Moreno et
788 al., 2001; Santos et al., 2002). In the intrusives and metasediments of the lower group, this
789 event reached high-P granulite and eclogite facies (Vogel, 1967; Hubregtse, 1973; Arenas and
790 Martínez Catalán, 2002) dated at *c.* 400 Ma (Schäfer et al., 1993b; Santos Zalduegui et al.,
791 1996; Ordóñez Casado et al., 2001; Fernández-Suárez et al., 2002, 2007). Subsequent
792 exhumation produced migmatization dated at 395-385 Ma (Fernández-Suárez et al., 2002),
793 and continued by amphibolite-facies mylonitization at 390-380 Ma (Dallmeyer et al., 1991,
794 1997; Valverde Vaquero and Fernández, 1996; Gómez Barreiro et al., 2006) and Variscan
795 recumbent folding and thrusting under amphibolite and mainly greenschist facies conditions.

796 In the upper group, the intermediate-P metamorphism is dated at 495-482 Ma, an age
797 slightly younger than that of the large igneous granodioritic and gabbroic protoliths, close to
798 which granulite facies was attained. Metamorphism is attributed to crustal thickening nearly

799 coeval to magmatic underplating in the ensialic arc (Abati et al., 1999, 2003, 2007;
1 800 Andonaegui et al., 2002). Rutile of 390-380 Ma (Abati, 1999) and Ar-Ar ages of
2 801 amphibolites and metasediments (400-370 Ma; Dallmeyer et al., 1997; Gómez Barreiro et al.,
3 802 2006) date an early Variscan Barrovian metamorphic overprint associated to recumbent folds
4 803 and thrusts.
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9 804 In the Upper Allochthon, changes in metamorphic grade are abrupt and occur along
10 805 extensional detachments (Díaz García et al., 1999b; Álvarez Valero et al., 2014). Even if their
11 806 metamorphic histories differ, the lower and upper groups are considered parts of the same
12 807 terrane because no suture has been found between them. Actually, relics of the Cambro-
13 808 Ordovician high-T metamorphic event characteristic of the upper group has also been
14 809 identified in the lower group (Fernández-Suárez et al., 2002).
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22 23 811 *3.7. Variscan intrusive magmatism in the Iberian Massif*

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25 812 The Variscan Belt is characterized by high-T metamorphism and the abundance of
26 813 granitoids. It is a hot orogen whose causes, according to Rey et al. (1997) include i) large
27 814 crustal thickening resulting from collision, ii) the fact that later extension was caused by
28 815 gravitational collapse instead of plate dynamics, and iii) perhaps high initial geothermal
29 816 gradients and mantle involvement. In the Iberian Massif, granitoids are abundant in the NW,
30 817 beneath and around the GTMZ Allochthon, in the Central System and in the SW at both sides
31 818 of the OMZ/SPZ boundary. Their types and the origin of such a distribution are discussed in
32 819 the next sections.
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41 821 *3.7.1. Synkinematic granitoids*

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43 822 In NW Iberia, Capdevila (1969) and Capdevila and Floor (1970) identified three
44 823 groups of Variscan granitoids, which were recognized in other parts of the Iberian Massif
45 824 (Fig. 2a). The first and oldest group includes per- and metaluminous, biotite-rich
46 825 synkinematic granodiorites, often with tonalitic facies, representing magmas with poor fluid
47 826 content. They are allochthonous in relation to their country rocks and derived from partial
48 827 melting of the lower crust with some contribution from the mantle (Galán et al., 1996).
49 828 Vaugnerites (durbachites) have been found associated to these rocks in NW Iberia (Gil
50 829 Ibarra, 1981; López-Moro and López-Plaza, 2004; Gallastegui, 2005). The age of this
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830 group was established at 345-320 Ma (Serrano-Pinto et al., 1987; Fernández-Suárez et al.,
831 2000; Gallastegui, 2005; Gloaguen, 2006; Gloaguen et al., 2006; López-Moro et al., 2017).

832 Early Carboniferous plutons, dated at 347-337 Ma, intruded in or close to the
833 Parautochthon and Lower Allochthon of the GTMZ (Gutiérrez-Alonso et al., 2018). The age
834 of thrusting of the GTMZ upon the CIZ is 345-340 Ma (Dallmeyer et al., 1997; Abati and
835 Dunning, 2002), and none of the plutons piercing the GTMZ is older than 345 Ma, while
836 several of them intruded along faults that cut across thrust faults related to the emplacement
837 of that allochthonous zone. The transition from building the allochthonous accretionary
838 wedge to the establishment of a collisional regime took place at 360 Ma (Martínez Catalán et
839 al., 2007, 2009), and crustal thickening of the orogenic crust started then. Deep-generated
840 magmas reached upper crustal levels 15-25 m.y. later, when the Allochthon had been
841 emplaced yet. Intrusion of the younger plutons of the granodiorite suite at 325-320 Ma is
842 coeval with an extensional event which followed 15-25 m.y. of thermal relaxation after
843 thrusting of the GTMZ at 345-340 Ma (Martínez Catalán et al., 2014; Alcock et al., 2015).
844 Partial melting of the lower crust and upper mantle probably resulted from radiogenic heating
845 aided by decompression and by water supplied through crustal-scale shear zones.

846 The second group of granitoids consists of peraluminous two-mica synkinematic
847 granites and adamellites, rich in fluid phases and metasedimentary enclaves. They derive from
848 metasediments and orthogneisses of the middle crust, and are subautochthonous to
849 allochthonous. Their ages fall within the 322-305 Ma time range, with a maximum between
850 310-305 Ma. At that time, the middle crust had reached the temperature necessary to generate
851 melts of monzogranitic composition. These granitoids are concentrated in the GTMZ and
852 around it in the CIZ, suggesting a relationship with crustal thickening and subsequent
853 collapse, as in the case of the younger synkinematic granodiorites. 1-D and 2-D numerical
854 thermal models (Alcock et al., 2009, 2015) show that thickening induced by recumbent
855 folding and thrusting in the GTMZ, CIZ and WALZ may explain the low-P/high-T late-
856 Variscan metamorphic event and also granite production.

857 Granitoids of the two synkinematic groups were deformed during emplacement along
858 the extensional event, and by strike-slip ductile shear zones clustering around 310-305 Ma
859 (Gutiérrez-Alonso et al., 2015) associated to which late folds developed (Iglesias Ponce de
860 León and Choukroune, 1980).

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862 3.7.2. *Postkinematic granitoids*

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2 863 The third group of granitoids (Capdevila, 1969; Capdevila and Floor, 1970) includes a
3
4 864 wide range of compositions from gabbro (scarce) to granodiorite and monzogranite, which
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6 865 have the common characteristic of being post-kinematic and allochthonous in relation to the
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8 866 country rocks. They are abundant in the CIZ surrounding the GTMZ, in the Central System
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10 867 and to the south until the limit with the OMZ (Fig. 2a). Late- to post-kinematic granitoids
11 868 have been dated at 310-285 Ma (Gutiérrez-Alonso et al., 2011 and references therein).

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13 869 For those surrounding the GTMZ, thermal modeling by Alcock et al. (2015) explains
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15 870 the voluminous granitic magmatism by crustal thickening followed by thermal relaxation and
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17 871 collapse. They show that the lithospheric mantle gets thinner, facilitating a contribution of the
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19 872 asthenosphere to the heat budget needed for the voluminous magmatism in the internal zones
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21 873 of NW and Central Iberia.

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23 874 South of the GTMZ, and in the western part of the Central System batholith, most
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25 875 massifs are postkinematic (Fig. 2a), but some older massifs exist (322-310 Ma), and a smooth
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27 876 transition from the Barrovian event to low-P/high-T conditions and granite intrusion is
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29 877 observed (Valle Aguado et al., 2005, 2017; Díaz-Alvarado et al., 2013; Pereira et al., 2018).
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31 878 In the eastern part of the Central System batholith, all granitoids are postkinematic (308-298
32 879 Ma), result from partial melting of the lower crust, and their high-grade country rocks were
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34 880 cold at the time of intrusion (Orejana et al., 2012; Villaseca et al., 2017). In anatectic areas of
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36 881 central Iberia, Castiñeiras et al. (2008) found that the later generation of granitoids was
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38 882 unrelated with mid-crustal anatectic processes and had a deeper crustal derivation, so that the
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40 883 mechanisms for granite production there was different, with a stronger mantelic heat input
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42 884 and little contribution of crustal thickening. Several small stocks of postkinematic gabbros
43 885 and diorites support this interpretation (Zeck et al., 2007b; Orejana et al., 2009).

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45 886 An alternative model put forward by Pereira et al. (2015a) defends that the late
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47 887 Carboniferous and early Permian batholiths (315-280 Ma) are calc-alkaline, related with the
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49 888 subduction of Paleotethys oceanic lithosphere and the formation of the Iberian orocline
50
51 889 (IAA). Their model is that of plume-assisted relamination, related to a Paleotethyan magmatic
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53 890 arc and unrelated with the Variscan collision.

54
55 891 In the southern part of the CIZ, two large postkinematic elongated plutons occur
56
57 892 parallel to the NW-SE CIZ/OMZ boundary. In the NW, the Nisa-Albuquerque pluton, dated
58
59 893 at 310-305 Ma, straddles the Portugal/Spain border (Fig. 2a) and consists of S-type

894 monzogranite to syenogranite, with a core of I-type tonalite-granodiorite (Solá et al., 2009). In
1 895 the SE, Los Pedroches is a long batholith (Figs. 2a and 3b) dated at 314-304 Ma, whose
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3 896 composition varies from granodioritic to granitic, and that hosts a bimodal dike complex also
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5 897 striking NW-SE (Carracedo et al., 2009). The ensemble occupies the core of the late syncline
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7 898 where the Pedroches Culm is preserved (Fig. 3b). The pluton intruded along a previous crustal
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9 899 to lithospheric-scale fracture (Ábalos and Díaz Cusí, 1995) during an extensional episode
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11 900 responsible for a S-dipping extensional shear zone and faults at the northern limit of the Culm
12
13 901 (Martín Parra et al., 2006). In this case, no previous crustal thickening may justify crustal
14
15 902 melting, while mantle input (basaltic underplating?) and extension seems the main causes of
16
17 903 postkinematic magmatism.

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21 905 *3.7.3. Granitoids of the Rhenohercynian arc*

23 906 A magmatic complex straddles the OMZ/SPZ boundary. It occurs at the SW of the
24
25 907 OMZ, mostly intruded in the EAMB (Fig. 2a), where it is known as the Beja Igneous
26
27 908 Complex. In the east of the SPZ, it is known as the Northern Sierra de Sevilla Batholith. The
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29 909 complex is an association of ultramafic to felsic intrusions dated at 355-335 Ma, the same age
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31 910 as the BAU and partially coeval with the Volcano-Sedimentary Complex formed during the
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33 911 uppermost Devonian to Viséan extensional episode in the SPZ. The left-lateral Southern
34
35 912 Iberian Shear Zone (SISZ) displaced the northern part to the west relative to the southern part,
36
37 913 and the BAU is essentially a stripe of the Beja gabbro (342 or 352 Ma; Jesus et al., 2007; Pin
38
39 914 et al., 2008) strongly sheared by the SISZ.

40 915 The rocks belong to two magmatic suites, tholeiitic-calc-alkaline (basic intrusives and
41
42 916 ultrabasic cumulates) and calc-alkaline (monzogranites, subvolcanic granitoids and
43
44 917 intermediate rocks; de la Rosa et al., 1992; de la Rosa and Castro, 2004). Basic rocks
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46 918 originated from a basaltic magma and produced some crustal assimilation, generating the
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48 919 intermediate and felsic rocks. The abundance of Mesoproterozoic zircons in a composite
49
50 920 pluton (gabbro to granite) adjacent to the BAU suggests that the rocks pierced the SPZ crust,
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52 921 as zircons of that age are uncommon in the OMZ (Gladney et al., 2014), which supports the
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54 922 contention that the SPZ basement is of Avalonian type (Fig. 3c).

55 923 Given the calc-alkaline character of magmatism, genetic models implying subduction
56
57 924 are the more reasonable. These include a suprasubduction setting (Quesada et al., 1994)
58
59 925 perhaps including slab break-off (Pin et al., 2008) or ridge subduction (Castro et al., 1996a,

1996b). Because of its age, subduction cannot be that of the Rheic lithosphere, but that of the
Renohercynian Ocean, the closure of which formed the Renohercynian arc identified in the
NW of the Bohemian Massif and active between 360-325 Ma (Zeh and Will, 2010).

A group of granitoids in the central and eastern parts of the OMZ are also related with
the Renohercynian arc (Fig. 2a). Their composition includes ultramafics, gabbro, tonalite,
granodiorite, and minor bodies of granite and leucogranite. Intermediate rocks are the most
abundant. They are metaluminous, show a calc-alkaline character and are dated at 355-330
Ma (Casquet and Galindo, 2004).

3.8. Other Variscan massifs of the Iberian Peninsula

3.8.1. La Demanda and Iberian Chains

The basement outcrops of La Demanda and the Iberian Chain are considered the
continuation of the WALZ (Fig. 2a). Both share a similar Neoproterozoic and Paleozoic
sequence (Gutiérrez-Marco, 2004a; Gutiérrez-Marco et al., 2002, 2004; Liñán et al., 2002,
2004) with the difference of a thick Devonian sequence of siliciclastics and carbonates in the
Iberian Chain (Carls et al., 2004), missing in the WALZ due to erosion. The Variscan
structure in these Alpine basement uplifts includes steep to overturned early folds (pre-
Westphalian), some thrust faults in La Demanda and only upright early folds in most of the
Iberian Chain. Shortening and crustal thickening was less important than in the WALZ.
Metamorphism is of low-T greenschist facies and a slaty cleavage accompanies the folds,
which delineate the northern limb and part of the hinge zone of the CIA (Fig. 2a).

3.8.2. Pyrenees and Catalonia Coast Ranges

The massifs occurring at the northern limb of the IAA include the central and eastern
Pyrenees and the Catalonia Coast Ranges. They can be correlated with the WALZ and CIZ of
the Iberian Massif, but two main differences exist. One is the age of the Ordovician
voluminous felsic magmatism, which is younger in the Pyrenees (477-446 Ma; Casas et al.,
2017; Casas and Murphy, 2018; Clariana et al., 2017), than in the CIZ (500-465 Ma; Solá et
al., 2008; Montero et al., 2009a; Díez Montes et al., 2010; Talavera et al., 2013). The other is
the Sardinian unconformity, found elsewhere in the Mediterranean province (Álvaro et al., 2017;
Casas, 2010) but not found in the CIZ (González-Clavijo et al., 2017). In western Pyrenees,

957 the Basque Massifs (Fig. 2a) differ from the rest of the chain because the Lower Ordovician
958 includes shallower water successions and the Sardinic unconformity is absent (Gutiérrez-
959 Marco, 2004b). This, and the absence or very low grade metamorphism rather suggest a
960 correlation with the CZ. Moreover, a similar late Carboniferous conodont fauna has been
961 found in the CZ and the Basque Quinto Real Massif (Sanz-López and Blanco Ferrera, 2012,
962 2013; Rodríguez-Cañero et al., 2018).

963 In the central and eastern Pyrenees (Axial Zone), an early Carboniferous Barrovian
964 event was followed by a low-P/high-T event (Guitard et al., 1995; de Hoÿm de Marien et al.,
965 2019). A detachment separates an upper, low-T crust (superstructure) from a middle crust
966 with numerous gneiss domes and postkinematic granitoids (infrastructure). Partitioning of
967 deformation occurred at 310-290 Ma in a context of dextral transpression (Carreras and
968 Druguet, 2014; Cochelin et al., 2017): while the middle crust recorded E-W lateral flow and
969 dome emplacement, the upper crust was thickened developing E-W folds and a steep, low-
970 grade axial planar cleavage. Late growth of kyanite occurred locally, caused by isobaric or
971 slightly compressional cooling during dextral transpression (Guitard et al., 1995).

972 Folds previous and oblique to the E-W system, are recognized in the Basque Massifs.
973 They are overturned toward the west or SW and often cut across by thrust faults, which is
974 coherent with a continuation of the northern limb of the IAA. These early structures affect the
975 synorogenic Namurian flysch, which supports correlation with the CZ.

976 The Catalonia Coast Ranges show stratigraphic affinities with the Variscan
977 Mediterranean province and can be correlated with the Pyrenean and Sardinian successions
978 (Gutiérrez-Marco, 2004a). No differences exist between the NE and SW massifs (Gutiérrez-
979 Marco et al., 2004), which leaves little room for the continuation of the CZ to the south. A
980 right-lateral fault trending NW-SE (Malod, 1982; Boillot et al., 1984) may have contributed to
981 hide part of the eastern prolongation of the CZ acting as reverse during the late stages of arc
982 closure. The CZ may however continue in the island of Menorca, where pre-orogenic
983 Devonian and synorogenic Carboniferous sequences occur (Obrador et al., 2004).

984 North of the North Pyrenean Fault (NPF), the basement of the Aquitaine Basin is
985 known from boreholes. According to Paris and Le Pochat (1994), lithologic and faunal
986 evidences suggest affinity with the Pyrenees, the Montagne Noire and the Cantabrian
987 Mountains. The existence in Aquitaine of recumbent folds, cleavage and low grade
988 metamorphism suggests correlation with either the WALZ or CIZ. As there are no criteria to

989 distinguish between CIZ and WALZ in the northern branch of the IAA, both zones have been
990 merged as the continuation of the CIZ because it is wider, Ordovician orthogneisses are
991 common in the CIZ and the Axial Zone and absent in the WALZ, and tectonometamorphic
992 evolution of the Pyrenean Axial Zone can be compared with that of the CIZ.

993 A dextral displacement of 300 km along the NPF balances the appearance of the CZ in
994 the Cinco Villas Basque Massif, north of the NPF (Fig. 1). Its Permian cover yielded
995 paleodeclinations with a European orientation (NNE-SSW; Van der Voo and Baessenkool,
996 1973), while south of the NPF, paleodeclinations are NNW-SSE (Boillot et al., 1984). The
997 NPF occurs south of the Cinco Villas Massif (Gil Peña and Barnolas, 2004), and may be
998 continued north of the Quinto Real Massif (Fig. 2a), implying that the latter could have
999 remained in the Iberian side, while the northern Basque massifs formed part of the northern
1000 block. In Figure 1, the contour of the Basque Massifs is duplicated, but the darker color
1001 suggests the position of the northern and southern massifs. The 300 km of late Variscan
1002 dextral displacement nearly balances the 400 km of left lateral movement during the
1003 Cretaceous (Boillot et al., 1984)), so that the northern and southern massifs would have been
1004 relatively close during most of the Variscan deformation.

4. French and related massifs

1005 The main Variscan outcrops in France occur in the Armorican Massif and the Massif
1006 Central. Other massifs include the Axial Zone of the Pyrenees, Maures-Tanneron, Corse, the
1007 Vosges, a small part of the Rhenish Massif, which continues in Belgium, Luxemburg and
1008 Germany, and part of the External Massifs of the Alps, (Fig. 4). This section describes these
1009 massifs, except the Pyrenees, which were described in the Iberian Peninsula section, and the
1010 Rhenish Massif, essentially formed by the Rhenohercynian Zone and which will be described
1011 together with the Bohemian Massif. In addition, the Sardinia and the Schwarzwald are
1012 described here because of their close relation with Corse and the Vosges respectively. For a
1013 comprehensive text describing the Variscan geology of France, see the book edited by Keppie
1014 (1994). Compilations for the main massifs, mostly dealing with the Allochthon, can be found
1015 in the volume edited by Schulmann et al. (2014a).

1019 *4.1. Armorican Massif*

1
21020 The most salient characteristic of this massif is the transpressional type of tectonics.
3
41021 Strike-slip faults and shear zones, mostly right-lateral, cut across the Massif and the more
5
61022 important of them separate three domains named North, Central and South. Only the
7
81023 Autochthon exists in the North and Central domains, with the exception of a narrow band of
9
101024 Parautochthon in the latter, whereas the South Armorican Domain is largely occupied by the
111025 Allochthon.
12

13
141026
15
161027 *4.1.1. Central Armorican Domain (CAD)*
17

181028 This domain is bounded by the North Armorican Shear Zone, whereas its southern
19
201029 boundary is usually located at the Nort-sur-Erdre Fault (Fig. 4), as it represents the northern
21
221030 limit of the three groups forming most of the Allochthon in the South Armorican Domain
23
241031 (Ballèvre et al., 2014). So defined, the CAD is a WNW-ESE striking band 45-90 km wide.
25

261032 To the north of the Nort-sur-Erdre Fault, the St. Georges-sur-Loire Unit includes
27
281033 Upper Ordovician to Middle Devonian metasediments (Lardeux and Cavet, 1994), whose
29
301034 prevailing lithologies are low-grade schists, often with greenish and purple colors. Also
31
321035 frequent are black cherts and volcanics (spilites, basic tuffs, quartz-porphyrries) which show
33
341036 little continuity. Cartier and Faure (2004) describe this unit as formed by two allochthonous
351037 subunits. The southern subunit includes olistostromes and exotic blocks of black cherts,
36
371038 limestone, sandstone, acid and basic volcanics, of Silurian and Devonian ages, embedded in a
38
391039 pelitic or sandy matrix. The age of sedimentation is unknown, but must be more recent than
40
411040 the younger limestone blocks, which are Middle Devonian. Metamorphism is of greenschist
42
431041 facies, close to blueschist facies (Rolet et al., 1994). The northern subunit consists of
44
451042 sandstone and pelite alternation with a thick quartzitic level, and its age is probably
461043 Ordovician. Thrusting of the unit upon the northern autochthonous CAD is established at 360-
47
481044 350 Ma (Rolet et al. 1994).
49

501045 To the north of the St. Georges-sur-Loire Unit, the CAD is represented by
51
521046 Neoproterozoic (Brioverian) strata, representing a foreland where the effects of the Cadomian
53
541047 orogeny are limited to gentle folding (Trautmann et al., 1994), and by Paleozoic pre- and
55
561048 synorogenic sequences. No Cadomian or Cambrian intrusions, neither pre-Variscan
57
581049 deformation or metamorphism are recognized, and the sequence is thick and terrigenous,
59
601050 often bearing clasts of black cherts, and including some carbonates and volcanic alternations
61

1051 (Chantraine et al., 1994a, b). The Brioverian is unconformably overlain by Lower Ordovician
1052 detritals, starting with continental red beds and overlain by the Armorican Quartzite. The
1053 Ordovician is complete and culminates with Hirnantian glaciation-related deposits (Vidal et
1054 al., 2011; Dabard et al., 2021). A few Ordovician granitic plutons dated at 495-450 Ma occur
1055 in this domain. The Silurian is also rather complete, and characterized by the abundance of
1056 black shales, and the Devonian is marine, with abundant silts and quartzites, often
1057 ferruginous, alternating with shales and limestones (Robardet et al., 1994).

1058 Synorogenic early-middle Carboniferous deposits lie unconformably upon the
1059 Paleozoic sequence, although the latter is often fully preserved. Weak folding and erosion
1060 preceded Carboniferous volcanism and deposition, which allowed Stille (1928) to define the
1061 Bretonnic phase based in the maps of the central Brittany fold set (Paris et al., 1982). The
1062 main carboniferous outcrop of the CAD occurs in the NW, at the core of the large E-W
1063 Chateaulin Syncline. It is terrigenous with volcanic alternations and evidence of syntectonic
1064 deposition highlighted by exotic blocks, Carboniferous platform carbonates olistoliths and
1065 mudflows. Its age is Tournaisian to early Namurian (Pelhate, 1994).

1066 Variscan deformation is dominated by a set of late Variscan folds slightly oblique to
1067 the limits of the domain, which are dextral strike-slip faults. Early pre-Carboniferous folds are
1068 responsible for the Bretonnic unconformity, but they are open, scarce, and their geometry is
1069 mostly unknown. Some NE- and NW-trending folds in the Crozon Peninsula, at the western
1070 termination of the CAD may correspond to this phase. Also, the interference pattern shown by
1071 the central Brittany fold set may result from superposition of roughly E-W late folds on a set
1072 of previous, NNE-trending folds (Fig. 4).

1073 The main Variscan deformation in the CAD has been interpreted as resulting from
1074 crustal-scale simple shear deformation (Gumiaux, 2003; Gumiaux et al., 2004). Shearing was
1075 dextral (Figs. 4 and 5a), and generated the late Variscan “en échelon” folds subsequently
1076 flattening and rotating them toward the flow plane of regional shearing. A statistical estimate
1077 of cleavage orientation (N 95°-100° E) and its relation with the shear plane (N 125° E) yields a
1078 mean angle of 26°, equivalent to a shear value $\gamma = 1.5$, equivalent to a shear angle of 56°
1079 (Gumiaux et al., 2004). No significant vertical stretching occurred. Assuming that
1080 deformation was only by simple shear, this implies 50 % shortening perpendicular to the
1081 cleavage.

1082 Metamorphism is generally of low to very low grade, although P-T conditions increase
1083 to the west, where the amphibolite facies is reached, and south of the CAD, close to the
1084 northern branch of the South Armorican Shear Zone. The baric type is of low- to
1085 intermediate-P type with kyanite occurrences in shear zones, and of high-T around granite
1086 plutons, associated to which, migmatitic gneisses may occur (Rolet et al., 1994).

4.1.2. North Armorican Domain (NAD)

1087 This domain, bounded by the North Armorican Shear Zone, is interesting for the
1088 occurrence of an old, Paleoproterozoic basement, abundant evidences of the Cadomian
1089 orogeny, and three main Paleozoic unconformities. The Icartian basement crops out at the
1090 outermost coastal regions to both sides of the Saint-Malo Gulf and in the outer Channel
1091 Islands. It includes granodioritic and monzogranitic orthogneisses often migmatitic, as well as
1092 banded gneisses of sedimentary and volcanic origin. Their ages, ranging 1.8-2.2 Ga, point to a
1093 Gondwanan provenance (Chantraine et al., 1994b).

1094 The sedimentary and volcanic succession is Neoproterozoic and collectively named
1095 the Brioverian (Barrois, 1899). It is made of two groups, the lower of which bears bedded
1096 black cherts whereas the upper group includes this lithology only as clasts. The lower
1097 Brioverian is profusely intruded by Cadomian plutons dated at 670-580 Ma, and is interpreted
1098 as developed in an active margin on which three SW-NE trending subdomains have been
1099 recognized from NW to SE: a magmatic arc, a back-arc and a continental margin. The latter is
1100 characterized by the abundance of bedded black cherts and tholeiitic volcanics (Bale and
1101 Brun, 1983; Chantraine et al., 1994a, c), and its development was contemporaneous with arc
1102 magmatism. The upper Brioverian was deposited at *c.* 580-540 Ma and intruded by
1103 postorogenic 540-520 Ma plutons. Voluminous volcanism, mostly rhyolitic and ignimbritic,
1104 characterizes the Proterozoic-Paleozoic boundary, and has been dated between 545-520 Ma.

1105 The main Cadomian orogenic phase occurred at 600-580 Ma, and zonation of the
1106 deformation includes a large area with vertical cleavage and late-tectonic plutonism (540 Ma)
1107 occupying a triangular zone between the Granville-Cancale Fault and the North Armorican
1108 Shear Zone (Fig. 4). The SW-NE trending band to the north includes, from SE to NW, a zone
1109 of high-T metamorphism with migmatites, another characterized by thrusting and sinistral
1110 wrenching, with 615-585 Ma syntectonic plutonism, and inside the latter to the north, areas of
1111 no cleavage with post-orogenic plutonism (Rabu et al., 1994).

1114 Thick terrigenous early Cambrian deposits unconformably overlay the Brioverian in
1115 the NE, alternating with ignimbrites, rhyolites and scarce andesites, as well as carbonates.
1116 These are in turn unconformably covered by Lower Ordovician red beds deposited in deltaic
1117 environments and by the Armorican Quartzite, which may show a gradual transition from the
1118 red beds or lay directly upon the Cambrian or the Brioverian (Robardet et al., 1994). No
1119 Ordovician granitic plutons occur in the NAD. The rest of the preorogenic succession does
1120 not differ significantly from that of the CAD, although ages of the detrital zircons in
1121 Ordovician strata differ notably (Dabard et al., 2021). The third significant unconformity is
1122 marked by the synorogenic Carboniferous deposits, characterized by an initial extensional
1123 phase marked by rhyolites and ignimbrites, and by conglomerates, sandstones, mudstones,
1124 coal beds and platform limestones, reflecting a paralic environment and reaching the early
1125 Namurian (Pelhate, 1994).

1126 In the Pays de Léon, to the north of the westernmost Armorican Massif, high-P rocks
1127 have been identified in what seems a different domain or zone. There, a lower unit of coarse-
1128 grained augengneisses, probably Paleozoic, is the footwall to an upper unit of ortho- and
1129 paragneisses, migmatized and with lenses of eclogite at its base. High-P metamorphism was
1130 dated not very precisely around 440-415 Ma (Cabanis and Godard, 1987; Paquette et al.,
1131 1987; Ballèvre et al., 1994). Later, pervasive migmatization gave rise to a gneiss dome (Rolet
1132 et al., 1994). This unit has been interpreted as probably derived from the South Armorican
1133 Domain, displaced at least 150 km to its present position by dextral transcurrent shear zones
1134 (Balé and Brun, 1986), as representing the Saxothuringian Zone (Schulz et al., 2007), and as
1135 an equivalent to the Mid German Crystalline High (Faure et al., 2010a).

1136 Except for the Léon, Variscan reworking was weak in the NAD. North of the Gravelle-
1137 Cancale Fault, deformation started during the Upper Devonian by remobilization of Cambrian
1138 faults, which produced WSW-ENE folds in the north and continued during the early
1139 Carboniferous by NNE and SSW-verging thrusts and N 110° E folds. Folds in the north are
1140 locally associated with a strong, greenschist facies cleavage (Rolet et al., 1994). The
1141 triangular zone between the Gravelle-Cancale Fault and the North Armorican Shear Zone
1142 (Fig. 4) is characterized by Cadomian late-tectonic plutonism and vertical cleavage. Variscan
1143 deformation there is so weak that a N-trending dolerite dike swarm (Le Gall, 1999), early
1144 Carboniferous and perpendicular to the late Variscan folds, is undeformed.

1146 4.1.3. South Armorican Domain (SAD)

1
2 1147 Most of the Armorican Massif SW to the Nort-sur-Erdre Fault is occupied by
3
4 1148 allochthonous units which have been grouped in three ensembles that correlate with
5
6 1149 equivalents in NW Iberia (Ballèvre et al., 2014 and references therein). The Allochthon crops
7
8 1150 out in a triangular zone between the Nort-sur-Erdre Fault and the southern branch of the
9
10 1151 South Armorican Shear Zone, and also as klippen south of the latter (Fig. 4). The original
11 1152 contacts between units are interpreted as thrust faults, although in many cases, including the
12
13 1153 contact with the Autochthon, they were overprinted by extensional detachments.
14

15 1154 The Autochthon consists of Cambrian to Devonian metasediments and metavolcanics
16 1155 in the SE and mostly schists, granitic orthogneisses and migmatites toward the NW. In the
17 1156 south (Vendée littorale), the sequence includes from bottom to top migmatites and high-grade
18
19 1157 gneisses, a thick succession of Cambro-Ordovician aluminous schists with rare carbonates,
20
21 1158 Lower Ordovician porphyroids, thin quartzite layers and monotonous grey slates. Above, and
22
23 1159 separated by a thrust fault, synorogenic deposits, low-grade schists, ampelites and black
24
25 1160 cherts probably represent the Parautochthon (Ballèvre et al., 2012). Lower Ordovician felsic
26 1161 magmatism in the Autochthon and Parautochthon is interpreted by Pouclet et al. (2017) as
27
28 1162 related with a continental rift where magmatic underplating produced abundant crustal
29
30 1163 melting, and which can be followed from central and northern Iberia to southern France.
31
32 1164

33
34 1164 Above, a thrust sheet with Lower Ordovician porphyroids is still considered the
35
36 1165 Parautochthon, and occurs beneath the Lower Allochthon and the high-P Middle Allochthon
37
38 1166 represented by the Bois-de-Cené Unit (Figs. 4 and 5a). Migmatization of the lower parts of
39
40 1167 the Autochthon postdate emplacement of the allochthonous pile, and was overprinted by the
41
42 1168 southern branch of the South Armorican Shear Zone.
43

44 1169 The Lower Allochthon is well represented in the Champtoceaux Complex, along the
45
46 1170 Vendée coastline to the south, and in poorly exposed klippen NW of the latter (Figs. 4 and
47
48 1171 5a). The better exposures occur at the core of the antiform where the Champtoceaux Complex
49
50 1172 crops out. It consists of a thick sequence of albite-bearing micaschists derived from late
51 1173 Neoproterozoic greywackes, fine-grained leucocratic gneisses with eclogite lenses, garnet
52
53 1174 micaschists, and graphitic schists. Leucocratic orthogneisses of calc-alkaline and alkaline
54
55 1175 composition were dated at *c.* 495-475 Ma (Ballèvre et al., 1987, 1989, 2014). The eclogite
56
57 1176 protoliths are doleritic dikes intruding the Lower Ordovician granitoids and felsic volcanics
58
59 1177 (Godard, 1988).
60
61
62
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64
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1178 Early Variscan high-P metamorphism is recorded in eclogites, orthogneisses and felsic
1179 gneisses. In the Champtoceaux Complex, an along-strike gradient shows increasing
1180 metamorphic conditions toward the west (from 1.5-2 GPa to *c.* 2.5 GPa; Godard et al., 1981;
1181 Ballèvre and Marchand, 1991; Bosse et al., 2000), indicating continental subduction toward
1182 the NW in present coordinates. High-P metamorphism has been dated at *c.* 360 Ma in a
1183 glaucophane-bearing eclogite by U-Pb and Sm-Nd methods (Paquette, 1987; Bosse et al.,
1184 2000; Paquette et al., 2017). Ar-Ar data on phengite yielded *c.* 350-340 Ma (Bosse et al.,
1185 2000; Maurel et al., 2003), probably reflecting decompression, which took place during
1186 stacking of the units and thrusting onto the Autochthon. This stage was associated with
1187 inverted metamorphism, followed by fast cooling at 340-330 Ma (Pitra et al., 2010).

1188 The Middle Allochthon includes a group of Cambro-Ordovician oceanic units and at
1189 least an Upper Devonian unit. The older group is exposed in the Audierne Bay, Ile-de-Groix
1190 and Vendée littorale (Bois-de-Cené Unit; Figs. 4 and 5a). In the Audierne Bay Complex, a
1191 metaophiolite unit includes serpentized peridotites, metagabbros and greenschists that most
1192 probably derive from basaltic flows and tuffs (Peucat, 1973; Peucat and Cogné, 1974;
1193 Hanmer, 1977). The basic rocks belong to a tholeiitic suite with MORB compositions
1194 (Bernard-Griffiths and Cornichet, 1985; Lucks et al., 2002). The unit has a distinct magnetic
1195 signature that can be followed offshore to the SW for more than 100 km (BRGM, 1972; de
1196 Poulpiquet, 1988). A plagioclase-rich metagabbro of the ophiolite was dated at 482 Ma
1197 (Paquette et al., 2017). The rocks were intensely deformed under amphibolite facies
1198 conditions.

1199 Other rocks of oceanic affinity include the blueschist-facies metabasites and
1200 micaschists of Ile-de-Groix and Bois-de-Cené units. The latter unit lies above the Lower
1201 Allochthon occupying the core of a synform, while the structural position of Ile-de-Groix is
1202 unknown. The metabasites derive from basaltic lava flows or sills that underwent extensive
1203 alteration by oceanic hydrothermalism, but E-MORB affinities can be recognized (Bernard-
1204 Griffiths et al., 1986; El Korh et al., 2009). Oceanic sediments include cherts and
1205 manganiferous nodules. The age of this sequence is constrained by rare layers of fine-grained
1206 felsic gneisses which have provided a zircon age of 481 Ma (El Korh et al., 2012). The
1207 metamorphic history is dominated by blueschist-facies metamorphism (Ballèvre et al., 2003;
1208 El Korh et al., 2009). The metamorphic zonation in the Ile-de-Groix (Triboulet, 1974) has
1209 been interpreted as due to the superposition of two units that slightly differ in their P-T paths
1210 (Bosse et al., 2002; Ballèvre et al., 2003). Rb-Sr and Ar-Ar data indicate that the high-P event

1211 took place 370-360 Ma ago, while cooling was achieved in rocks overprinted by greenschist-
1212 facies metamorphism at *c.* 350-340 Ma (Bosse et al., 2005). U-Pb data on titanite confirmed
1213 the Late Devonian age of the high-P event (El Korh et al., 2012).

1214 A narrow belt of oceanic rocks occurs in Vendée east of the Bois-de-Cené Unit
1215 (Ballèvre et al., 2009). Its composition ranges from amphibolites of basaltic composition to
1216 rhyodacitic leucocratic banded gneisses. Geochemically, the suite follows a tholeiitic trend,
1217 and has been interpreted as a volcanic sequence perhaps erupted in an island-arc or forearc
1218 setting (Thiéblemont et al., 1987a). Metamorphism is characterized by the lack of eclogite-
1219 facies relics and peak P-T conditions at *c.* 0.7 GPa at 470-550 °C (Thiéblemont et al., 1988).
1220 The ages of the protoliths and their metamorphism have not been investigated, but they are
1221 tentatively considered Early Ordovician in age (Ballèvre et al., 2014). Another possible
1222 ophiolitic unit of unknown age is intercalated in the Lower Allochthon of the Champtoceaux
1223 Complex (Figs. 4 and 5a).

1224 A Devonian ophiolitic unit bounds the Champtoceaux Complex, separating the Lower-
1225 from the Upper Allochthon. In the east, it is made of serpentized peridotites and
1226 metagabbros. Most metagabbros have been severely deformed under amphibolite facies
1227 conditions, but some of them preserve magmatic textures and mineralogy. The metagabbros
1228 show compositions of cumulates or of N-type MORB (Bernard-Griffiths and Cornichet, 1985;
1229 Paquette, 1987), and one of them was dated as Upper Devonian (*c.* 380 Ma; Paquette, 1987;
1230 Paquette et al., 2017). Toward the west, the unit changes into greenschists and micaschists
1231 possibly representing the upper part of the oceanic crust.

1232 The Upper Allochthon in the SAD includes a lower group of units characterized by an
1233 eclogite-facies event and an upper group of low-grade Cadomian basement unconformably
1234 overlain by Cambrian to Carboniferous sediments. The best example of the high-P group is
1235 the Essarts Unit, cropping out SW of the southern branch of the South Armorican Shear Zone
1236 (Figs. 4 and 5a). This unit consists of eclogites, orthogneisses and migmatized paragneisses.
1237 The eclogites derive from metagabbros forming a MORB-type tholeiitic suite (Godard, 1988,
1238 2001) and interlayered banded gneisses derived from plagiogranites point to an oceanic crust
1239 provenance (Montigny and Allègre, 1974; Godard, 1983, 1988; Bernard-Griffiths and
1240 Cornichet, 1985). Magmatic zircons from an eclogite provided a protolith age around 490 Ma
1241 (Paquette et al., 2017). The unit also contains orthogneisses, one of them dated at 483 Ma
1242 (Lahondère et al., 2009). Moreover, polycyclic gneisses record a high-T metamorphism
1243 predating the eclogite-facies event (Godard, 2001, 2009). The early high-temperature event is

1244 recorded by migmatitic, cordierite-bearing gneisses that also show evidences of cooling. A
1245 first generation of monazite is associated with the high-T event and was dated at 460 Ma, no
1246 far from protolith ages of gabbros and granitoids, to which it is possibly related. A younger
1247 population gave 375 Ma, potentially dating the high-P event or later exhumation of the unit.

1248 An eclogitic unit also occurs in the Audierne Bay Complex, where a few tectonic
1249 slices display garnet-diopside mafic granulites, eclogites and garnet-kyanite gneisses (Velde,
1250 1972; Marchand, 1982; Lucks et al., 2002). They occur on top of the Cambro-Ordovician
1251 oceanic unit and are separated from it by a strongly deformed orthogneiss which may occur at
1252 a ductile extensional detachment (Ballèvre et al., 1994). All slices record a high-P/high-T
1253 metamorphism (*c.* 1.8-2.0 GPa, 800-900 °C). U-Pb data on zircons (Peucat, 1983; Paquette et
1254 al., 1985) suggest an Early Ordovician age for the granulite protoliths (485-480 Ma) and the
1255 orthogneiss (480-470 Ma). The high-P event is constrained at 390-380 Ma.

1256 Another unit has been recognized in the Audierne Bay Complex, consisting of
1257 metagreywackes and interbedded amphibolites. The age of the protoliths is unknown, and the
1258 metamorphic grade is of the albite-epidote-amphibolite facies. The affinity of this unit is still
1259 unclear, but due to its structural position above the ophiolite complexes, it can be tentatively
1260 ascribed to the Upper Allochthon and correlated with the intermediate-P units of NW Iberia
1261 (Ballèvre et al., 2014).

1262 The upper group of the Upper Allochthon in the SAD includes two low-grade units,
1263 one overlying the Champtoceaux Complex (Mauges Unit) and the other above the Essarts
1264 Unit (Roc-Cervelle). The former displays a Proterozoic basement made of low-grade schists
1265 with a layer of black cherts associated with mafic volcanics. An increasing grade of
1266 metamorphism is recorded towards the contact with the underlying Devonian ophiolitic unit,
1267 where a sequence of mylonitized amphibolite-facies gneisses 1 km thick occurs.

1268 In the southern part of the Mauges Unit, the Paleozoic cover consists of Cambrian
1269 conglomerates and schists (Cavet et al., 1966; Ducassou, 2010) overlain and partly intruded
1270 by a complex of mafic to felsic volcanic rocks dated at *c.* 520 Ma (Thiéblemont et al., 1987b,
1271 2001). Along the northern boundary, the Neoproterozoic basement is unconformably overlain
1272 by Lower Ordovician reddish siltstones (Cavet et al., 1971). Cambrian volcanics (Perroud and
1273 van der Voo, 1985) and Lower Ordovician siltstones (Perroud et al., 1986) have provided
1274 paleomagnetic evidence of high southerly latitudes, consistent with a northern Gondwana
1275 setting. Above lies unconformably a Lower-Middle Devonian formation starting with

1276 sandstone, following by reefal limestone and culminating with immature sandstones rich in
1277 plant debris (Ducassou et al., 2009). A more distal sequence includes Upper Ordovician
1278 sediments displaying a typical glaciomarine record covered by Silurian graptolite-rich cherts
1279 (Piçarra et al., 2002, 2009), pelites and Emsian pelagic carbonates. Synorogenic sediments
1280 exist in the north: early Carboniferous lacustrine and alluvial deposits laid down in a
1281 transtensional basin (Ballèvre and Lardeux, 2005; Ducassou, 2010).

1282 Another Upper Allochthon low-grade unit (Roc-Cervelle) is located SE of the Essarts
1283 Unit and consists of low-grade schists perhaps Cadomian (Bouton and Branger, 2007; Bouton
1284 and Camuzard, 2012). A Devonian sequence cropping out nearby might have been deposited
1285 on top of the schists. It includes from bottom to top Givetian fluvatile sandstones, shales and
1286 reefal limestones (Le Maître, 1937), recording a marine transgression close to an emerged
1287 land. The low-grade units of the Upper Allochthon in the SAD record the sedimentary
1288 evolution at the beginning of the convergence. Because the sediments contain abundant
1289 volcanic-derived material, it is possible that the Upper Allochthon were part of a continental
1290 magmatic arc during late Lower-Middle Devonian (Ducassou et al., 2014).

1291 4.2. *French Massif Central (FMC)*

1292 The FMC is essentially formed by allochthonous terranes, while the Autochthon crops
1293 out only to the south, in the Montagne Noire region. Lardeux et al. (2014) correlate this
1294 Massif with that of Bohemia and include its allochthonous part in the Moldanubian Zone. The
1295 following description is based on their overview and that of Faure et al. (2009), and proceeds
1296 describing the structural pile from bottom to top.

1297 A fold and thrust belt affecting Cambrian to early Carboniferous sediments and
1298 igneous rocks developed in the Montagne Noire region, which is considered the continuation
1299 of the Pyrenean Axial Zone. It consists of Cambrian to middle Carboniferous low-grade
1300 metasediments representing a passive margin. Sediments are mostly terrigenous except for a
1301 thick carbonate platform developed during the Middle Devonian to early Carboniferous. In
1302 the north, the structure consists of large, south-vergent recumbent folds and thrust faults
1303 deformed in the Axial Zone of the Montagne Noire by two large, adjacent flattened gneiss
1304 domes whose emplacement is linked with a pull-apart structure created during dextral
1305 transcurrence (Franke et al., 2011). The domes consist essentially of Cambro-Ordovician
1306 orthogneisses pervasively migmatized. Protoliths have yielded ages between 550 and 450 Ma,
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1308 and low-P/ high-T metamorphism is dated at 340-320 Ma (Cocherie et al., 2005; Faure et al.,
1309 2010b, 2014). An earlier high-P event is represented by eclogites resulting from the
1310 metamorphism of basic dikes. High-P metamorphism has been dated at *c.* 355-350 Ma, and
1311 interpreted as the result of intracontinental subduction (Faure et al., 2014). Large recumbent
1312 folds developed between 340-330 Ma in the Paleozoic series, and were later deformed by
1313 doming and flattening related to transpression at 330-315 Ma. Migmatization in the Axial
1314 Zone took place between 320-295 Ma (Poujol et al., 2017) or 315-303 Ma (Trap et al., 2017).

To the south, a synorogenic turbiditic basin developed during the Visean-Namurian
and can be continued under the recent cover until the Pyrenees. The basin was affected by
recumbent folds and thrust faults with a NE vergence contemporaneously with its
sedimentation, and large olistoliths were emplaced in proximal turbidite facies during the
deformation (Engel et al., 1980; Chardon et al., 2020).

Above the Autochthon, a thrust sheet of low-grade metasediments represents the
Parautochthon. It crops out in the SE of the FMC, and in tectonic windows on its western and
northern parts (Fig. 4). It consists of schists, quartzites, rare limestones and felsic and mafic
metaigneous rocks (Lardeaux et al., 2014). Metamorphism is Carboniferous and generally of
low grade (greenschist to epidote-amphibolite facies), although anatexis was locally reached
in areas intruded by voluminous Variscan granites.

The Parautochthon is overlain by the Lower Gneiss Unit (LGU), whose lithology is
similar to that of the underlying unit but it is metamorphosed mostly under amphibolite-facies
conditions. Cambro-Ordovician magmatism is tholeiitic or alkaline, similar to that of the
Parautochthon, and derived from the northern Gondwana passive continental margin (Briand
and Santallier, 1994). Basic and ultrabasic rocks are rare except in the Limousin (NW Massif
Central). Relics of high-P granulites and eclogites have been found in some units of the
western part of the FMC (Ledru et al., 1994a). Barrovian metamorphism reaching the
sillimanite zone is identified in several units and with different ages, related with the
diachronous evolution of the nappe pile. It is 360-355 Ma old in the west and 340-325 Ma in
the south (Costa, 1989; Costa et al., 1993; Ledru et al., 1994b), clearly corresponding to
different pulses. Partial melting and anatexis affected large parts of this unit, and is dated at *c.*
375-365 Ma in the west. There, it might have followed a high-P event, being in turn followed
by the widely recognized Barrovian imprint.

1339 An ophiolitic unit has been identified above the LGU in the Limousin (Girardeau et
1340 al., 1986, 1994). It consists of harzburgites, dunites, wehrlites, troctolites and layered gabbros
1341 mostly undeformed, and metamorphosed under low-P (0.2-0.4 GPa) static conditions,
1342 probably by oceanic hydrothermal metamorphism (Berger et al., 2005). The ophiolites occur
1343 in dismembered massifs 1-5 km long and up to 1 km thick in a *mélange* on which chlorite-
1344 rich mylonites can be identified. The authors interpret the occurrence as marking a suture
1345 between upper and intermediate allochthonous units. Lack of pervasive deformation and of
1346 high-P metamorphism allow excluding them from the Cambro-Ordovician oceanic group
1347 while a Devonian age is considered possible.

1348 The Upper Gneiss Unit (UGU) includes at its base the Leptyno-Amphibolitic Complex
1349 (Forestier, 1963; Santallier et al., 1988), an assemblage of spinel and garnet peridotites and
1350 the metamorphic by-products of gabbros, basalts, granites, rhyolites, tuffs and sediments.
1351 Igneous protoliths are Cambro-Ordovician and high-P granulite/eclogite-facies
1352 metamorphism is Lower-Middle Devonian (415-380 Ma; Pin and Lancelot, 1982; Paquette et
1353 al., 1995; Lotout et al., 2018). This was followed by decompressive partial melting at 385-370
1354 Ma which ended under amphibolite-facies conditions (Mercier et al., 1991; Faure et al., 2008;
1355 Lotout et al., 2018). The high-P rocks occur at the base of Leptyno-Amphibolitic Complex,
1356 where dismembered peridotites appear at several places, often associated with ultrahigh-P
1357 eclogites equilibrated at 2.5-3.0 GPa at 412 Ma (Berger et al., 2010).

1358 The Leptyno-Amphibolitic Complex has been interpreted as representing an oceanic
1359 lithosphere, a subduction *mélange*, a back-arc basin and a transitional continent-ocean
1360 lithosphere (Lotout et al., 2018 and references therein). Several of these hypotheses, if not all,
1361 are possibly correct, as the complex may include units of different origins and evolutions.
1362 Perhaps the 35 Ma age interval found for high-P metamorphism can be explained by these
1363 differences. We will consider that at least some units of the complex form the base of the
1364 Upper Gneiss Unit and represent a subducted and exhumed part of a Cambro-Ordovician
1365 terrane with a transitional but essentially continental crust, which continues upward by an
1366 assemblage of paragneisses and orthogneisses that underwent Middle-Upper Devonian
1367 migmatization but apparently not the earlier high-P event.

1368 The uppermost part of the nappe pile includes several units in different parts of the
1369 Massif Central (Fig. 5b, c). In the SW, the Thiviers-Payzac Unit is formed by low- to medium
1370 grade Cambrian greywackes, quartzites and rhyolites. Surrounded by this unit, Saint Génis is
1371 either an *in situ* small ophiolitic unit including greenschist-facies metabasalts, cherts and

1372 Middle Devonian limestones, or an early Carboniferous olistostrome reworking a Middle
1373 Devonian ophiolite (Faure et al., 2009).

1374 In the NE, the uppermost units include Middle-Upper Devonian calc-alkaline
1375 volcanics forming the Morvan Arc (Pin et al., 1982; Lardeaux et al., 2014), and to the south,
1376 the Brévenne Unit, an Upper Devonian ophiolite metamorphosed under greenschist-facies
1377 conditions representing its back-arc. The latter is dated at 370-360 Ma (Pin and Paquette,
1378 1998), and was deformed between 350-340 Ma (Leloix et al., 1999). Even younger, the so-
1379 called Tuffs Anthracifères are formed by late Visean terrigenous and coal sediments
1380 alternating with rhyolites and dacites unconformable upon metamorphic and granitic rocks.
1381 The calc-alkaline character of Late Devonian to early Carboniferous magmatism and its
1382 arc/back-arc polarity reflects south-dipping subduction (Lardeaux et al., 2014).

1383 The tectonometamorphic evolution of the FMC includes an early Variscan episode of
1384 subduction, Lower-Middle Devonian in age although not very precisely dated (415-380 Ma),
1385 recognized in the UGU. The conditions reached 1.8-2 GPa at 650-750 °C, although a
1386 minimum ultrahigh-P of 2.8 GPa was identified in the Lyonnais (Lardeaux et al., 2001). By
1387 380-360 Ma this unit was exhumed to *c.* 30 km depth by underthrusting or subduction of the
1388 LGU, and was followed by migmatization of the latter at *c.* 375-365 Ma, younger than in the
1389 UGU (385-370 Ma). An ocean possibly separated both units, although ophiolitic relics are
1390 very scarce in the FMC. The intrusives of the tonalitic line of Limousin, emplaced at 380-360
1391 Ma, were probably generated during its closure, as they occur only in the UGU (Figs. 4 and
1392 5b). The same accounts for the Morvan Arc and the Saint Génis and Brévenne ophiolitic
1393 units.

1394 According to Faure et al. (2009), the next event was Upper Devonian?-early
1395 Carboniferous metamorphism characterized by Barrovian-type conditions, indicating that a
1396 collisional regime had replaced previous subduction-related convergence. Recumbent folds
1397 and thrust sheets were emplaced between 360-345 Ma producing significant crustal
1398 thickening. This event was succeeded by middle Carboniferous (Visean) synmetamorphic
1399 deformation recognized in the southern part of the FMC, characterized by southward vergence
1400 and dated at 340-335 Ma, which propagated to the south until 325 Ma. Subsequently, late
1401 Carboniferous (Namurian-Westphalian) extension dated at 325-295 Ma allowed the intrusion
1402 of large plutons with fabrics indicating NW-SE crustal stretching which was however coeval
1403 with continued compression to the north of the FMC and in the Montagne Noire and
1404 Pyrenees.

1405 Kinematic criteria indicate tectonic transport toward the west (NW to SW) for early
1406 Variscan fabrics (older than 350 Ma), and toward the SW-SSW for younger fabrics (Ledru et
1407 al., 1994b). The older were formed under amphibolite-facies conditions, and often represent
1408 retrogressive foliations following high-P, subduction-related metamorphism, and considered
1409 coeval with nappe emplacement. The younger SW-SSW kinematic criteria occur mostly in
1410 the southern part of the FMC according to Ledru et al. (1994b), and are distributed along the
1411 whole massif for Faure et al. (2008). They are also related to nappe emplacement, including
1412 that of the Allochthon above the Parautochthon and Autochthon, and that of the thrust sheets
1413 of the northern Montagne Noire.

1414 Transpression characterizes some of the deformation episodes and was partly coeval
1415 with local extensional structures. Dextral shear zones, roughly striking WNW-ESE, occur in
1416 continuation of those in the Armorican Massif. To the east their strike changes to SW-NE due
1417 open oroclinal folding that formed the Massif Central Arc (MCA). A fold system associated
1418 to these faults is dated at 345-335 Ma (Gay et al., 1981; Costa et al., 1993). To the east, the
1419 Velay Dome witnesses a late-Variscan low-P/high-T event dated at 320-295 Ma (Ledru et al.,
1420 2001; Lardeaux et al., 2001), while a ductile detachment bounding the dome in the north is
1421 dated at 320 Ma (Malavieille et al., 1990). W-E sinistral strike-slip faults occurred also in the
1422 northern part of the FMC between 320-300 Ma (Ledru et al., 1994b), either contemporaneous
1423 with a long lasting dextral system or intercalated among two separate dextral events.

4.3. *Vosges and Schwarzwald Massifs (VM, SWM)*

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1437 migmatitic paragneisses and includes amphibolite and marble intercalations, while the
1438 Granulite Unit derives from a *c.* 500 Ma igneous protolith. In the two latter, initial
1439 metamorphism was of high-P/high-T type, and all units evolved later to low-P/high-T
1440 conditions. The three units include lenses of peridotite, while amphibolites possibly
1441 retrograded after granulites alternate with felsic rocks at the boundary between the Varied and
1442 Granulite units, which together may be equivalent to the Leptyno-Amphibolitic Complex of
1443 the FMC. The high-T rocks have been interpreted as vertical extrusions at mid-crustal levels,
1444 contemporaneous with ultra-potassic magmatism (Schulmann et al., 2002; Skrzypek et al.,
1445 2014).

1446 The Southern Vosges include a relative autochthon with Upper Devonian limestone
1447 overlain by early Carboniferous flysch and continental sediments and volcanics intruded by
1448 ultra-potassic rocks dated at 345-335 Ma (Tabaud et al., 2015). Above, an allochthonous unit
1449 consists of an Upper Devonian ophiolite covered by thick Viséan flysch deposits. The
1450 ophiolites, known as the Ligne des Klippes (Schneider et al., 1990), and dated at *c.* 372 Ma,
1451 are equivalent to the Brévenne Unit of the FMC, and represent the opening of a small ocean,
1452 probably a back-arc basin (Skrzypek et al., 2012).

1453 In the Schwarzwald, the dextral Baden-Baden Shear Zone, the eastward continuation
1454 of the Lalaye-Lubine Shear Zone in the Vosges, separates the Moldanubian Zone from a
1455 narrow belt of Cambro-Ordovician metasediments and metaigneous rocks in greenschist to
1456 amphibolite facies representing the Saxothuringian Zone. The Moldanubian Zone includes the
1457 Central and Southern Schwarzwald Gneiss complexes separated by a belt of Lower Paleozoic
1458 to Carboniferous deposits. The Central Schwarzwald Gneiss Complex is formed by the same
1459 three units described in the Central Vosges, although some differences exist: the Monotonous
1460 Unit here includes eclogitic relicts indicating a high-P event previous to the low-P/high-T
1461 conditions displayed by the migmatitic paragneisses, dated at *c.* 330 Ma. Conversely, no high-
1462 P relics are preserved in the Varied Unit. The Granulite Unit reflects an igneous event of
1463 Cambrian-Lower Ordovician age, and includes metasediments of probable Silurian-Devonian
1464 age. The age of high-P metamorphism is estimated around 340 Ma (Kalt et al., 1994;
1465 Marschall et al., 2003).

1466 The metasedimentary belt separating the two gneiss complexes consists of Ordovician-
1467 Silurian metasediments, metavolcanics and deformed granitoids, but no metamorphic Lower
1468 Devonian to Upper Carboniferous sediments and volcanics. The latter are correlated with
1469 coeval rocks of the Southern Vosges. The Southern Schwarzwald Gneiss Complex is made of

1470 gneisses, high-P granulites and migmatites similar to those of the Varied Unit, overridden by
1471 a thrust sheet with migmatitic gneisses, amphibolite and gabbro boudins with protolith ages of
1472 350 Ma.

1473 The Variscan history of the VM and SWM is marked by the opening during the Upper
1474 Devonian of a narrow back-arc, its inversion during the early Carboniferous, high-P/high-T
1475 metamorphism and low-P/high-T overprint (Lardeaux et al., 2014). Although complex, this
1476 evolution reflects Devonian-Carboniferous subduction toward the SE below the Moldanubian
1477 crust (Edel and Schulmann, 2009; Skrzypek et al., 2014). According to the age of high-P
1478 metamorphism and that of the Ligne des Klippes ophiolite, the subducted ocean was the
1479 Saxothuringian, defined by Franke (2000) between the Saxothuringian Zone and Bohemia.

1480 1481 *4.4. Massifs of the southern Variscan Belt (MTM, Corsica, Sardinia, EmA)*

1482 The southern European Variscan belt includes several relatively small crystalline
1483 massifs whose tectonometamorphic and magmatic evolution is similar among them, and also
1484 to that of the large massifs (Corsini and Rolland, 2009). These are the Maures-Tanneron
1485 Massif (MTM), Corsica, Sardinia and the External Massifs of the western Alps (EmA).

1486 The MTM crops out along the Mediterranean coast south of the FMC, and consists of
1487 a nappe stack with N-S structures that define three domains. The following description is
1488 based on Corsini and Rolland (2009), and Schneider et al. (2014). In the Western Domain, the
1489 thrust sheets dip to the NW and from west to east include Silurian metapelites, sandstones,
1490 quartzite and limestone, a Cambrian volcano-sedimentary rift-related succession with alkaline
1491 rocks, and a Precambrian basement with orthogneisses and metasediments. The whole
1492 sequence is inverted by late Variscan back-thrusts, as demonstrated by sedimentary structures
1493 in an early Carboniferous synorogenic flysch (Bronner and Bellot, 2000). Metamorphism in
1494 this domain is of low- to medium grade, with temperature increasing toward the SE and
1495 reaching the sillimanite zone.

1496 The Central Domain, described as the Leptyno-Amphibolitic Complex by Schneider et
1497 al. (2014), comprises metapelites, quartzites and calc-schists, a bimodal magmatic suite with
1498 alkaline and tholeiitic affinities, and peridotites with high-P relicts. Metamorphic grade is
1499 higher than in the previous domain and partial melting is widespread. The structure is
1500 antiformal, a flattened gneiss dome with migmatites surrounding a syntectonic Variscan
1501 granite. The Eastern Domain consists of high-grade metasediments, metavolcanics and

1502 orthogneisses ranging in age from Neoproterozoic to Silurian, as well as abundant syntectonic
1503 Variscan granitoids. Some ultrabasics also occur, as well as Cambro-Ordovician tholeiitic
1504 metabasalts and metagabbro that Schneider et al. (2014) interpreted as a possible
1505 dismembered suture. Relicts in metabasic rocks record an early high-P metamorphic event.

1506 The structural evolution includes a high-P event poorly dated but probably Devonian,
1507 and whose P-T conditions have not been precisely established. It is identified in
1508 metaperidotites and retroeclogites of the Central and Eastern domains, and reflects eastward
1509 subduction in the internal zones. Barrovian metamorphism characterizes the Western Domain,
1510 where early Carboniferous recumbent folds and thrust sheets with NW vergence were
1511 overprinted by E-verging back-thrusts during the middle Carboniferous (Schneider et al.,
1512 2014). Some of these were reactivated as normal faults during the late Carboniferous, which
1513 is also characterized by granite intrusions and the formation of gneiss domes and late
1514 Carboniferous intramontane basins, all related with the Variscan orogenic collapse. Coeval
1515 with this stage, wrench structures more or less parallel to the trend of earlier ones developed.
1516 In the MTM, these are roughly N-S shear zones and faults with dextral motion which had first
1517 recorded a sinistral kinematics (Bellot et al., 2002).

1518 In Corsica, the pre-Alpine basement is mostly occupied by Variscan granites, but the
1519 metamorphic rocks represent the Allochthon (Fig. 4). In Sardinia, the Allochthon occupies the
1520 northern and central parts, while an Autochthon comparable to the Mediterranean peri-
1521 Gondwanan province is exposed in the south (Fig. 1). The Allochthon is divided in two zones.
1522 The Nappe Zone, in the centre, occupies most than half of the island surface, and is equivalent
1523 to the Western Domain of the MTM while the Axial or Internal Zone, in the north, is
1524 continuation of the Central and Eastern domains (Rossi et al., 2009; Schneider et al., 2014).
1525 Both zones are separated by the Posada-Asinara Shear Zone, trending NW-SE and interpreted
1526 as a dextral shear zone cutting across the thrust belt separating the Allochthon and the
1527 Parautochthon (Fig. 1). What follows is a short description of zones and units from bottom to
1528 top, based on Rossi et al. (2009) and Cocco et al. (2018).

1529 The Autochthon of SW Sardinia (Iglesiente) consists at its base of a terrigenous
1530 Cambrian to Lower Ordovician sequence including a thick Lower Cambrian carbonate unit,
1531 which is truncated by the Sardic unconformity. The ensemble is overlain by partially
1532 syntectonic Upper Ordovician clastics and followed by shallow-marine glaciation-related
1533 deposits including alkaline volcanics, overlain in turn by Silurian-Devonian black shales,
1534 limestones and Culm deposits. Paleozoic deposits are at most anchimetamorphic. The original

1535 position of the Iglesias in the Variscan puzzle is poorly constrained. It might be derived
1536 from a relatively remote domain significantly distinct from the Nappe Zone (Cocco et al.,
1537 2018).

1538 Overlying the Autochthon, the Nappe Zone is divided into external and internal parts.
1539 The External Nappes are a stack of units transported to the SW which include Cambrian to
1540 Lower Ordovician sandstones and acid volcanites dated at 490 Ma, unconformably overlain
1541 by thin conglomerates and a thick calc-alkaline volcanic sequence related to an Ordovician
1542 arc (460-450 Ma). The conglomerates lie above the Sardinian unconformity, and an Upper
1543 Ordovician transgression is marked by shallow-marine sandstones which continued by a
1544 typical Silurian succession with black shales, marls and limestones representing the whole
1545 Devonian and the lowermost Carboniferous. All rocks are metamorphic, with the grade
1546 increasing northward and upward, from the lower greenschist to the amphibolite facies. A
1547 klippe of the External Nappes has been reported lying above the Autochthon by Pavanetto et
1548 al. (2012).

1549 The Internal Nappes, to the north, are a relatively narrow band 15-30 km wide mostly
1550 characterized by amphibolite-facies rocks (Giacomini et al., 2006). At the contact with the
1551 overlying Internal Zone, a strongly deformed metasedimentary belt with an inverted
1552 Barrovian sequence includes retrogressed eclogites derived from MORB-type protoliths.
1553 Conditions of the eclogitic event are estimated at *c.* 1.5 GPa and 600-700 °C (Cortesogno et
1554 al., 2004), and the rocks did not record a high-T granulitic stage. This belt represents the
1555 thrust contact between the Nappe and Internal zones, and indicates a southward emplacement
1556 of the latter (Cappelli et al., 1992). The Posada-Asinara late Variscan dextral shear zone,
1557 dated at *c.* 325-300 Ma (Carosi et al., 2020), generated a green-schist-facies phyllonite.

1558 The Internal Zone in Corsica is similar to the Leptyno-Amphibolitic Complex of the
1559 FMC, and consists of early Paleozoic metasediments and orthogneisses with eclogite boudins
1560 but also high-P/high-T basic granulites and ultrabasics. Peak conditions of high-P
1561 metamorphism are estimated at 1.4-1.8 GPa and 900-1000 °C and dated at *c.* 360 Ma
1562 (Giacomini et al., 2008) or *c.* 420-400 Ma (Rossi et al., 2009). In northern Sardinia and
1563 Asinara Island, the Leptyno-Amphibolitic Complex includes orthogneisses dated at *c.* 460 Ma.
1564 Magmatic affinity is alkaline in Asinara and calc-alkaline in northern Sardinia. An eclogitic
1565 stage *c.* 400 Ma old was followed by a granulitic event and then by pervasive migmatization.

1566 What appears to be the uppermost allochthonous unit crops out in northern Corsica,
1567 where quartzites and greywackes with interbedded metabasalts represent the Neoproterozoic.
1568 A well-preserved Cambrian to Silurian detrital sequence includes Upper Ordovician
1569 glaciomarine diamictites and Silurian radiolarites. An Upper Devonian Culm deposit appears
1570 in faulted contact with the Paleozoic succession. Regional metamorphism was at most of
1571 lower greenschist facies, but many occurrences are roof pendants of the late Variscan Corsica-
1572 Sardinia batholith and were transformed into hornfelses.

Rossi et al. (2009) compare the low-grade Paleozoic succession of northern Corsica
1573 with that of SW Sardinia, finding significant differences. They conclude that northern Corsica
1574 was a different sedimentary area than SW Sardinia, the Maures or the Montagne Noire, and
1575 interpret it as deposited in a microcontinent separated from Gondwana during the Ordovician.
1576 Actually, this succession is quite similar in facies and thickness to that of the Mauges, in the
1577 SAD.

The External Massifs of the Alps (EmA) have not been included in the map of Figure
1579 1 because their position in the late Carboniferous is imprecise, even if generally considered
1580 part of the Allochthon because some of their units are polymetamorphic and include eclogites
1581 and high-P granulites. The EmA are shown on their present position in Figure 4. A good
1582 synopsis of their geology is offered by von Raumer et al. (1993). In the Belledonne and Mont
1583 Blanc massifs (Fig. 4), Ballèvre et al. (2018) separate three domains. The first is the western
1584 Belledonne domain, which consists of medium-grade schists derived from a monotonous
1585 terrigenous sequence of unknown age. The second domain occurs in southern Belledonne,
1586 where a Variscan nappe stack includes the Cambro-Ordovician Chamrousse Ophiolite, and
1587 ultramafic and gabbro complex with protoliths dated at *c.* 500 Ma (Pin and Carme, 1987;
1588 Ménot et al., 1988). Still forming part of the second domain as the relative autochthon of the
1589 ophiolite, a Devono-Carboniferous bimodal suite of leptynites and amphibolites occurs, in
1590 what may represent an active margin or be related with rifting (von Raumer et al., 1993).
1591 Moreover, a low-grade detrital and volcanic sequence of argilites, sandstones, conglomerates,
1592 spilites and keratophyre, probably Viséan, represent an intracontinental pull-apart basin
1593 (Ménot, 1988). The third domain is widely represented in the External Massifs and
1594 characterized by a polymetamorphic evolution. The age of eclogite or high-P granulite facies
1595 metamorphism is uncertain. Although largely taken as Ordovician, Silurian and Early
1596 Devonian ages have been reported (Paquette, 1987; Paquette et al., 1989), and more recent

1598 data point to a Carboniferous age (c. 330 Ma; Rubatto et al., 2010). It was overprinted by an
1599 amphibolite-facies event culminating in partial melting.

1600 The internal zones of the Alps share also a Variscan basement and a Gondwanan
1601 derivation, but their original position relative to the external zones and the Variscan Belt is
1602 difficult to establish because they are separated from the EmA by the Zone Houillère,
1603 interpreted by Ballèvre et al. (2018) as a transtensional basin developed along a late Variscan
1604 dextral shear zone. According to von Raumer and Neubauer (1993), the Alpine basement may
1605 include pieces detached from the north African margin after the Variscan orogeny, which may
1606 be monometamorphic or polymetamorphic. The first can be ascribed to the Autochthon, while
1607 the second, generally characterized by early high-P conditions, are possibly related with the
1608 Allochthon, but could also represent different terranes. Due to this uncertainty, no further
1609 discussion on the internal zones will be included. The same accounts for other basement
1610 outcrops in Italy described by Molli et al. (2020), who propose a Permian paleoposition SW
1611 to the Alps. For a discussion on linkages between the Alpine basement and other more or less
1612 Alpinized Mediterranean Variscides see Neubauer and von Raumer (1993).

4.5. *Variscan intrusive magmatism in France*

1615 Variscan plutonism is abundant in the two main French massifs, being more
1616 voluminous in the Massif Central, probably because this was the area where crustal
1617 thickening reached its maximum. Variscan magmatism is related to crustal thickening and
1618 resulting crust/mantle interaction. In the Armorican Massif, a link with crustal-scale wrench
1619 tectonics has been established. Devonian-early Carboniferous magmatism is also important,
1620 as it reflects subduction processes and can be related with plate convergence. Variscan
1621 magmatism is also voluminous in other French and related massifs: Vosges, Schwarzwald,
1622 Corsica, Sardinia and the External Massifs of the Alps.

4.5.1. *Granitoids of the Armorican Massif*

1625 In the Armorican Massif, Variscan granitoids occur mostly to the west in the CAD and
1626 NAD, and also along the whole SAD (Fig. 4). According to their regional distribution, three
1627 groups are distinguished: the red granite belt along the northern shore of Brittany, the Middle
1628 Armorican Batholith with large elliptic plutons, and the leucogranite belt, with elongated

1629 plutons emplaced during the activity of the South Armorican Shear Zone (Carron et al.,
1630 1994).

1631 The red granites of the northern belt are late Variscan granodiorites and syenogranites
1632 of the I-type, emplaced at *c.* 300 Ma. These subalkaline granitoids probably derive from
1633 fractional crystallization of basaltic magmas plus an important crustal contribution. The
1634 middle Armorican plutons are mostly grey, porphyritic monzogranites, often with concentric
1635 zoning and associated with both peraluminous and mafic bodies. Two series have been
1636 distinguished depending on the presence or absence of cordierite. They were emplaced
1637 between 340-290 Ma. Some are subalkaline, but even the cordierite-bearing plutons, which
1638 have an important crustal component, may be related to mafic magmas. The plutons of the
1639 leucogranite belt are progressively younger from the north and NE to the south and SW. They
1640 are mostly of the S-type, derived from partial melting of the continental crust, and were
1641 strongly deformed in the two branches of the South Armorican Shear Zone. Their
1642 emplacement took place around 340-330 Ma in the N-NE, 330-310 Ma in the central band,
1643 and 310-290 Ma in the S-SW (Carron et al., 1994; Edel et al., 2018).

1644 In addition to the Variscan granites, a few migmatitic orthogneisses in the Léon are
1645 Lower-Middle Devonian (390 Ma) porphyritic granitoids (Marcoux et al., 2009). By their
1646 position in the Variscan belt and age, they can be related with the Rheic suture of the model
1647 of Zeh and Will (2010) for the Mid German Crystalline High.

1648 A younger early Variscan intrusion of gabbro-diorite with calc-alkaline affinity occurs
1649 in the Upper Allochthon south of the Mauges Unit and north of the Chantonay Syncline (Fig.
1650 4). It is dated at *c.* 373 Ma (Cuney et al., 1993), and could be related with a poorly dated
1651 Devonian back-arc basin represented by mafic volcanics in the Chantonay syncline (Wyns et
1652 al., 1989), and also with extensive felsic volcanism with massive sulphides developed during
1653 the Famennian-Viséan in the eastern Massif Central (Delfour, 1989) and Vosges (Lefèvre et
1654 al., 1994; Krecher et al., 2007). It is also interpreted as the westernmost representative of a
1655 series of intrusives exposed in the western Massif Central, aged 380-360 Ma and collectively
1656 known as forming the tonalitic line of Limousin (Bernard-Griffiths et al., 1985a; Bertrand et
1657 al., 2001). It

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1659 4.5.2. *Granitoids of the French Massif Central*

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21660 Upper Paleozoic granitoids occupy *c.* 50 % of the outcropping surface in the FMC,
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41661 with ages ranging 380-280 Ma. They include a calc-alkaline association with tonalites,
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61662 granodiorites and monzogranites, and potassic subalkaline metaluminous and peraluminous
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81663 associations with monzogranites, muscovite and cordierite granitoids, and two-mica
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101664 leucogranites. The large massifs are igneous complexes composed of multiple phases
111665 spanning in time 40 Ma.

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141666 Early Variscan magmatism was widely developed in the FMC linked with plate
151667 convergence. It started with pervasive migmatization in the Upper Gneiss Unit (385-370 Ma),
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171668 closely followed by that of the Lower Gneiss Unit (375-365 Ma), related in both cases to
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191669 exhumation following subduction (Santallier et al., 1994; Faure et al., 2008), which induced
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211670 the formation of migmatitic domes (Ledru et al., 1994b). The Upper Devonian plutons
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231671 forming the tonalitic line of Limousin (380-360 Ma) occur only in the Upper Gneiss Unit,
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251672 equivalent to the Iberian and Armorican Upper Allochthon. They are related to oceanic
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271673 subduction previous to that of continental Lower Gneiss Unit. Contemporaneous rocks of the
281674 Morvan Arc and ophiolitic Brévenne Unit may be shallow manifestations of the same
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301675 phenomenon, the closure of the ocean separating the Upper Allochthon from the rifted margin
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321676 of northern Gondwana (Lower Allochthon). Conversely, the calc-alkaline volcanism of the
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341677 Tuffs Anthracifères witnesses younger (350-330 Ma) back-arc magmatism (Lardeaux et al.,
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361678 2014), and probably results from subduction of the Rhenohercynian Ocean.

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381679 Ages and magma types do not show a clear polarity, although they depict some
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401680 zonation (Ploquin and Stussi, 1994; Ledru et al., 1994b). Most granitoids older than 330 Ma
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421681 are distributed in the central part of the FMC, and are surrounded by younger intrusives. This
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441682 central region is characterized by diachronous alumino-potassic granitoids, and includes the
451683 Velay Dome (Fig. 4), a large anatectic complex composed of a number of intrusions of
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471684 varying composition, but essentially peraluminous although with spatially but not genetically
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491685 related vaugnerites (Ploquin and Stussi, 1994). A calc-alkaline province with granodiorites
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511686 and granites (325-315 Ma) occur in the Limousin cross-cutting the older plutons of the
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531687 tonalitic line (Fig. 4), and also in the NE, where alumino-potassic, subalkaline and calc-
541688 alkaline granitoids intruded between 345 and 280 Ma, partially coeval with the formation of
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561689 the Tuffs Anthracifères. In the southern region (Cévennes and Montagne Noire), plutons are
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581690 subalkaline to calc-alkaline, localized, and younger than 330 Ma.

1691 With the exception of the early Variscan calc-alkaline rocks of the Limousin and
1692 Morvan, and those of the Tuffs Anthracifères, no relation with a subductive environment may
1693 be established. Although a mantle contribution is not excluded, most Variscan granitoids
1694 derive from a late Proterozoic crustal source including peraluminous metasediments,
1695 orthogneisses and lower crustal granulites (Ploquin and Stussi, 1994; Chelle-Michou et al.,
1696 2017). The large amount and long interval of granite generation implies maintained heat
1697 production by radioactive decay, and points to significant crustal thickening. Amongst the
1698 large Variscan Massifs, the FMC is the one where it can be granted that the Variscan
1699 Allochthon occupied a largest proportion of its present extent (more than 90 %, nearly 100%
1700 if the eclogites of the Montagne Noire are related with overburden induced by the
1701 Allochthon).

1702 According to thermal models by Alcock et al. (2015), crustal thickening induced by
1703 the Allochthon and to a lesser extent by recumbent folding in the Autochthon in NW Iberia,
1704 may have provided a very significant contribution to the heat budget necessary for generalized
1705 crustal melting, induced gravitational collapse and crustal extension and thinning. It is
1706 noteworthy that the eastern part of the FMC is characterized by the thinnest crust among the
1707 large Variscan Massifs (Dezes and Ziegler, 2002; see Fig. 5c).

1708 1709 *4.5.3. Granitoids of other French and related massifs*

1710 Variscan granitoids are abundant in the Vosges and Schwarzwald. In the small
1711 occurrence of the Saxothuringian Zone in the northern Vosges, diorites, granodiorites,
1712 monzogranites and granites intruded between 335 and 320 Ma associated with ignimbritic
1713 volcanism, represent a magmatic arc related with the subduction of the Rhehercynian
1714 Ocean (Edel and Schulmann, 2009; Edel et al., 2013; Lardeaux et al., 2014). In the Central
1715 Vosges, ultra-potassic magmas including durbachites were emplaced at 345-332 Ma, while
1716 younger granite bodies intruded at 335-325 Ma (Tabaud et al., 2015). Intrusion of the ultra-
1717 potassic rocks was coeval with diapiric ascent of high-T rocks from lower crustal levels
1718 (Schulmann et al., 2002; Skrzypek et al., 2014). Ultra-potassic rocks dated at 345-335 Ma
1719 also occur in the Southern Vosges, intruding Devonian and Carboniferous sediments.

1720 In the Schwarzwald, layered basic rocks dated at 350 Ma are interpreted as intrusions
1721 related to early Carboniferous convergence. Later Variscan granitoids are abundant:
1722 muscovite or two-mica granites intruded at 330-325 Ma in the Central Gneiss Complex, while

1723 in the Southern Gneiss Complex, emplacement of biotite and two-mica granites is dated at
1724 335-330 Ma.

1725 In the Maures-Tanneron Massif, several magmatic events are described by Corsini et
1726 al. (2009). The older consists of granodioritic melts related to the early Variscan subduction.
1727 Although dated as Silurian-Lower Devonian (440-410 Ma; Demoux et al., 2008), they are
1728 probably products of the Middle-Upper Devonian (385-375 Ma) partial melting that followed
1729 *c.* 415-380 Ma early Variscan subduction of the Upper Gneiss Unit in the FMC (Lardeaux et
1730 al., 2014). Later, crustal thickening produced anatectic granitoids intruded at *c.* 340 Ma, and
1731 were followed by late Variscan leucocratic granites (320-315 Ma) related to isothermal
1732 decompression, and by late-kinematic granites linked to dextral shearing (*c.* 300 Ma; Corsini
1733 et al., 2009).

1734 In Corsica, Mg-K granitoids intruded at 345-335 Ma in relation with collision related
1735 with the early Variscan suture (Rossi et al., 2009). In northern Sardinia, peraluminous
1736 intrusions (330-308 Ma) predate the metaluminous intrusions of the Corsica-Sardinia
1737 batholith. The latter consists of late Carboniferous-early Permian granitoids (310-290 Ma)
1738 contemporaneous with the activity of a dextral shear zone developed between Gondwana and
1739 Laurussia. The more abundant intrusive rocks are granodiorites and monzogranites, but the
1740 whole spectrum includes also ultrabasic and basic rocks, tonalites and leucocratic granites,
1741 and all of them have their counterpart in coeval volcanic rocks (Rossi et al., 2009). These
1742 authors interpret the batholith as resulting from a Carboniferous asthenospheric rise producing
1743 mafic magmatic underplating and voluminous crustal melting.

1744 In the External Alps, several early Carboniferous granitoids occur in the Belledonne
1745 Massif (350-320 Ma), where Mg-K granites were emplaced during nappe stacking and
1746 sinistral transpression, which also occurred in the Mont Blanc Massif at 330 Ma (Corsini et
1747 al., 2009; Ballèvre et al., 2018). Regional crustal melting started at *c.* 320 Ma and most
1748 granitoids derived from crustal and mantle sources intruded at 310-300 Ma, in relation with
1749 major dextral shear zones. Upper Carboniferous granites occur mainly in the External
1750 Massifs, while Permian granitoids characterize the more internal massifs (Ballèvre et al.,
1751 2018). On the other side, Carboniferous sedimentation differs in age and nature at both sides
1752 of the boundary between the EmA and more internal massifs, being Namurian-Westphalian
1753 and shallow marine in the internal Zone Houillère and Stephanian and lacustrine in the EmA.
1754 These differences drove these authors to interpret this zone as a transtensional basin opened
1755 along a major dextral shear zone whose movement took place at 320-310 Ma.

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21757 **5. Bohemian and related massifs**

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This massif represents the cradle of subdivision in zones of the Variscan Belt and of correlation among its massifs by the pioneering work of Suess (father and son; 1888 1912, 1926) and Kossmat (1927). The following description is based on many sources, among which the contributions by Franke (2000), Franke and Żelaźniewicz (2000, 2002) for the whole massif and Mazur and Aleksandrowski (2001), Aleksandrowski and Mazur (2002), and Linnemann and Romer (2010) for significant parts of it.

The Bohemian Massif (Fig. 6) is flanked by the Rhenohercynian and Moravo-Silesian zones, which represent foreland thrust belts (Fig. 7). Its central part includes the Saxothuringian, Teplá-Barrandian and Moldanubian zones, whose interpretation varies depending of the authors. The Saxothuringian Zone is mostly considered the Autochthon although it was covered by an allochthonous nappe stack partially preserved in three klippen. The Teplá-Barrandian is now considered allochthonous and the root of the Saxothuringian klippen, while the character of the Moldanubian Zone is discussed.

5.1. *Rhenohercynian Zone (RHZ)*

Most of this zone actually crops out in the Rhenish Massif (Figs. 1 and 4), but is included here because it crops out in the Harz Mountains and underlies the Saxothuringian Zone. In the western part of the Rhenish Massif and to the north of it, in the Brabant Massif, an autochthonous Cadomian basement of Avalonian affinity, and a Cambrian to Silurian cover are unconformably overlain by Lower Devonian clastics of Laurentia or Baltica provenance (Franke, 2000; Sintubin et al., 2009). A Lower Devonian rifting is documented by abundant felsic volcanism which turned dominantly basic during the Middle Devonian to early Carboniferous.

At the limit of the RHZ with the STZ, a narrow belt named Northern Phyllite Zone (Fig. 6) includes a mixture of metasediments of different provenances and metavolcanics which may represent a late Silurian-Lower Devonian arc. According to Franke (2000), the zone may host the suture of the Rheic Ocean.

A new ocean, the Rhenohercynian, opened in the Emsian (~405 Ma) and closed in the early Frasnian (380 Ma; Franke, 2000) or early Carboniferous (~350 Ma; Zeh and Gerdes,

1787 2010). This was followed by early Carboniferous turbidites derived from an active margin to
1788 the SE, the Mid German Crystalline High (MGCH), which marks the closure of the
1789 Rhenohercynian basin.

1790 NW-directed thrusting developed at 335-325 Ma (Oncken et al., 2000). The Giessen
1791 Nappe of the SE of the Rhenish Massif (Fig. 4) which continues to the central part of the Harz
1792 Mountains (Fig. 1), carries at its bottom tectonics slices of a Lower Devonian oceanic crust
1793 and a mélange of sedimentary rocks of Saxothuringian provenance (Franke, 2000). The root
1794 of the nappe seems to occur between the Northern Phyllite Zone and the MGCH, and could
1795 represent the suture of the Rhenohercynian Ocean (Franke, 2000).

1796 5.2. Moravo-Silesian Zone (MSZ) and Brunovistulian block

1798 The Moravo-Silesian zone represents a *c.* 50 km wide and 300 km long tectonic
1799 boundary between the internal Moldanubian zone and the easterly Neoproterozoic
1800 Brunovistulian continental block (Dudek, 1980). This microcontinental block was formed as
1801 an Ediacarian intraoceanic (730 Ma) and later Japan-type continental arc (600 Ma) established
1802 on Paleoproterozoic crust (Collett et al., 2021). In Upper Silesia, the Brunovistulian basement
1803 is unconformably covered by a thick early to mid Cambrian siliciclastic sequence and
1804 sporadic upper Ordovician deposits (Bula et al., 2015). Detrital zircon and white mica age
1805 populations (Habryn et al., 2020), as well as early Paleozoic biogeographic data indicate
1806 derivation from NW Gondwana during the Cambrian, similarities with Avalonia and docking
1807 to Baltica between the Cambrian and Silurian (Belka et al., 2000). This zone formed part of
1808 southern Laurussia during the Devonian. The Devonian to Carboniferous sedimentary cover
1809 of the western margin of the Brunovistulian block is traditionally compared to the RHZ.

1810 Both RHZ and MSZ share rifting during the Middle Devonian, indicated by mafic
1811 volcanism, and shallow carbonate deposition during the Middle Devonian, followed by early
1812 Carboniferous, synorogenic detrital and carbonate sedimentation and late Carboniferous
1813 molasse. Lower Devonian (Pragian) clastic sedimentation (Chlupáč, 1989) was followed in
1814 Silesia by deposition of volcanosedimentary material associated with mafic to intermediate
1815 arc volcanism dated at 374 Ma (Janoušek et al., 2014). The Devonian rifting cycle terminated
1816 with development of a thick Givetian carbonate platform semi-continuously covering the
1817 whole western margin of the Brunovistulian block. The early Carboniferous Culm facies,
1818 Tournaisian to lower Namurian, reaches a thickness of 7 km forming a doubly-vergent

1819 accretionary wedge thrust to the SE and NW that is deeper toward the westerly internal MZ
1820 in Silesia and becomes more shallow toward the southeastern Moravian branch (*e.g.* Tomek et
1821 al., 2019).

1822 The structure of the MSZ is imbricated, involving the basement at its inner west part
1823 and of thin-skinned type toward the E (Fig. 7a). The western part of the MSZ emerges
1824 through a thick Moldanubian nappe in form of three tectonic semi-windows: the Svatka and
1825 Thaya domes and the Silesian domain (Suess, 1912, 1926). In the latter, thick-skinned
1826 tectonics involve three basement and cover thrust sheets folded in three large antiforms (Fig.
1827 7a): the Velké-Vrbno, Keprník and Desná domes (Schulmann and Gayer, 2000). To the
1828 central and southern Svatka and Thaya windows, they are formed by basement and Devonian
1829 cover duplexes at the bottom and two thick crystalline nappes at the top (Schulmann et al.,
1830 1991, 1994). Variscan Barrovian metamorphism is inverted, grading from the chlorite zone in
1831 the east to the kyanite-sillimanite zone in the west at the top of the sequence (Štípská and
1832 Schulman, 1995; Štípská et al., 2020). The highest part of the pile is formed by middle to late
1833 Cambrian ophiolitic sequences (*e.g.* Soejono et al., 2010; Collett et al., 2021) associated with
1834 eclogites (1.6 GPa, 650-700 °C), kyanite-staurolite micaschists and serpentized ultramafics
1835 (Konopásek et al., 2002; Štípská et al., 2006). The age of HP metamorphism is estimated as
1836 Upper Devonian based on geochronology of metamorphosed and deformed rocks of the Staré
1837 Město ophiolitic sequence in Silesia interpreted as an oceanic suture (Jastrzebski et al., 2013).
1838 Structurally above ophiolitic sequences in Silesia and Moravia occur migmatites, partially
1839 molten orthogneisses and high-P granulites of the Moldanubian nappe. The contact between
1840 the highest allochthonous Moravo-Silesian units and the Moldanubian is thus formed by
1841 remnants of a Cambrian ophiolite called in Silesia the Nýznerov Thrust (Skácel, 1989), which
1842 is an equivalent of Moldanubian thrust in the south, formed by micaschists of the kyanite-
1843 sillimanite zone containing a *mélange* of dismembered ophiolitic fragments and eclogites.

1844 Polyphase tectonic evolution of the MSZ is characterized by a first subduction of
1845 oceanic crust probably during Late Devonian (Jastrzebski et al., 2013) followed by early
1846 Carboniferous (345-340 Ma) westward continental underthrusting of the Brunovistulian
1847 margin beneath the Moldanubian hot fold nappe (Ráček et al., 2017; Štípská et al., 2015;
1848 2020). During this event started deposition of earliest sediments in the easterly flexural Culm
1849 basin (Ráček et al., 2017; Tomek et al., 2019). The margin of the underthrust
1850 Brunovistulian block was later imbricated producing two thick crustal nappes that were
1851 emplaced over the autochthonous basement together with the hot Moldanubian subduction

1852 channel flow region at around 340-335 Ma (Schulmann et al., 2008; Racek et al., 2017). This
1853 stage was associated with maximum deposition of sediments in the Culm foreland basin
1854 further east (Ktaková et al., 2007; Tomek et al., 2019). At this stage the western and oldest
1855 part of Culm sediments was underthrust and metamorphosed beneath the Moravo-Silesian
1856 nappe pile thereby forming a bivergent accretionary wedge (Tomek et al., 2019). The tectonic
1857 evolution of the MSZ is terminated by dextral shearing that reactivated the Moldanubian-
1858 Brunovistulian boundary at around 330-325 Ma (Schulmann et al., 1994; Racek et al., 2017).

1860 5.3. Saxothuringian Zone (STZ)

1861 There is abundant geological information about this zone, much of which is collected
1862 in the contributions gathered in Linnemann and Romer (2010). Brief accounts are provided by
1863 Franke (2000) and Franke and Żelaźniewicz (2002). Most of this zone represents the
1864 Autochthon of the Bohemian Massif, but several allochthons occur in klippen in the internal
1865 parts, the more important being the Münchberg klippe in the west, those surrounding the
1866 Krkonoše Massif, and the Góry Sowie Massif in the east. Furthermore, a poorly outcropping
1867 belt in the north known as the Mid German Crystalline High is often included as a domain of
1868 the STZ (*e.g.* Linnemann et al., 2007) but also considered as a separate entity (Franke and
1869 Żelaźniewicz, 2000, 2002). These domains are described separately in the following sections.

1870 The STZ has two branches that surround the more internal parts of the massif. The
1871 western branch is oriented SW-NE, and although mostly covered, can be continued through
1872 the Odenwald and northern Vosges to join the western Variscan massifs (Figs. 1, 4 and 6).
1873 The eastern branch has a NW-SE attitude derived from dextral shearing of the Elbe zone, of
1874 which the Elbe Fault Zone represents its southern limit (Fig. 6). This branch is known as the
1875 Lausitz Block.

1877 5.3.1. Mid German Crystalline High (MGCH)

1878 This domain crops out in a few inliers, the three more important of which are
1879 Odenwald, Spessart and Ruhla (Figs. 4 and 6), but information is also supplied by numerous
1880 drill holes. According to Franke (2000), it represents the SE active margin of the RHZ,
1881 characterized by arc-related plutonism. The MGCH may have a Cadomian basement, and
1882 includes i) a Cambro-Ordovician succession of paragneisses and metabasites, ii) Upper
1883 Ordovician-early Silurian granitoids, iii) Silurian and Lower Devonian arc-related magmatism

1884 similar to that of the Northern Phyllite Zone, iv) Upper Devonian calc-alkaline plutons
1885 probably from a Devonian-early Carboniferous magmatic arc, v) Middle Devonian to
1886 Tournaisian unmetamorphized sediments, and vi) arc-related middle Carboniferous granitoids
1887 (340-325 Ma; Franke, 2000; Zeh and Will, 2010). Zircon provenance studies show
1888 sedimentary derivation from Gondwana and Baltica in different groups of the Ruhla inlier and
1889 in other units, implying that the MGCH hosts the Rheic suture, according to Zeh and Will
1890 (2010) and Zeh and Gerdes (2010).

1891 Metamorphism is dated as early as 370 Ma, and indications of high-P rocks are
1892 generally absent, but eclogite lenses occur in the NE of Odenwald. A fault zone separating
1893 western from eastern Odenwald and Spessart is considered part of the Rheic suture by Will et
1894 al. (2017). Granulite facies metamorphism in western Odenwald is dated between 430 and
1895 349 Ma, whereas eclogite facies in eastern Odenwald has a minimum age of 363-350 Ma
1896 (Scherer et al., 2002), and amphibolite facies there and in central Spessart yielded *c.* 324-316
1897 Ma (Will et al., 2017). According to Will and Schmädicke (2001, 2003), minimum pressure
1898 of 1.6 GPa at 650-750 °C characterize the eclogites, while granulites in the west reached a
1899 maximum P of 0.4 GPa at minimum T of 680 °C.

1900 The eclogites crop out in the core of an antiformal structure interpreted as a tectonic
1901 window which continues with a SSW-NNE attitude to the Spessart (Will et al., 2017). An
1902 antiformal structure exists also at Ruhla with a similar attitude, which is oblique to present
1903 SW-NE zone limits, interpreted as late Variscan strike-slip faults truncating earlier structures
1904 (Franke, 2000).

1905 According to Zeh and Gerdes (2010), the Silurian-Devonian arc (425-392 Ma) resulted
1906 from NW-directed subduction of the Rheic lithosphere beneath Laurussia continental shelf.
1907 That subduction created a back arc leading to the opening of the Rhenohercynian basin and
1908 the Lizard-Giessen Ocean to the NW at the Lower-Middle Devonian. The Variscan evolution
1909 (360-320 Ma) started by the oblique (sinistral) collision between the peri-Gondwanan STZ
1910 and the Rhenohercynian domain. Different parts of the MGCH underwent different structural
1911 and metamorphic evolutions. The still preserved parts of the Rheic Ocean subducted to the
1912 south, starting to form a new arc in the STZ. Then the Silurian-Devonian arc became accreted
1913 to the base of a new arc and the Rhenohercynian Ocean was in turn closed, culminating the
1914 growth of the new, Rhenohercynian arc. Accretion started at 360-350 Ma, contemporaneous
1915 with flysch sedimentation in the Rhenohercynian basin, and ceased at *c.* 310 Ma. Formation
1916 of the Rhenohercynian arc started at 360 Ma and reached a peak at 340-330 Ma.

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21918 *5.3.2. Saxothuringian Autochthon*

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The rocks forming this domain are mostly low-grade metasediments of Neoproterozoic and Paleozoic age deposited at the northern or northwestern margin of Gondwana. The Neoproterozoic and lowermost Cambrian include volcanosedimentary complexes and massive intrusions developed between 570-535 Ma. They reflect the Cadomian orogenic event and the transition between plate convergence responsible for the arc and subsequent rifting, which opened the Rheic Ocean and pulled apart the peri-Gondwanan Avalonia microcontinent. Rifting was renewed at the Cambro-Ordovician boundary, with intrusions dated at 500-480 Ma (Linnemann et al., 2007, 2008). Bounding the MGCH to the south, the Vesser Rift or Southern Phyllite Zone is a Late Cambrian bimodal magmatic belt affected by greenschist facies Variscan metamorphism.

In the Cadomian basement, Linnemann et al. (2008) differentiate a northern external domain with distal turbidites including black cherts and volcano-sedimentary complexes dated at 570-543 Ma, which would represent the back-arc. The southern internal domain includes Ediacaran sediments of passive margin and flysch types, which represent respectively the margin of the back-arc and a Cadomian retro-arc basin. A *c.* 565 Ma glaciation has been identified by Ediacaran glaciomarine deposits of the Elbe Zone in the south of the Lausitz Block (Linnemann et al., 2018). Plutons of granodiorite composition are abundant and voluminous here, dated at 540-530 Ma.

For the Paleozoic, preorogenic sequences are mostly terrigenous, with carbonates of early-middle Cambrian and Devonian age, Lower Ordovician quartzites and pelites, and Silurian graptolite slates. All these are typical of relatively inner parts of the Gondwanan continental shelf, and comparable to successions in the Armorican and Iberian Massifs. Cambrian faunas in eastern Germany, including trilobites and archaeocyatha are comparable to those of France, Iberia and Morocco (Elicki, 2007). Upper Ordovician glacigenics have been found in the SW of the STZ (Katzung, 1961; Linnemann et al., 2003; Hofmann and Linnemann, 2013). Detrital zircon age spectra also indicate correlation with the OMZ of SW Iberia (Linnemann et al. 2008). The synorogenic sedimentation includes Tournaisian to Viséan Carboniferous flysch, progressively younger toward the NW, and late Viséan molasses (Franke and Engel, 1986). Both pre- and synorogenic autochthonous sequences are

1948 described together as the Thuringian Facies, mostly of shallow platform facies for the
1949 preorogenic sequence.

1950 Variscan deformation in the western branch of the STZ is dominated by upright SW-
1951 NE wrench shear zones and folds with an axial planar, low-grade pervasive cleavage. These
1952 are late folds, as one or more previous cleavages are identified at the NW and SW wrench and
1953 thrust zones and in the core of the antiform between them (Kroner et al., 2010). The folds,
1954 cleavage and metamorphism affect early Carboniferous synorogenic deposits. The authors
1955 describe the previous deformation events as consisting of crustal stacking and continental
1956 subduction followed by exhumation. Their SW-NE wrench and thrust zones refer to the bands
1957 coinciding with the MGCH in the NW and the allochthonous domain in the SE. Although
1958 they do not mention the kinematics of wrenching, Franke and Żelaźniewicz (2002) consider it
1959 dextral.

1960 Later dextral shear zones and faults with a NW-SE attitude cut across the whole
1961 Bohemian Massif, with the Elbe zone, in the north, as the most important (Fig. 6). This zone
1962 affects the Lausitz block, where the age of dextral shearing has been estimated between 334
1963 Ma and 327 Ma by U-Pb ages of a synkinematic and an undeformed granite respectively, both
1964 occurring close to the Elbe Fault Zone (Hofmann et al., 2009). Ar-Ar dating of the Intra-
1965 Sudetic Fault (Fig. 6) yielded a younger age interval of 325-320 Ma (Marheine et al., 2002).
1966 Franke and Żelaźniewicz (2002) interpret that the SW-NE early dextral system (380-340 Ma)
1967 was cut across by the NW-SE dextral system, which was in turn overprinted by the NNE-
1968 SSW Moldanubian Thrust, even when the ages of the two later are similar.

1969 In contrast with the general low-grade metamorphic imprint of most of the
1970 autochthonous STZ, several gneiss domes occur surrounding the inner part of the Bohemian
1971 Massifs. The Granulitebirge and Erzgebirge domes were formed beneath the outer rim of the
1972 Allochthon, and the same was probably the case of the Krkonoše and Orlica-Śnieżnik domes
1973 (Figs. 6 and 7a). Together, they form a curved alignment delineating an open Bohemian Arc
1974 that was yet drawn by Suess (1885-1909) and Kossmat (1927).

1975 The Saxonian Granulitebirge and the Erzgebirge (Fig. 6) include high-grade and high-
1976 and ultrahigh-P rocks (Massone, 2001; Massone et al., 2007; Collett et al., 2017). Their domal
1977 structure indicates a provenance from intracontinental subduction of the Saxothuringian crust
1978 and emplacement as gneiss domes or extrusions along the subduction zone, aided perhaps by
1979 hydraulic overpressure (Franke and Stein, 2000).

1980 The Saxonian Granulitebirge is described by Rötzler and Romer (2010). Its rocks crop
1 out in a SW-NE elongated dome with metamorphism progressively increasing toward its core,
2 1981 out in a SW-NE elongated dome with metamorphism progressively increasing toward its core,
3 1982 which consists of granulite-facies Cambro-Ordovician mainly bimodal igneous protoliths of
4 calc-alkaline and tholeiitic character. These are surrounded by a detachment zone with slivers
5 1983 of dismembered high-grade metapelites and mafic and ultramafic rocks, and by a low-P schist
6 cover of early Paleozoic to Upper Devonian age. The granulites recorded near isothermal
7 1984 decompression from metamorphic peak at 2-2.5 GPa, and post-peak recrystallization at 1.1
8 GPa. Peak T is dated at 341 Ma, whereas retrogradation in the base of the schist cover is dated
9 1985 at 334 Ma. Average rates of exhumation and cooling of the granulites have been calculated in
10 7-13 mm/year and 40-70 °C/Ma. Kinematic criteria indicate top to the SE sense of shear in the
11 1986 detachment zone (Fig. 7a; Kroner, 1995; Franke and Stein, 2000).
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20 1991 The Erzgebirge forms the structurally deepest part of the Saxothuringian basement. It
21 is formed by a complex stack of Cadomian basement and early Paleozoic sedimentary and
22 1992 bimodal volcanics (500-460 Ma) reflecting a passive margin environment, equivalent to that
23 of the Thuringian Facies. The lowermost Gneiss-Amphibolite Unit includes Neoproterozoic
24 1993 greywackes intruded by Cadomian granitoids (570-530 Ma), and affected by medium-
25 P/medium-T amphibolite facies metamorphism. Above, a pile of Paleozoic rocks starts with
26 1994 an ultrahigh-P/high-T Gneiss-Eclogite Unit, and continues upward by a high-P/high-T
27 Gneiss-Eclogite Unit, high-P/low-T Mica schist-Eclogite Unit, medium-P/low-T Garnet
28 1995 Phyllite Unit and low-P/low-T Phyllite Unit (Rötzler and Plessen, 2010). Mylonitic shear
29 zones separate the units and the ensemble from the overlying schist mantle. The Gneiss-
30 1996 Eclogite units also include granulites and ultramafic rocks, and micro diamonds occur in the
31 underlying ultrahigh-P unit, pointing to pressures equivalent to at least 150 km . Závada et al.
32 (2018) shown that the ultrahigh-P granulites and high-P granofelses preserved at the SE flank
33 1998 of the Erzgebirge testify a paleogradient of continental subduction originally dipping to the
34 east, and where the deepest rocks were exhumed along a subduction channel to mid-crustal
35 1999 depths.
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50 2007 The eclogites had been exhumed yet before stacking, and later, the entire stack became
51 overprinted by amphibolite-facies metamorphism, and was exhumed in the dome. The thrust
52 2008 faults responsible from the metamorphic inversion show a tectonic transport to the west or to
53 the N-NNE (Schmädicke, 1994; Franke and Stein, 2000). The upper detachment shows a
54 2009 westward component of motion for the hanging wall (Krohe, 1996, Krawczyk et al., 2000).
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2012 According to Franke and Stein (2000) and references therein, ages of high-P and
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2013 ultrahigh-P metamorphism are somewhat scattered, starting at *c.* 370 Ma and cooling through
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3014 *c.* 350 Ma until a significant event around 340 Ma. However, specific ages for eclogite-facies
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5015 metamorphism range 360-335 Ma (Schmädicke et al., 1995). The older ages come from Ar-
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7016 Ar and Sm-Nd analyses, and are discussed by Rötzler and Plessen (2010), who prefer the age
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9017 group clustering around 340 Ma as indicating peak of high-P metamorphism during
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11018 intracontinental subduction. Modern petrochronology studies (Závada et al., 2021) show that
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13019 the ultrahigh-P conditions for coesite-diamond bearing rocks were reached at 350 Ma, while
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15020 the granulites formed at around 340 Ma. The whole system was exhumed at around 338 Ma.
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17021 A later group of ages around 330 Ma date retrograde penetrative deformation during
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19022 exhumation.

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21023 There are two other domes which also bear the imprint of intracontinental subduction
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23024 and are overlain by units belonging to the Allochthon. One is the Krkonoše Massif, in the
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25025 Lausitz Block, and the other the Orlica-Šnieżnik Dome in Central Sudetes.

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27026 The Krkonoše Massif (Figs. 6 and 7a) is a dome whose core is occupied by a late
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29027 Carboniferous granitic pluton dated at 313 Ma (Marheine et al., 2002; Žák et al., 2013; Kryza
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31028 et al., 2014). From bottom to top, four units have been described (Mazur and Aleksandrowski,
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33029 2001; Žáčková et al., 2010; Jeřábek et al., 2016). The upper three units are interpreted as
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35030 forming part of the Allochthon, and will be described in the next section. The lowermost of
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37031 them is the Jizera-Kowary Unit, which may represent the STZ Autochthon. It consists of
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39032 Cambro-Ordovician orthogneisses (*c.* 500 Ma) intruded in metasediments of probable
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41033 Neoproterozoic age. To the west, the orthogneisses have intrusive contacts with Cadomian
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43034 granitoids of the Lausitz Block, which supports that the Jizera-Kowary Unit forms part of the
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45035 Autochthon.

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47036 This unit outcrops mostly to the north of the Variscan Krkonoše granite, where
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49037 metamorphism ranges from greenschist to lower amphibolite facies. To the south and east of
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51038 the Krkonoše granite, the unit includes albite-bearing micaschists, graphitic schists, quartzite,
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53039 marble and calc-silicate rocks, amphibolite and granitic orthogneiss. An estimate of early
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55040 metamorphic conditions in the south yielded 1.2-1.4 GPa at 460-520 °C, implying high-P
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57041 conditions (Žáčková et al., 2010). On the other hand, the relatively young age of this
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59042 metamorphism, *c.* 340 Ma (Konopásek et al., 2019) suggests a correlation with the
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61043 Erzgebirge, instead of with the overlying units, where the high-P is Upper Devonian.

2044 The Orlica-Śnieżnik Dome represents the more oriental outcrop of the STZ
1 2045 Autochthon and, as the Krkonoše Massif, it is overlain by the Allochthon, here represented by
2 2046 the Moldanubian Zone (Figs. 6 and 7a). The Autochthon is dominated by amphibolite-grade
3 2047 orthogneisses, locally containing lenses of high and ultrahigh-P eclogites and granulites,
4 2048 alternating with bands of biotite- to staurolite-grade supracrustal rocks (Don et al., 1990;
5 2049 Štípská et al., 2004; 2012). The orthogneisses have been dated at 520-490 Ma (Oliver et al.,
6 2050 1993; Turniak et al., 2000; Kröner et al., 2001; Lange et al., 2005). The supracrustal
7 2051 succession consists of a monotonous series of plagioclase-rich paragneisses and a variegated
8 2052 formation consisting mostly of mica schists. A Saxothuringian affinity is revealed by zircon
9 2053 provenance studies (Mazur et al., 2012, 2015). Occurrences of ultrahigh-P rocks provide a
10 2054 record of continental subduction and subsequent melt-assisted extrusion of deeply subducted
11 2055 rocks partly driven by buoyancy forces (Chopin et al., 2012) similar to that of the Erzgebirge.
12 2056 The structure is interpreted as a large gneiss dome created by the expulsion of lower crustal
13 2057 material at the rear of the Sudetic accretionary prism (Chopin et al., 2012; Štípská et al.,
14 2058 2019). The dome comprises dismembered fragments of the Saxothuringian passive margin
15 2059 that were subducted during the Variscan collision and then exhumed in front of the rigid
16 2060 buttress of Brunovistulian block, according to Duretz et al. (2011) and Chopin et al. (2012).
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31 2061 32 33 2062 *5.3.3. Saxothuringian allochthons*

34 2063 Figure 7b shows a section across the SW part of the Bohemian Massif perpendicular to
35 2064 the western branch of the STZ. It shows the relationships between the Autochthon and the
36 2065 allochthons of this zone, represented here by the Münchberg Klippe, and also with the also
37 2066 allochthonous Teplá-Barrandian and Moldanubian. The Allochthon and Parautochthon in the
38 2067 western branch of the STZ is represented by the Münchberg, Wildenfels and Frankenberg
39 2068 klippen, whose units are separated by thrust faults (Fig. 6). Münchberg is the largest and
40 2069 shows the most complete succession (Klemd, 2010), which is described from bottom to top.
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49 2070 The lower part of the Parautochthon consists of early Carboniferous, mainly Viséan
50 2071 proximal synorogenic flysch, often a wildflysch with blocks of probably Cambrian and also
51 2072 Ordovician, Silurian and Devonian blocks and olistoliths (Franke, 1989; Linneman et al.,
52 2073 2003; Klemd, 2010). The upper Parautochthon is represented by the distal and relatively deep
53 2074 Bavarian facies, including Ordovician bimodal volcanism and clastic sediments, as well as
54 2075 Silurian-Devonian volcanics, graphitic schists and cherts.
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2076 Above, a unit formed by phyllites and prasinites occurs, followed by another
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2077 consisting of amphibolite represent the Middle Allochthon. They were formed around 400 Ma
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2078 (Koglin et al., 2018), and metamorphosed at *c.* 365 Ma (Kreuzer et al., 1989).
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62079 The two remaining units represent the Upper Allochthon. The first consists of
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2080 metasediments, orthogneisses, metagabbros and serpentinites, while the culminating unit is
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92081 made of banded hornblende gneisses, amphibolites and leucocratic gneisses, and includes
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112082 eclogites at the base (Klemd, 2010). The igneous protoliths have ages of 525-480 Ma, and the
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132083 eclogites point to equivalence with the high-P/high-T units, with peak pressure conditions
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152084 roughly dated in Münchberg at 400 Ma and bearing a record of subsequent amphibolite-facies
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172085 metamorphism at 380 Ma (Franke, 2000; Koglin et al., 2018).
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192086 The structural succession of the Frankenberg klippe consists of a low-grade
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212087 volcanosedimentary complex that can be ascribed to the STZ autochthon. Above, a prasinite
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232088 unit would be equivalent to the phyllite and prasinite unit of Münchberg, and represent an
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252089 upper crustal piece of a Devonian oceanic crust. The klippe culminates with high-P
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272090 amphibolite and gneiss units (Rötzler et al., 1999; Klemd, 2010) probably representing the
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292091 Upper Allochthon. In Wildenfels, only the Upper Allochthon is represented by augengneisses,
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302092 gneissic micaschists, amphibolitic gneisses and amphibolite units (Klemd, 2010).
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332093 Kinematic criteria indicate final emplacement of the Münchberg klippe toward the
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352094 NW, but previous kinematic criteria indicate top to the SW or WSW in augengneisses and
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372095 mylonites of the Upper Allochthon (Blümel et al., 1990). Interestingly, a similar kinematics
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392096 has been described in the Erbendorf-Vohenstrauß Unit, at the base of the Moldanubian Zone
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412097 (Franke, 1990; Blümel et al., 1990).
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442098 In the eastern branch of the STZ, the Allochthon occurs in the west and central
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462099 Sudetes, surrounding the Krkonoše Massif, and in the Góry Sowie Massif respectively. Figure
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482100 7a shows a section partly oblique to the structure, but subparallel to it in the eastern branch of
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502101 the STZ. It cuts across the Erzgebirge, Krkonoše and Orlica-Śnieżnik domes, which are
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522102 equivalent (not successive) structures which delineate the Bohemian Arc. This part of the
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542103 profile is intended to show the suggested mutual relationships among the different
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562104 allochthons.
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592105 South and SE of the Krkonoše Massif, three units have been described. The lowermost
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612106 of them crops out in just a small area SW of the pluton. It consists of very low grade Devonian-
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632107 Carboniferous sediments including Culm facies on top, and is considered the Parautochthon
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2108 by Mazur and Aleksandrowski (2001). Above, the South Krkonoše Unit crops out along the
1 2109 southern flank of the dome. It is formed by garnet-free micaschists, phyllites and marbles with
2 2110 a high proportion of metavolcanics, of which the metabasites show relics of blueschist facies
3 2111 (1.2-1.5 GPa; Majka et al., 2016; Konopásek et al., 2019), dated around 365 Ma (Maluski and
4 2112 Patočka, 1997; Konopásek et al., 2019).

8 2113 To the east of the Krkonoše Massif, another unit overlies the South Krkonoše Unit. It
9 2114 consists of medium-P, MORB-type Cambro-Ordovician metabasites (Oliver et al., 1993;
10 2115 Kryza et al. 1995; Winchester et al. 1995). It represents the older oceanic units of the Middle
11 2116 Allochthon, and can be continued northward to the Kaczawa Complex (Figs. 6 and 7a).

16 2117 The Kaczawa Complex represents the Middle Allochthon, including its two oceanic
17 2118 components (Martínez Catalán et al., 2020). Below, Cambro-Ordovician metasedimentary
18 2119 and bimodal metavolcanics, with dominant alkaline basalts, represent the development of a
19 2120 continental rift. Metamorphism in all these units reached blueschist facies, and was
20 2121 overprinted by greenschist facies (Kryza et al., 1990, 2004, 2011). Above, a couple of Silurian
21 2122 and Devonian units with thick, pillowed MORB-type basalts represent the evolved stages of a
22 2123 rift and a deep-water environment, perhaps oceanic supracrustals (Furnes et al., 1994; Seston
23 2124 et al., 2000). Another characteristic component is an Upper Devonian-early Carboniferous
24 2125 flysch with chaotic assemblages and tectonic mélanges, always of very low metamorphic
25 2126 grade (Kryza and Zalasiewicz, 2008). Some of these synorogenic deposits separate the lower
26 2127 and upper units of the Kaczawa Complex. Moreover, a rather undeformed basaltic unit of
27 2128 probable Upper Devonian age may represent an even younger ophiolite (Seston et al., 2000).

32 2129 The most oriental Allochthon of the STZ outcrops in and around the Góry Sowie
33 2130 Massif (Figs. 6 and 7a). Although it is bounded by late Variscan faults, its lithological
34 2131 association and tectonometamorphic evolution, together with the ophiolites surrounding a
35 2132 catazonal unit favor an allochthonistic interpretation. These ophiolites are Lower Devonian
36 2133 and collectively considered a dismembered ophiolite suite (*e.g.* Kryza et al., 2004). For
37 2134 instance, the Ślęza ophiolite north of the massif consists of variably serpentinized
38 2135 harzburgites and lherzolites, coarse grained gabbros and a roof assemblage of fine-grained
39 2136 metagabbros, diabase and volcanics, including pillow lavas, and is dated at 400 Ma (Oliver et
40 2137 al., 1993; Floyd et al. 2002; Dubińska et al., 2004).

45 2138 The Góry Sowie Massif consists of migmatitic paragneisses previously assumed to be
46 2139 late Neoproterozoic to Cambrian based on microfossils (Gunia, 1999) and have been recently

2140 dated at 600-493 Ma using detrital zircons (Tabaud et al., 2021). There are also orthogneisses,
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2 2141 dated at 490-480 Ma (Kröner and Hegner, 1998; Kryza and Fanning, 2007), and some
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4 2142 coronitic metagabbros, amphibolites and ultramafics. A high-P/high-T granulite facies event
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6 2143 is dated at 400-395 Ma, and an amphibolite-facies overprint occurred at 385-365 Ma (O'Brien
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8 2144 et al., 1997; Bröcker et al., 1998; Kryza and Fanning, 2007; Tabaud et al., 2021). This
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10 2145 tectono-thermal event was coeval with emplacement of Devonian ophiolitic rocks presently
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12 2146 surrounding the Góry Sowie Massif.

15 2148 *5.4. Teplá-Barrandian and Moldanubian*

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18 2149 The classical Moldanubian Zone (Kossmat, 1927) is considered a peri-Gondwanan
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20 2150 coherent terrane bounded by sutures that separate it from the Saxothuringian Zone in the north
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22 2151 and NW, and from the Moravo-Silesian Zone in the SE (Matte et al., 1990; Franke, 2000).
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24 2152 Which are the oceans witnessed by these sutures is a matter of debate, as it is the correlation
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26 2153 of the Moldanubian with other Variscan terranes or Zones. High-P basic and ultrabasic rocks
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28 2154 occur in parts of this zone, and have been interpreted as klippen of a large thrust sheet, the
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30 2155 Gföhl Nappe, emplaced to the N, E and ESE and having covered the whole zone (Tollmann,
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32 2156 1982; Matte et al., 1990; Franke, 2000). This interpretation has been challenged by
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34 2157 Schulmann et al. (2005, 2009, 2014b), who see the high-P units as exhumed pieces of the
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36 2158 lower crust underlying the zone and being a consequence of continental relamination (Hacker
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38 2159 et al., 2011).

39 2160 Two eclogitic units occur at the base of the Teplá-Barrandian and Moldanubian zones,
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41 2161 Mariánské Lázně and Erbendorf-Vohenstrauß respectively (Figs. 6 and 7b). The Teplá-
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43 2162 Barrandian Zone is considered either a separate terrane (Matte et al., 1990; Franke, 2014) or
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45 2163 part of the Moldanubian terrane, occupying a relatively higher structural position, and
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47 2164 separated from the Moldanubian Zone by large-scale normal and wrench faults (Pitra et al.,
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49 2165 1999; Franke, 2000; Zulauf et al., 2002). It is possible that these faults hide the lateral
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51 2166 transition between two parts of the same terrane with different characteristics (Schulmann et
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53 2167 al. 2009; Lardeaux et al., 2014). In this case, the two zones would represent the upper
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55 2168 ensemble of the Bohemian nappe pile, perhaps being equivalent to the Upper Allochthon of
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57 2169 the French and Iberian massifs. Then, the two eclogitic units occurring at the base of the
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59 2170 Teplá-Barrandian and Moldanubian zones (Mariánské Lázně and Erbendorf-Vohenstrauß
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61 2171 respectively; Figs. 6 and 7b) would represent the suture of such a terrane. In the next sections

2172 the Moldanubian Zone is described first followed by the Teplá-Barrandian Zone, according to
2173 the bottom to top mode followed in previous sections.

2174 2175 5.4.1. Moldanubian Zone (MZ)

2176 This zone was divided by Fuchs (1976, 1986) into two major ensembles, Drosendorf
2177 and Gföhl. The Drosendorf Unit includes a lower ensemble called the Monotonous Group,
2178 consisting of migmatitic paragneisses, quartzites, amphibolites and numerous Paleo- to
2179 Neoproterozoic gneiss fragments ranging from 1.3 to 0.56 Ga (Lindner and Finger, 2018;
2180 Lindner et al., 2019) and gneissified early Paleozoic intrusions (*e.g.* Friedl et al., 2004). The
2181 detrital zircon data suggest deposition of the Monotonous group sediments during Early
2182 Paleozoic, most likely early Ordovician times (Košler et al., 2013). The higher ensemble
2183 called the Varied Group comprises paragneisses, quartzites, marbles, amphibolites and
2184 leptynites of Neoproterozoic and early Paleozoic age, interpreted as deposited in a back-arc
2185 setting (Janoušek et al., 2008; Schulmann et al., 2009). The detrital zircon record points to
2186 Early Paleozoic, possibly Silurian deposition (Košler et al., 2013). The Monotonous and
2187 Varied groups underwent medium-P amphibolite facies (0.8-1 GPa, 600-700 °C) at *c.* 350-330
2188 Ma, although eclogite boudins occur mainly at the contact between them (Schulmann et al.,
2189 2005).

2190 The Gföhl Unit is viewed here as a collective name for several massifs, most of them
2191 distributed in the north and SE of the MZ, but also occurring in the central part of the
2192 Moldanubian (Kutna Hora and Blanský les units; Fig. 6), between the two main plutonic
2193 belts. Rocks of the Gföhl Unit include felsic migmatites, felsic granulites, Ordovician
2194 orthogneisses, retrograded eclogites of mantle and crustal origin, and mantle peridotites.
2195 Mantle rocks attained P-T conditions of 6-3.3 GPa and 875-1150 °C, while crustal rocks
2196 reached peak conditions above 3-4 GPa and 900-1000 °C at *c.* 350 Ma (Perraki and Faryad.,
2197 2014; Nahodilová et al., 2020). This event was followed by major reequilibration of high-P
2198 granulites at 1.8-2 GPa at 340 Ma (Štípská et al., 2016). Coeval potassic and ultra-potassic
2199 intrusions (durbachites) occur spatially related to the granulites and include relics of high- and
2200 ultrahigh-P basic and ultrabasic rocks (Janoušek and Holub, 2007; Janoušek et al.,
2201 2020).

2202 The meaning of the Gföhl Unit and the role of the STZ subducted crust has been
2203 discussed. Matte et al. (1990) and Franke and Żelaźniewicz (2000, 2002) interpreted the unit

2204 as a thrust sheet emplaced above the MZ, while Schulmann et al. (2005, 2009, 2014b)
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2205 describe the evolution of the MZ in terms of subduction of the STZ beneath a continental
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32206 terrane. The association of granulites and durbachites in the MZ has been interpreted as a
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52207 Visean product of relamination, as defined by Hacker et al. (2011) in subduction zones
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72208 bearing sediments or involving continental or arc crust. It implies emplacement of felsic
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92209 gneisses derived from lower plate and incorporated to the upper plate, mostly at the lower
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112210 crust. In the MZ, partially molten gneisses derived from the subducted continental lower crust
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132211 of the STZ may have penetrated horizontally from the subduction channel to the SE
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142212 relaminating the Moldanubian lower crust, part of which extruded later vertically reaching
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162213 upper crustal levels (Lexa et al., 2011; Kusbach et al., 2015).
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192214 Chopin et al. (2012) have proposed flat subduction of the STZ continental crust
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202215 beneath the Teplá-Barrandian rocks and extrusion of ultrahigh-P to high-P rocks in the Orlica-
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222216 Šniežnik Dome. If true, the felsic granulites with high-P relics could derive directly from this
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242217 crust, and their occurrence can be explained by extrusion from the STZ crust and diapiric
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262218 intrusion close to the Moldanubian Thrust in the SE, and also in the central part of the MZ
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282219 (Fig. 7b; Lexa et al., 2011; Schulmann et al., 2014b; Kusbach et al., 2015).
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302220 31 322221 *5.4.2. Teplá-Barrandian Zone (TBZ)* 31 32 33

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352222 Underlying the Mariánské Lázně and Erbendorf-Vohenstrauß units, an association of
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372223 low-grade metasediments, basic rocks, acid tuffs and serpentinites reaching at most epidote-
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392224 amphibolite facies occurs (Kladská unit of Kachlík, 1997). It overlies the Thuringian-facies
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412225 autochthonous Paleozoic and is similar to the unit with phyllites and prasinites of Münchberg.
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432226 The Mariánské Lázně Unit consists of amphibolites, often migmatitic, with minor
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452227 alternations of felsic gneisses, and rare metasediments and pyroxenites. An eclogite-
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472228 amphibolite zone occurs between a basal layer of serpentinite and an overlying gabbro-
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492229 amphibolite zone (Collett et al., 2018). The protolith age of eclogites and amphibolites is *c.*
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512230 540 Ma old, while gabbros in the upper zone, often coronitic, have yielded 500 Ma
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522231 (Timmermann et al., 2004, 2006). Lu-Hf and Sm-Nd garnet-whole rock geochronology yield
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542232 *c.* 390 Ma for high-P metamorphism (*c.* 2.5 GPa, 650-750 °C) and 375 Ma for cooling after a
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562233 high-T overprint, in accordance with the previous age data (Collett et al., 2018 and references
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582234 therein). Calc-alkaline gabbros intruded the Mariánské Lázně eclogites during their
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602235 syncompressional and later extensional exhumation at *c.* 380 Ma (Deiller et al., submitted).
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2236 Overlying the Mariánské Lázně Unit, the Teplá Crystalline Complex is a band 10 km
1 2237 wide of Neoproterozoic metasediments and metabasalts where a Cambro-Ordovician low-
2 2238 P/high-T metamorphic event (*c.* 485 Ma) was overprinted by a Late Devonian, Barrovian
3 2239 event related with crustal thickening (*c.* 375 Ma), and then by low-P/high-T metamorphism
4 2240 related to a giant detachment along which the Maránské Lázně complex was exhumed
5 2241 (Peřestý et al., 2017).
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11 2242 The thick Neoproterozoic succession of the TBZ starts with Cadomian arc-related
12 2243 magmatism at the east (Davle group) which characterizes the lower part of this zone (Hajná et
13 2244 al., 2011), followed by late Ediacarian to Cambrian ocean floor stratigraphy and turbiditic
14 2245 accretionary wedge sequences of the Štěchovice group (Hajná et al., 2018; Ackermann et al.,
15 2246 2020). The uppermost unit is represented by a weakly metamorphosed Paleozoic section
16 2247 preserved at the core of the Praha or Barrandian Syncline. Metamorphic grade decreases
17 2248 upward from the basal Mariánské Lázně and Erbendorf-Vohenstrauß units through a thick
18 2249 Neoproterozoic succession until the core of the Praha Syncline (Peřestý et al., 2017).
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26 2250 Above, a medium to low-grade Neoproterozoic succession is followed by the
27 2251 Paleozoic of the Praha Syncline. This sedimentary pile starts with an Ordovician siliciclastic
28 2252 succession (485-443 Ma) followed by early Silurian graptolitic shales which continuously
29 2253 passes to late Silurian and Lower Devonian (Lochkovian and Pragian) limestones. The
30 2254 sequence terminates with siliciclastic deposition of the Givetian Srbsko Formation, which
31 2255 contains detrital zircons dated at 382 Ma (Vacek et al., 2019). Similarities between this
32 2256 sequence and that of the Iberian Autochthon had been claimed based mostly in stratigraphic
33 2257 and faunal correlations (Verneuil and Barrande, 1885; Delgado, 1908; Gutiérrez-Marco et al.,
34 2258 1999b). However, a study of Ordovician trilobites carried out by Ballèvre (2016) shows that a
35 2259 very limited number of species are common between the Barrandian and the Armorican and
36 2260 Iberian Autochthon, which points to the TBZ as a terrane separated from mainland Gondwana
37 2261 at the Cambro-Ordovician transition.
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49 2262 The first event affecting the TBZ produced upright folding and development of a NE-
50 2263 SW trending vertical fabric associated to crustal thickening at around 380 Ma (Hajná et al.,
51 2264 2012; Peřestý et al., 2017; 2020). The folding of the Praha syncline had to occur after the
52 2265 Givetian hiatus. Its age can correlate with formation of strong vertical cleavage affecting the
53 2266 eastern margin of the TBZ (Jílové Zone) which developed coeval with the syntectonic
54 2267 intrusion of arc-derived magmas of the Central Bohemian Pluton (Žák et al., 2005). Therefore
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2268 the shortening of the eastern TBZ and folding of Praha synforms can be indirectly dated at
2269 360-350 Ma (Hajná et al., 2012; Vacek and Žák, 2017).

2271 5.5. Variscan intrusive magmatism in the Bohemian Massif

2272 The main tectonic events have their counterpart in the granitic plutonism of the
2273 Bohemian Massif. The earliest magmatic rocks are calc-alkaline gabbros and tonalites
2274 intruding the eclogites of the Mariánské Lázně Unit at around 390-380 Ma (Deiller et al.,
2275 submitted), c. 375 Ma granodiorite stocks (Čistá pluton, Venera et al., 2000), and gneissified
2276 granitoids intruding the TBZ rocks further east (Mirovice and Sedlčany, Košler et al., 1993).
2277 These magmatic rocks are interpreted as a manifestation of earliest Variscan arc magmatism
2278 in the Bohemian Massif. However, large magmatic arc intrusions are dated at 354-337 Ma,
2279 and mostly occur at the Central Bohemian Plutonic Complex (Žák et al., 2014). These calc-
2280 alkaline magmas were generated during Upper Devonian to early Carboniferous subduction,
2281 first of oceanic lithosphere and then of the continental Saxothuringian crust beneath the
2282 Teplá-Barrandian Unit. For that reason, they are referred to as the Saxothuringian arc (Figs. 6
2283 and 7b). The juvenile magmas intruded SE to the Praha Syncline at 355-340 Ma (Sázava
2284 suite) while the magmatic pulse migrated eastward with magmas becoming progressively
2285 richer in potassium (Janoušek et al., 2004; 2010; Žák et al., 2014). These rocks can be
2286 compared with the intrusives of the tonalitic line of Limousin and the young ophiolites in the
2287 French Massif Central (Peiffer, 1986; Pin and Paquette, 2002; Faure et al., 2009) and the
2288 Vosges (Skrzypek et al., 2012).

2289 The next pulse of granitoid intrusions was that of the durbachites or Mg-K granitoids
2290 (ultra-potassic and potassic melasyenites and melagranites), intruded in the MZ to the SE of
2291 the Central Bohemian Plutonic Complex at 345-335 Ma (Janoušek and Holub, 2007; Žák et
2292 al., 2014). These rocks are also found in NW Iberia (López-Moro and López-Plaza, 2004;
2293 Gallastegui, 2005), the Vosges (Tabaud et al., 2015) and Corsica (Rossi et al., 2009). To the
2294 scale of the Variscan belt, von Raumer et al. (2014) consider them melting products of
2295 enriched subcontinental mantle related with the subduction of the Rheic slab into the mantle.
2296 A similar view is sustained by Schulmann et al. (2009, 2014b) and Janoušek et al. (2007),
2297 although for them, the durbachites may derive from the subducted crust of the STZ.

2298 Meanwhile, the northern margin of the Variscan Belt recorded intense activity. After
2299 opening of the Rhenohercynian basin, the RHZ collided with the STZ, initiating the Variscan

2300 dynamics that lasted between 360-320 Ma. Initially, preserved parts of the Rheic lithosphere
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22301 adjacent to the STZ were subducted to the south, followed by accretion of the Silurian-
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42302 Devonian arc (Zeh and Will, 2010). This formed the Rhenohercynian arc, starting at 360 Ma,
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62303 and whose activity lasted until *c.* 325 Ma with a magmatic peak at 340-330 Ma. Its rocks are
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82304 very poorly exposed in the MGCH (Odenwald and Ruhla inliers; Figs. 4, 6 and 7b).
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102305 Interestingly, their volcanic equivalents are widely exposed to the SW, represented by the
11
122306 abundant granitoids and ignimbritic volcanism in the Vosges (Tabaud et al., 2014), the
13
142307 layered basic intrusions and granitoids in the Schwarzwald, and the Tuffs Anthracifères in the
15
162308 northern Massif Central (Lardeaux et al., 2014).

172309 Post-collisional, mostly anatectic plutons characterize the last magmatic pulse. This
18
192310 event is associated with low-P/high-T metamorphism which followed a previous Barrovian
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212311 event associated with crustal thickening (Gerdes et al., 2000). Its most voluminous occurrence
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232312 is at the South Bohemian Batholith (335-300 Ma), which represents the SE limit of a ring of
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252313 many plutons surrounding the core of the Bohemian Massif (Schulmann et al., 2009; Žák et
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272314 al., 2014), of which the youngest are dated at *c.* 310-300 Ma.

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6. Morocco

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60 bounding the High Atlas. The segments of the Variscan belt exposed in southernmost

2332 Morocco and the northwestern corner of Mauritania (Zemmour area; Souttoufides of
1 2333 Villeneuve et al., 2015) and those defining the Mauritanides beyond the Reguibat Shield in
2 2334 western Mauritania (Villeneuve, 2005) are not included in this description.
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5 2335 6 7 8 2336 *6.1. Rif Domain* 9

10 2337 Pieces of the Variscan Belt are incorporated into the Alpine belt describing the
11 2338 Alboran Arc, with the Betic Range in the north (Spain) and the Maghrebides (Rif in Morocco)
12 2339 in the south (Do Couto et al., 2016). In the Betics, the sequences vary from little and no-
13 2340 metamorphic Ordovician to early Carboniferous sediments in the Maláguide Complex
14 2341 (Martín-Algarra et al., 2004) to metasediments difficult to separate from Mesozoic sediments
15 2342 with Alpine metamorphic imprint in the Alpujarride and Nevado-Filábride complexes
16 2343 (Alonso-Chaves et al., 2004; Puga et al., 2004). The deposition of the clastic protolith of the
17 2344 latter complex was recently determined using detrital zircon ages ranging between 349-282
18 2345 Ma (Santamaría-López and Sanz de Galdeano, 2018). Based on Carboniferous conodonts
19 2346 found in the lowest Nevado-Filábride Unit, Rodríguez-Cañero et al. (2018) correlate that
20 2347 Variscan basement with the Cantabrian Zone, while occurrences in the Alpujarride and
21 2348 Maláguide complexes, integrated in the Alborán Domain, are placed in the northern margin of
22 2349 a Paleotethys Ocean.
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25 2350 High-grade Variscan metamorphic rocks are present in both Betic and Rif cordilleras
26 2351 in southern Spain and Morocco. The metamorphic core of this previously thickened Cenozoic
27 2352 orogen was dismembered during Neogene extension forming several discontinuous outcrops
28 2353 along the Alboran Arc (Michard et al., 1997; Do Couto et al., 2016; Gimeno-Vives et al.,
29 2354 2020). The polymetamorphic Variscan outcrops are found in the Internal Zone of the Alboran
30 2355 Arc, namely in the Alpujarride-Sebtide complex, which form the metamorphic envelope of the
31 2356 Ronda in the Betic Cordillera and the Ceuta and Beni Bousera peridotite massifs in the Rif
32 2357 (*e.g.* Michard et al., 1997, Ruiz-Cruz and Sanz de Galdeano, 2013). The metamorphic pattern
33 2358 of these complexes is characterized by downward increase of metamorphic gradient with
34 2359 structurally deepest granulites, migmatites and higher gneisses and schists (Rossetti et al.,
35 2360 2010, Acosta-Vigil et al., 2014). These units reveal a polymetamorphic record of both
36 2361 Variscan and Alpine orogenic cycles for which the determination of age and pressure-
37 2362 temperature conditions remains problematic (Michard et al., 1997; Massone, 2014). However,
38 2363 it becomes increasingly evident that the granulite-migmatite unit of the Moroccan Sebtide
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2364 complex surrounding Beni Bousera peridotite preserves remnants of high-P/high-T late
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2 2365 Carboniferous-Permian metamorphism reaching *c.* 1.6 GPa at 800 °C (Bouybaouene et al.,
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4 2366 1998). These granulites and granite sills representing products of deep crustal anatexis were
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6 2367 dated at 300-290 Ma using U-Pb zircon method (Rossetti et al., 2010, 2020). Similar peak
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8 2368 pressure conditions of 1.2-1.4 GPa at 850 °C followed by reequilibration at 0.5-0.6 Gpa and
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10 2369 800-850 °C were determined in the anatectic Jubrique envelope of the Ronda massifs (Barich
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12 2370 et al., 2014). Here, the age of anatexis was determined by U-Pb zircon geochronology to 280-
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14 2371 290 Ma (Acosta-Vigil et al., 2014) confirming EMP monazite age *c.* 287 Ma from the same
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16 2372 unit (Massonne et al., 2014). These data corroborate the previously obtained metamorphic
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18 2373 300 Ma ages from amphibolite facies schists (Zeck and Williams, 2000; Zeck and
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20 2374 Whitehouse, 2002) and *c.* 283 Ma age of late Variscan pegmatite magmatism (Sánchez-Navas
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22 2375 et al., 2014) from the metamorphic core of the central Betic Cordillera. Similar *c.* 281 Ma
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24 2376 granite magmatism was also documented by U-Pb zircon geochronology of granite pebbles
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26 2377 from a Tertiary conglomerate lying on a Ghomaride nappe of the eastern Rif (Olivier and
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28 2378 Paquette, 2018).

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30 2379 Recently, ultrahigh-P conditions were discovered in diamond-bearing garnets
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32 2380 preserved in the migmatitic Jubrique complex in Spain (Ruiz Cruz and Sanz de Galdeano,
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34 2381 2014) and diamondiferous granulites of Ceuta inlier in Morocco (Ruiz Cruz and Sanz de
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36 2382 Galdeano, 2013). Even if the precise age of ultrahigh-P conditions remains currently unknown
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38 2383 the textural criteria indicate that it certainly preceded the main late Carboniferous-Permian
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40 2384 Variscan high-T event and is considered older than 330 Ma (Ruiz Cruz and Sanz de
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42 2385 Galdeano, 2014).

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44 2386 Altogether, the Variscan outcrops preserved in the Spanish and Morocco branches of
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46 2387 Alboran arc show similar lithological and metamorphic evolution suggesting the existence of
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48 2388 a high-P/high-T metamorphic belt before and during *ca* 300 Ma. Because of these similarities,
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50 2389 this belt can be considered as a stitching zone between Iberian Variscides to the north and the
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52 2390 Moroccan Meseta to the south. It coincides with the South Iberia Fault, which suggests that
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54 2391 the latter was a plate boundary more complex than other subparallel wrench faults, and
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56 2392 included convergence and probably extension during late Variscan evolution.

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2394 6.2. *Meseta Domain*

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22395 Although the main part of the Meseta is covered by Meso- and Cenozoic sediments,
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42396 several massifs allow the separation of various domains (Fig. 8), namely the Eastern and
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62397 Western Meseta, the later including the Coastal and Sehoul blocks (Piqué and Michard,
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82398 1989). Except for the Coastal Block, these domains are characterized by a deformation
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92399 stronger than in the Anti-Atlas, associated metamorphism, pre- and late orogenic magmatism
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112400 and weakly deformed continental late Carboniferous and Permian deposits. The
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132401 paleogeographic significance of these domains, limited by extensive shear zones (Michard et
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152402 al., 2008; Accoto et al., 2020), however remains debatable, and only the Sehoul Block have a
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172403 clearly distinctive signature (Tahiri et al., 2010; Pérez-Cáceres et al., 2017). The domain of
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192404 strong Variscan deformation also includes most of the basement of the Atlas Mountains
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212405 included within the South Meseta Zone, up to the South Meseta Front. It will be here
222406 described collectively with the Meseta Domain, yet this domain has Anti-Atlas affinities.
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252407 The preorogenic sequences can be compared with those of the Anti-Atlas. The lower
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272408 Paleozoic is similar although the Cambrian carbonates are thinner and thick Cambrian clastics
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292409 were deposited in graben structures fed by local sources and associated with tholeiitic
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312410 volcanism, which accounts for early to late Cambrian extension (Pouclet et al., 2008; Letsch
322411 et al., 2018). The graben system might have had a southward prolongation toward the Anti-
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342412 Atlas (Piqué, 2003). The Ordovician is comparatively thinner, with a marked distalisation
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362413 toward the NE across the Western Meseta (Baudin et al., 2003). Lower Ordovician red beds
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382414 comparable to those of the Armorican Massif and southern CIZ are reported in Central
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402415 Morocco and the Eastern Meseta, but are absent in the Anti-Atlas and nearly so in most of the
412416 Coastal Block (Ouanaimi et al., 2016). The Hirnantian glacial episode is widely recorded
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432417 throughout the Meseta, with the ice sheet having probably reached the Rehamna massif (Le
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452418 Heron et al., 2007), and an ice-distal turbiditic succession being recognized in the Tazeka
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472419 Massif (Ghienne et al., 2018). Extension continued during the early Paleozoic, marked by
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492420 intraplate volcanism in the Lower Ordovician (El Hadi et al., 2014) and the Silurian (El
502421 Kamel et al., 1998). Carbonate platforms developed in the Devonian. A more pronounced
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522422 tectonic activity was recorded during the Lower-Middle Devonian around the Variscan
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542423 Western Meseta Shear Zone (WMSZ; Fig. 8), where coarse continental conglomerates filled
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562424 restricted fault-bounded basins (Michard et al., 2008).

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582425 Differences in the style and age of deformation occur among the different domains.
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602426 The Eastern Meseta (EM) was affected by an early folding event, synmetamorphic, of early

2427 Carboniferous age. Detailed analysis of fold geometries at the scale of several massifs of the
1 2428 Moroccan Meseta shows that these folds are recumbent, with variable vergence: W to SW
2 2429 (Piqué, 2001; Michard et al., 2010a) and SE (Accotto et al., 2020). Axial traces delineate an
3 2430 arc convex toward the west (Piqué, 2001; Michard et al., 2010a), called here Eastern Meseta
4 2431 Arc (EMA). These recumbent folds and the associated subhorizontal schistosity have been
5 2432 attributed to the Bretonnic phase (Piqué, 1988, 1989, 2001; Michard et al., 2008; Ouanaimi et
6 2433 al., 2019), formed in a supra-subduction environment (Michard et al., 2008), or in a fore-arc
7 2434 setting (Accotto et al., 2020). Metamorphism related to axial planar schistosity of the early
8 2435 recumbent folds is of the greenschist facies, reaching amphibolite facies in the south. A
9 2436 younger folding event produced upright folds trending ENE-WSW, dated at late Visean-early
10 2437 Westphalian (*c.* 330-315 Ma), except in the west, where thrust sheets were emplaced
11 2438 westward in the Nappe Zone (NZ; Fig. 8). There, thin-skinned tectonics was active and coeval
12 2439 with synorogenic sedimentation during the same time interval, although early Variscan
13 2440 structures are also recognized (Michard et al., 2008).

25 2441 The Western Meseta (WM; Michard et al., 2010a) is divided into a western, almost
26 2442 undeformed part named the Coastal block, and an eastern part characterized by three principal
27 2443 Variscan massifs (Central, Rehamna, and Jebilet). This domain is separated from the Eastern
28 2444 Meseta by the Middle Meseta Fault Zone (MMFZ) and from the Coastal Block by the
29 2445 Western Meseta Shear Zone (WMSZ). The structural and Raman spectroscopy thermometry
30 2446 study of Lahfid et al. (2019) suggests that in the easternmost part of Western Meseta, the
31 2447 latest Devonian to early Visean folding event, reminiscence of the above-described tectonic
32 2448 evolution in the Eastern Meseta, was preceded by a high-T extensional Late Devonian event.
33 2449 The late Variscan evolution of the Western Meseta is debated, but NW-verging, late
34 2450 Carboniferous-Permian folds and thrust faults prevailed (Michard et al., 2010a). Chopin et al.
35 2451 (2014) proposed that a south- to SSW-verging shortening event occurred in the latest
36 2452 Carboniferous-earliest Permian, having no genetic connection with extensive Visean basins.
37 2453 This event is bracketed by Delchini et al. (2018) between 325-310 Ma, while a second
38 2454 deformation phase producing transpressional westward thrusting took place at 310-280 Ma in
39 2455 relation with the WMSZ. This scenario is more compatible with the late Carboniferous nappe
40 2456 tectonics recognized in the northern Anti-Atlas (Cerrina Feroni et al., 2010), as well as with
41 2457 the Anti-Atlas tectonic evolution (Baïdder et al., 2016). Later, an Early Permian orthogonal
42 2458 convergence occurred, and a post-tectonic pluton was emplaced at *c.* 275 Ma.

2459 The Coastal Block (CB), which includes the western parts of the Rehamna and Jebilet
1 2460 massifs as well as parts of the Western High Atlas (Fig. 8), is weakly deformed except close
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3 2461 to the WMSZ, a dextral shear zone where Cambrian to Lower Devonian sediments are
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5 2462 strongly deformed. The Paleozoic overlies a Proterozoic basement including Paleoproterozoic
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7 2463 relicts (El Houicha et al., 2018) but also Cadomian rocks: an orthogneiss has been dated a
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9 2464 600-585 Ma (Baudin et al., 2003). Early-middle Cambrian recorded alkaline to tholeiitic
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11 2465 volcanism as in the Anti-Atlas, and Lower Ordovician pillow lavas cropping out south of the
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13 2466 Rabat-Tiflet Fault Zone were probably generated in an extensional environment (El Hadi et
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15 2467 al., 2014). The Ordovician, coarser grained when compared to more eastern parts of the
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17 2468 Western Meseta, can be correlated with that of the Anti-Atlas. Lower Ordovician red beds are
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19 2469 scarce (Ouanaimi et al., 2016). Locally, a hiatus exists between the late Ordovician and the
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21 2470 Silurian and Devonian formations (Michard et al., 2008, 2010a; Ouanaimi et al., 2016),
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23 2471 whereas glaciogenics are still known laterally. At the WMSZ, Lower to Middle Devonian
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25 2472 sandstones and conglomerates rest unconformably upon Cambrian and Ordovician
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27 2473 formations, suggesting reactivation of pre-Variscan faults (Piqué and Michard, 1989).

28 2474 The Sehoul Block (SB; Fig. 8) is only represented in outcrops to the north but adjacent
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30 2475 to the Rabat-Tiflet Fault Zone (RTFZ). It consists of thrustured Cambro-Ordovician
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32 2476 metasediments which do not differ from other sequences of Morocco (Pérez-Cáceres et al.,
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34 2477 2017), but was thought having been affected by pre-Silurian metamorphism, deformation and
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36 2478 later granite intrusion, in consonance with old available isotopic dating (Piqué, 1989; Michard
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38 2479 et al., 2008). More recent U-Pb dating of the calc-alkaline Rabat granite yielded *c.* 367 Ma
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40 2480 (Tahiri et al., 2010), which together with the fact that the granite intruded previously folded
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42 2481 Cambrian sediments, would imply an orogenic event previous to the early Variscan folding in
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44 2482 the Eastern Meseta. It could be coeval with the early folds in the OMZ of SW Iberia.
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46 2483 According to Ouanaimi et al. (2016), the SB forms probably part of the peri-Gondwanan
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48 2484 realm in the continuation of the OMZ and SPZ of the Iberian Massif, and was displaced to its
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50 2485 present position by the dextral RTFZ (Fig. 8).

51 2486 Thick Upper Devonian-early Carboniferous flysh-type sediments include volcanic
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53 2487 rocks: tholeiitic to alkaline basalts in the Western Meseta and calc-alkaline in the Eastern
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55 2488 Meseta (Kharbouch et al., 1985; Kharbouch, 1994; Simancas et al., 2005; Delchini et al.,
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57 2489 2018). Recently, detrital zircon dating has evidenced a still unrecognized upper Devonian (*c.*
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59 2490 376 Ma) magmatic source for the Upper Devonian turbidites of the Eastern Meseta (Accotto
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61 2491 et al., 2020), which interestingly echoes the zircon record of SW Iberia (Pereira et al., 2017a;

2492 Pérez-Cáceres et al., 2017). Younger Variscan granitoids occur as small plutons sparsely
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2493 distributed in all the Meseta domains, and generally postdate the early Variscan folding phase.
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2494 Ages range 345-320 Ma (Oukemeni et al., 1995; Ajaji et al., 1998; Michard et al., 2008) and
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52495 compositions include calc-alkaline and metaluminous type I granitoids derived from
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72496 previously enriched mantle sources as well as type S with an important anatectic contribution.
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92497 This distribution, together with the Upper Devonian-early Carboniferous bi-modal
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112498 magmatism in the WM, was interpreted in terms of a SE-directed Upper Devonian-
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132499 Carboniferous subduction context by Michard et al. (2008). Small postkinematic plutons
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152500 occur mainly in the Western Meseta, dated at *c.* 305-280 Ma. (Michard et al., 2008).

16 172501 18 192502 *6.3. Anti-Atlas*

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222503 This domain beyond the South Meseta Front (Fig. 8) is characterized by inliers of
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242504 Precambrian basement, reworked during the Pan-African event, and a rather complete
252505 preorogenic sequence from the late Neoproterozoic to the early Carboniferous. The
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272506 sedimentary sequence, including late Ediacaran sediments, is 8-9 km thick in the west and 4-5
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292507 km in the east. The Paleozoic was deposited in shallow marine environments, and reflects the
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312508 latest Neoproterozoic-Cambrian-Lower Ordovician rift and post-rift sequence, with no major
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332509 unconformity at the Ediacaran-Cambrian boundary. Carbonate horizons alternating with fine-
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352510 grained detritals characterize the early Cambrian while thick sandstones and fine-grained units
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372511 occur from the middle Cambrian to the Ordovician (Destombes et al., 1985). Rifting is
382512 indicated by Cambrian basalts of tholeiitic and alkaline composition. The Hirnantian event is
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402513 well represented by glaciogenic deposits, tunnel valleys and glacial pavements up to the High-
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422514 Atlas, indicating that the end Ordovician ice sheet extended repeatedly across the Anti-Atlas
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442515 (Le Heron et al., 2007; Ghienne et al., 2014). The Paleozoic continues dominated by shelfal
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462516 shales with alternating sandstones and carbonates (Lubeseder, 2008), and extensional fault
472517 activity was recorded in the Upper Devonian (Wendt, 1985; Toto et al., 2008; Frizon de
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492518 Lamotte et al., 2013) leading to a so-called ‘disintegration’ of the Sahara platform. In the
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512519 latest Devonian and during the early Carboniferous, clastic sedimentation renewed, including
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532520 unconformable terrestrial deposits in places. Thick Viséan depocenters developed in the
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552521 eastern Anti-Atlas, associated with synsedimentary extensional faults and flysch-type deposits
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572522 including Ordovician to Tournaisian olistoliths, which are observed up to the adjacent Béchar
582523 and Ben Zireg area in Algeria (Frizon Lamotte et al., 2013; Benyoucef et al., 2015).
59
602524 Sedimentation appears more or less continuous through the Viséan (Alvaro et al., 2014), with

2525 no clear intra-Visean unconformity in stark contrast with the Meseta. The generalization of
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2526 carbonate mud mounds in the upper part of the Visean suggests a progressively decreasing
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2527 tectonic activity and basin infill (Wendt et al., 2001, Alvaro et al., 2014), the youngest folded
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52528 strata being late Visean in age. Unconformable, ‘molassic’, Upper Carboniferous deposits are
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72529 preserved in Algeria (Fabre, 2005) above the folded sequence.

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2530 Structurally, the Anti-Atlas is a thick-skinned domain characterized by antiforms
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112531 where the pre-Variscan basement crops out, and synforms occupied by the Paleozoic
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132532 sequence (Burkhard et al., 2006; Baidder et al., 2016). The basement inliers are frequently
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152533 bounded by strike-slip and reverse faults inherited from the early Paleozoic Gondwanan
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172534 paleomargin that were reactivated (inverted) during the Variscan event. The Neoproterozoic
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192535 basement was formed at the margin of the West African Craton (at *c.* 2 Ga), and both were
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212536 affected by the Pan-African orogeny (Ennih and Liégeois, 2001; Hefferan et al., 2014;
222537 Michard et al., 2017; Soulaïmani et al., 2018; Brahimî et al., 2018; Aït Lahna et al., 2020). In
23
242538 the western part, Variscan deformation appears stronger than in the east (Michard et al.,
25
262539 2010a). Metamorphism is absent in general but can be of very low-grade in the west, where
27
282540 axial planar cleavage developed locally. Fold interference patterns are common, with strikes
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302541 varying around N-S, NE-SW, E-W and NW-SE, which are interpreted as related to the
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322542 attitude of reactivated Panafrican to Cambrian faults (Robert-Charrue et Burkhard, 2008;
332543 Baidder et al., 2008).

34 35 362544 37 382545 **7. Correlation of the Variscan massifs**

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412546 Several of the zones described are represented or can be correlated with equivalents in
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432547 the adjacent and even distant massifs, while others are limited to a single massif or their
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452548 correlation is not straightforward. For instance, the Rhenohercynian Zone can be followed
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472549 probably from the eastern Bohemian Massif through the Rhenish Massif and the southern of
482550 the British Isles until SW Iberia. Conversely, the Cantabrian Zone has no equivalents in other
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502551 massifs, except possibly in the western Pyrenees, and the Allochthon, with its several groups,
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522552 is not represented in Morocco.

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542553 Figure 1 depicts the zone correlations in a late Variscan map reconstruction, and
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562554 Figure 9 shows the correlation of zones, groups and units as well as that of early Variscan and
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582555 Variscan orogenic magmatism for the main massifs. Figure 10 shows the timing of events in
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602556 table format detailed by massifs, zones and units, and grouped by significant tectonic,

2557 metamorphic and magmatic events. This figure is intended to offer an outline of the age of
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2558 most significant events in the different massifs, with the time intervals corresponding to those
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42559 described in the text. It also serves for the purpose of age correlation of events among the
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62560 massifs. Note however that differences may be misleading due to scarcity of isotopic dating,
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82561 time of dating, methodological approach or difficulty to precisely date high-P events followed
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102562 by decompression at high-T.

11
122563 The following sections highlight similarities among stratigraphy, pre-orogenic
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142564 magmatism, structural and metamorphic evolution of the massif described, analyzing
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162565 separately the groups and units forming the Allochthon from the underlying and surrounding
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182566 zones. The correlation of the Variscan orogenic magmatism is sketched in Figure 9 and has
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202567 been partly included in the description of the different massifs.

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222568 The particular similitude between the Iberian and Armorican massifs, explained by
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242569 their proximity, together with the fact that they occupy a central position in the Variscan Belt
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262570 determines here their use as elements of comparison for the other massifs. In this section, the
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282571 correlations are discussed separately for the Autochthon and Allochthon.

29 30 312572 322573 *7.1. Correlations in the Autochthon*

33 342574 *7.1.1. Links of Morocco with northern Variscan massifs (AA, EM, WM, CB, SB, CZ, CIZ,* 35 362575 *OVD, OMZ, CAD, NAD)*

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382576 A link between the Anti-Atlas and the Cantabrian Zone of the Iberian Massif was
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402577 suggested by Michard et al. (2010a). It can be established based on their common
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422578 interpretation as foreland belts and the general sedimentary and igneous evolution, although a
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442579 precise correlation of formations is not feasible. Both areas show a comparable ice-proximal
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462580 setting during the end-Ordovician, suggesting they belong to a same area relative to the ice
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482581 fronts, yet laterally distant. From a tectonic point of view they differ, being thin-skinned in the
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502582 whole CZ vs. thick-skinned in most of the central and eastern Anti-Atlas. May be, the reason
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522583 has to do with the existence of better detachment levels in the CZ, although they also
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542584 potentially exist in the Anti-Atlas at the Neoproterozoic-Cambrian transition. Other possible
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562585 explanations are the absence of faults related to the Cambrian-Ordovician rifting in the CZ,
57
582586 that such faults were not reactivated due to the relative orientation of the stress field and the
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602587 faults, or to the different amount of Variscan shortening, 17-25 % in the western Anti-Atlas

2588 (Michard et al., 2008) and more than 50 % in the CZ (Álvarez-Marrón, 1995; Ayarza et al.,
1 2589 1998).

3
4 2590 Concerning the Meseta, Michard et al. (2008) argued against the possibility that it
5
6 2591 became separated from Gondwana and disconnected from the Anti-Atlas during the opening
7
8 2592 of the Paleotethys, as proposed by Stampfli and Borel (2002). Using stratigraphic, faunal and
9
10 2593 paleomagnetic data, they support restricted displacements relative to Africa, at least for the
11 2594 main part of the Meseta.

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13
14 2595 Based on the assumption that Morocco remained in Gondwana during the Paleozoic,
15 2596 the similitude of the Carboniferous deposits with those of Iberia and the typology of Variscan
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17 2597 granitoids, Simancas et al. (2009) proposed a continuation of the Iberian Massif by the
18
19 2598 Moroccan Meseta. More precisely, they correlate the WM with the CIZ and the EM with the
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21 2599 WALZ and CZ, while the OMZ and SPZ would continue west of the present coast of
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23 2600 Morocco except for the Sehoul block.

24
25 2601 Michard et al. (2010a) consider zones of the Iberian Massif other than the CZ as
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27 2602 equivalent to those of the Meseta, without proposing a detailed correlation among them
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29 2603 except for the potential equivalence of the SPZ with the WM and CB. This last point is
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31 2604 debatable, as it is based on similarities of the Carboniferous fauna and magmatism, including
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33 2605 associated pyrrhotite-chalcopyrite ores. These can be explained by the important phase of
34 2606 extension that also affected the SPZ, OMZ and the southern part of the CIZ during the early
35
36 2607 Carboniferous. It is possible that Carboniferous sedimentation and magmatism were partly
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38 2608 controlled by subduction related with the closure of the Rheic or Rhenohercynian oceans. In
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40 2609 this scheme, the suture would be submerged, but could occur close to the coast (Michard et
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42 2610 al., 2010a).

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44 2611 Franke et al. (2020) proposed a correlation of the Moroccan Meseta and Coastal Block
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46 2612 with the OMZ of SW Iberia, based on the possible continuation of the Badajoz-Cordoba
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48 2613 Shear Zone by a suspect suture hidden under young deposits east of the Meseta.

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50 2614 A more reasonable correlation seems the one of Ouanaimi et al. (2016), based on the
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52 2615 Paleozoic sequence and focused on the extent of Lower Ordovician red beds. They proposed a
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54 2616 link of the Moroccan Meseta (EM and WM) with the Iberian CIZ and the Armorican CAD
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56 2617 and NAD, while the WALZ and CZ of Iberia would better correlate with the Anti-Atlas.
57 2618 Additional clues are provided by the zircon record, with abundant 1 Ga-old detrital zircons in
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59 2619 the Lower-Middle Ordovician clastics of both the Tazekka massif (northern part of the WM-

2620 EM boundary, Fig. 8; Accotto et al., 2019) and coeval strata from the CIZ and the CAD but
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2621 not the NAD (Stephan et al., 2019 and references therein). The correlation between the EM
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2622 and the CIZ is also supported by a structural criterion: the common existence of early
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2623 Variscan folds which in both zones delineate an arc, the EMA and CIA respectively. In fact,
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2624 such an arc may be a key element for understanding the geometry of the Variscan Belt, as
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2625 shown in the essay of retrodeformation tackled in the next section.

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On the other hand, the Coastal Block shows links with the Anti-Atlas (Michard et al., 2010a), but has also affinities with the OVD in southern Iberia and the NAD in France: i) the relatively old (600-580 Ma) Cadomian basement, equivalent to the lower Brioverian, cropping out in the three domains; ii) an Ordovician richer in terrigenous facies there than in the CIZ, CAD and EM; iii) a zircon record with local sources and a virtual absence of 1 Ga zircons, which echoes that of the NAD (Letsch et al., 2018; Dabard et al., 2021); iv) a relatively weak Variscan deformation, except in the proximity of shear zones.

The Sehouf Block could represent a fragment of Meguma for Michard et al. (2010a). A similar although alternative correlation is supported by the Devonian zircon record from the EM (Accotto et al., 2020), which suggest a sourcing from an Avalonian promontory which also would explain similar zircon populations at the limit between the SPZ and OMZ in SW Iberia (Pereira et al., 2012, 2017a; Pérez-Cáceres et al., 2017) and which might be represented by the SB. However, the Cambrian deposits of the SB are similar to those of the Western Meseta, suggesting paleogeographic proximity. And the early Variscan event represented by the calc-alkaline Rabat granite can be related with the first orogenic event in the OMZ.

Based on Ouanaïmi et al. (2016), and on the arguments given above, a correlation is proposed for the Sehouf Block with the OMZ, the Coastal Block with the OVD and NAD, the Meseta with the CIZ and the Anti-Atlas with the CZ (Figs.1 and 8).

The possible suture envisaged by Michard et al. (2010a) offshore but close to the coast could continue to the south in the northern Mauritanides, where a thrust belt with eastern vergence includes basic rocks in eclogite facies dated at 333 ± 25 Ma (Le Goff et al., 2001), and in central and southern Mauritanides, marked by oceanic assemblages at the limit between internal and external zones (Caby and Kienast, 2009). This may imply pulling apart of a peri-Gondwanan terrane, perhaps at the late Neoproterozoic, and closure of the intervening ocean during the Carboniferous. That suture would be that of NW Africa with a peri-Gondwanan terrane approximately placed where the terrane called Florida-Suwannee

2652 (Dallmeyer, 1989; Matthews et al., 2016) or Carolina (Winchester et al., 2002), pulled apart
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2 2653 from NW Africa during the Jurassic.

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6 2655 *7.1.2 Links among Iberian and French autochthons (CZ, CIZ, OVD, OMZ, CAD, NAD, SAD,*
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8 2656 *VM, SWM, MN, Sardinia)*

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10 2657 The Variscan zonation of the Iberian Peninsula is not fully represented in France. The
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12 2658 CZ and WALZ do not crop except perhaps in the westernmost Pyrenees (Figs. 1 and 4). The
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14 2659 OMZ and SPZ probably continue under the Meso- and Cenozoic cover of the Paris Basin, but
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16 2660 only the OMZ may crop out north of Brest, in the western edge of the Armorican promontory.
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18 2661 Conversely, the CIZ continues in the Armorican Massif and in the southern part of the Massif
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20 2662 Central, and the NW Iberian Allochthon can be correlated with the same domain in all the
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22 2663 French and related massifs.

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24 2664 The upper Brioverian of the CAD and the Schist-Greywacke Complex of the CIZ are
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26 2665 similar. The Black Series of the OMZ, equivalent to the lower Brioverian, does not crop out
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28 2666 in the CIZ, north of the OVD, but pebbles of black quartzite and amphibolite indicate erosion
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30 2667 of the Black Series, which probably underlies the Schist-Greywacke Complex. For the
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32 2668 Ordovician-Devonian sequence, a close correlation can be established between the CAD and
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34 2669 CIZ (Henry et al., 1974; Paris and Robardet, 1977). A distribution of Middle Ordovician
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36 2670 facies and fauna allowed Robardet et al. (1990) to separate 3 subdomains progressively more
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38 2671 distal to the south in the Armorican Massif. A similar zonation exists in the southern part of
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40 2672 the CIZ, where distality increases to the north, so that the subdomains may be correlated
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42 2673 along the Ibero-Armorican Arc.

43 2674 The Paleozoic sequence is so similar in Buçaco (western Portugal) and Crozon
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45 2675 (western Brittany) that both localities were probably very close to each other before the late
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47 2676 Variscan strike-slip tectonics and the opening of the Bay of Biscay (Robardet et al., 1990;
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49 2677 Young, 1990). Moreover, detrital zircon ages in Ordovician strata are similar in the CAD and
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51 2678 CIZ (Dabard et al., 2021). On the other hand, the synorogenic Carboniferous in the CAD is,
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53 2679 as in the CIZ, generally unconformable but often subparallel with the preorogenic sequence,
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55 2680 and synsedimentary tectonism in the CAD may be related with early Carboniferous extension,
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57 2681 as in southern Iberia.

58 2682 The correlation of the NAD with the CIZ is more problematic. The lower Brioverian,
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60 2683 characterized by bedded black cherts, is well represented in the OMZ by the Black Series,

2684 while in the CIZ it does crop out only in the OVD. The southern CIZ, including the OVD, is
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2685 characterized by more terrigenous and proximal Middle Ordovician facies, as the NAD
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2686 (Robardet et al., 1990). So, although a NAD/CIZ correlation can be envisaged, a more precise
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52687 link should be established only with the southern part of the CIZ (Figs. 1 to 5).
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82688 There is not a clear equivalent to the OMZ in the Armorican Massif. A possible
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92689 candidate is the eclogite-bearing gneissic unit in the Léon, at the northwestern end of the
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112690 Armorican promontory. Balé and Brun, (1986) interpreted it as derived from the SAD and so,
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132691 representing a detached piece of the Allochthon. But its structural position is the same as that
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152692 of the Central Unit in the OMZ: at the contact with the OVD in Iberia and with the NAD in
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172693 Brittany. Rolet et al. (1986) consider the possibility of underthrusting of the Léon under the
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192694 rest of the CAD at the Upper Devonian. We will assume this idea and also follow Schulz et al.
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212695 (2007) regarding the eclogite-bearing unit in the Léon as a potential link between the OMZ
222696 and the STZ of the Bohemian Massif (Fig. 1).
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252697 The continuation of the OMZ may be found further to the east, in the relatively small
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2698 outcrops of the Vosges and Schwarzwald to the north of the Lalaye-Lubine and Baden-Baden
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282699 shear zones. There, low- to medium grade Cambro-Ordovician to Silurian deposits are
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302700 overlain by marine Middle and Upper Devonian marine facies and early Carboniferous flysch
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322701 ascribed to the STZ of the Bohemian Massif, which is considered equivalent to the OMZ
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342702 (Pereira et al., 2006a; Linnemann et al., 2008).
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362703 In the SAD, the Autochthon can be compared with the Paleozoic of the northern part
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382704 of the CIZ by the occurrence of Cambro-Ordovician monotonous schists and porphyroids, the
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402705 possible existence of the Armorican Quartzite, the Middle Ordovician gray slates and the
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422706 Silurian black cherts. The latter, as well as some synorogenic deposits, could form part of the
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442707 Parautochthon, which includes metavolcanic Ordovician porphyroids, as in the Cabo Ortegal
452708 and the Portuguese complexes.
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482709 It is worth noting that while the Autochthon in the SAD correlates with that on the
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502710 northern CIZ east and south of the Iberian Allochthon, the CAD is rather comparable with the
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522711 central part of the CIZ. This correlation, as well as that of some allochthonous units, drove
532712 Ballèvre et al. (1992) to propose the existence of a suture between the CAD and SAD which
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552713 would continue by NW and central Iberia. But in NW Iberia, the suture represented by the
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572714 Middle Allochthon, is rootless, only preserved in klippen, and does not continue to the SE of
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592715 the CIZ (Fig. 2a). This is one of the problems we try to solve in this contribution.
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2716 In the FMC, the Montagne Noire (MN) can be correlated with the Pyrenees and
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2 2717 continue, in the other side of the Bay of Biscay, with the branch of the CIZ limiting to the east
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4 2718 and south the NW Iberian Allochthon (GTMZ) and forming also its relative autochthon. The
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6 2719 Paleozoic sequence in the Pyrenees and MN shows differences with the CIZ and CAD as, for
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8 2720 instance, the lack of the characteristic Armorican Quartzite, and younger or more lasting early
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10 2721 Paleozoic felsic magmatism, which is here reported up to the Upper Ordovician. While the
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12 2722 MN represents the only outcrop of the Autochthon in the FMC, the occurrence of the
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14 2723 Parautochthon in several tectonic windows suggest that the massif is partially or totally
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16 2724 underlain by an autochthonous crust comparable to that of the CIZ (Figs. 4 and 5b, c).

17 2725 In Sardinia, the Autochthon of Iglesias represents a less-deformed domain of the
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19 2726 northern Gondwana foreland. The presence of the Upper Ordovician Sardinic unconformity, as
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21 2727 in the Pyrenees (Casas, 2010) confirms that it forms part of the northern Gondwana
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23 2728 Mediterranean branch. This, as younging of Ordovician magmatism only reflect lateral
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25 2729 changes along the northern Gondwana paleomargin, whose western continuation through the
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27 2730 Pyrenees and the Bay of Biscay reaches the part of the CIZ underthrust beneath the GTMZ.
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29 2731 There are however differences with the Iberian Parautochthon, where no calc-alkaline
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31 2732 Middle-Upper Ordovician arc is clearly represented.

32 2733 33 34 2734 *7.1.3. Links among Bohemian and western Variscan autochthons (STZ, OMZ, CIZ, RHZ, SPZ,* 35 36 2735 *MSZ)*

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38 2736 Several zones defined in the Bohemian Massif can be continued to the west into the
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40 2737 French and Iberian Massifs. Starting with the STZ, its Autochthon has a similar structural
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42 2738 relationship with its allochthonous complexes as the CIZ in NW Iberia in relation with the
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44 2739 GTMZ Allochthon. This section discusses possible correlations following the order used to
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46 2740 describe the zonation, highlighting similarities and signaling differences.

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48 2741 A key correlation can be established between the STZ and OMZ of SW Iberia.
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50 2742 Ediacaran succesions, Cadomian arc magmatism and the transition to Cambro-Ordovician
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52 2743 rift-related magmatism are quite similar on both zones (Pereira et al., 2006a; Linnemann et
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54 2744 al., 2008) and also in the transitional OVD at the southern part of the CIZ (Bandrés et al.,
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56 2745 2004). Moreover, the 565 Ma glaciation identified in the Lausitz Block of the STZ has its
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58 2746 counterpart in the southern part of the CIZ, north of the transitional OVD (Linnemann et al.,
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60 2747 2018). The Paleozoic succesions are also comparable between the STZ and OMZ. Cambrian

2748 faunas in eastern Germany compare with those of France, Iberia and Morocco (Elicki, 2007),
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2749 with links pointing to the CIZ and the Montagne Noire. Upper Ordovician glacigenics exist in
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2750 the STZ and OMZ, as well as in the other autochthonous zones of Iberia (CIZ, WALZ and
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2751 CZ) and in the Central and Northern domains of the Armorican Massif.
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2752 Subsidence patterns calculated for the Barrandian, the STZ and the eastern part of the
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2753 CIZ in Iberia are comparable, less regarding the OMZ, but they are especially striking for the
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2754 Cambrian, and start to diverge from the Lower Ordovician, when separation of the future
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2755 Upper Allochthon took place (von Raumer and Stampfli, 2008). Even if the patterns are
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2756 considered similar during the rest of the Paleozoic, this can be attributed to symmetrical
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2757 evolutions of the continental margins of Gondwana on one side and the pulled apart terrane
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2758 on the other.
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2759 With these data, a broad link can be established among the Autochthon of Iberia,
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2760 France and Bohemia, namely the STZ, the CAD and NAD of western France and the Iberian
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2761 OMZ and CIZ. A more accurate correlation can be attempted using detrital zircon age spectra.
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2762 Linnemann et al. (2008) have shown that spectra are similar for Ediacaran, early Cambrian
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2763 and Lower Ordovician metasediments of the STZ and SW Iberia (OMZ and OVD). The
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2764 spectra in the CIZ, WALZ and CZ are also similar except for a significant Mesoproterozoic
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2765 population centered at 1000 Ma (Gutiérrez-Alonso et al., 2003; Martínez Catalán et al.,
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2766 2004b), which is absent or nearly so in the OMZ, STZ and NAD (Linnemann et al., 2008;
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2767 Dabard et al., 2021). The source for this population is the East African-Arabian Zircon
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2768 Province (Gómez Barreiro et al., 2007; Bea et al., 2010), fed by the eastern Gondwana super-
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2769 fan system (Meinhold et al., 2013; Stephan et al., 2019). The abundance of this population in
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2770 the CIZ, CAD, WALZ and CZ is related to the paleoposition of this part of the Iberian Massif
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2771 close to present northern Africa, whereas its scarcity or absence indicates a more westward
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2772 position of the OMZ and quite probably, of the STZ, both of which facing a more western
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2773 segment of Northern Africa in agreement with the faunal record (Gutiérrez-Marco et al.,
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2774 2002, 2016). Also it is worth noticing that STZ, NAD and WM share pieces of an Eburnean
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2775 (c. 2.05 Ga) basement most likely derived from the West African Craton (Pereira et al.,
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2776 2015b). This is in line with a relatively western derivation of the STZ of Bohemia, a character
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2777 that might have been acquired since the Cadomian (Pérez-Cáceres et al., 2017), and amplified
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2778 during the Variscan orogen.

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2779 Another affinity between STZ and OMZ is the existence of intrusives related with the
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2780 closure of the Rheohercynian Ocean. In the STZ they were emplaced in the MGCH at 360-
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2781 325 Ma, while equivalent rocks intruded at 355-335 Ma mostly in the SW of the OMZ and
2782 eastern SPZ (Figs. 2, 4 and 6). Also supporting the correlation, the eclogites in the Odenwald
2783 could have their counterpart in those of the Cubito-Moura Unit, in the southern OMZ.
2784 Eclogite-facies metamorphism is probably older than 363-350 Ma in the Odenwald (Scherer
2785 et al., 2002) and is dated at 371±17 Ma in the southern OMZ (Moita et al., 2005).

2786 Also, the high-P metamorphism of the Central Unit (377±19 Ma; Abati et al., 2018) is
2787 probably older than the 360-335 Ma high-P granulites of the Granulitebirge (Franke and
2788 Stein, 2000), the eclogites of the Erzgebirge (Schmädicke et al., 1995; Rötzler and Plessen,
2789 2010) and the blueschists of the Krkonoše dome (Žáčková et al., 2010; Konopásek et al.,
2790 2019). That suggests an early Carboniferous contrasting behavior: in the Bohemian Massif the
2791 STZ underwent intracontinental subduction in relation to the emplacement of the Allochthon
2792 and the continuation of convergence during the collisional stage (Závada et al., 2021 and
2793 references therein). Conversely, in the OMZ an early Carboniferous extensional episode
2794 produced normal faults and basins where detrital and carbonate rocks were deposited together
2795 with volcanics. After, late folds developed during the middle-upper Carboniferous, largely
2796 produced by transpression which was dextral in the STZ (Franke and Żelaźniewicz, 2002) and
2797 sinistral in the OMZ (Simancas et al., 2001; Díaz Azpiroz and Fernández, 2005), where
2798 reverse and left-lateral faulting continued until the Stephanian.

2799 While a correlation between the autochthonous STZ and OMZ is well established, an
2800 equivalent zone is not properly identified in the Armorican Massif, except possibly in the
2801 Léon region, although it may occur hidden under the cover of the Paris Basin (Fig. 4). For a
2802 similar reason, and due to its relative position, no equivalent to the STZ occurs in the FMC.
2803 Closer to the Bohemian Massif, limited outcrops in the northern part of the Vosges and
2804 Schwarzwald massifs represent small pieces of a wide STZ covered by post-Variscan
2805 sediments or submerged under the English Channel between the Iberian and Bohemian
2806 massifs (Figs. 1, 4 and 6).

2807 Concerning the RHZ, it occupies the whole Rhenish Massif and continues to the west
2808 in the southern part of the British Isles and in the SPZ of SW Iberia (Fig. 1). There, as in the
2809 Bohemian and Rhenish massifs, the sutures of the Rheic and Rhenohercynian oceans are more
2810 or less hidden by the complexity of Variscan tectonics. In the NW of the Bohemian Massif
2811 the RHZ is mostly covered, with its main outcrop occurring at the Harz Mountains (Figs. 1
2812 and 6). Further to the east, the MGCH and the Northern Phyllite Zone (Fig. 6) were continued

2813 by Franke and Żelaźniewicz (2000, 2002) following the gentle arc delineated by the
2814 outcropping STZ, but the continuation of the RHZ to the east is uncertain.

2815 It has been proposed that the RHZ surrounds the rest of the Bohemian Massif and
2816 continues by the MSZ (Kossmat, 1927; Tollmann, 1982), so tracing a tight Bohemian Arc.
2817 This option was discussed by Franke and Żelaźniewicz (2002), who developed a model of
2818 interaction of the Variscan Belt with Baltica that first formed the gentle Bohemian Arc,
2819 opened about 90°, which was followed by a large dextral displacement along the Moldanubian
2820 Thrust and then by NW-SE faults along the Baltic margin of Laurussia. In their model, the
2821 Silesia Terrane Assemblage (the basement of the MSZ) was situated to the SE of Baltica and,
2822 as Avalonia, represents the southern margin of Laurussia. These authors stress the
2823 stratigraphic similarity and evolution between the RHZ and MSZ during the Devonian and
2824 Carboniferous and the probable southern continuation of equivalent terranes in the Balkans
2825 and northern Turkey.

2826 Avalonian affinities of the MSZ basement were described by Mazur et al. (2010), and
2827 the suture between the Brunovistulian block and the Moldanubian Zone was placed at the
2828 Upper Devonian by Mazur et al. (2010) and Jastrzębski et al. (2013), and interpreted as that of
2829 the Rheic Ocean. However, flysch development continued during the Carboniferous, rather
2830 pointing to the Rhenohercynian Ocean. As for the RHZ and the SPZ, that boundary could
2831 include both sutures. Southward continuation of the Brunovistulian block terranes of
2832 Avalonian derivation until the Serbo-Macedonian Massifs in the southern Carpathians and
2833 Balkans was proposed by Mazur et al. (2010) and Spahić and Gaudenyi (2019).

2835 *7.2. Correlations in the Allochthon*

2836 *7.2.1. Links among Iberian and French allochthons (GTMZ, CAD, SAD, FMC, VM, SWM, 2837 MTM, Corsica, Sardinia)*

2838 Starting with the CAD, the characteristics described by Lardeux and Cavet (1994) for
2839 the St. Georges-sur-Loire Unit, as well as those of the southern “blocky” subunit of Cartier
2840 and Faure (2004) are also typical of the NW Iberian Parautochthon (Ribeiro, 1974; Dias da
2841 Silva et al. 2014a, b; González Clavijo et al., 2016). So, this unit will be considered as
2842 Parautochthon, overthrust upon the Autochthon of the rest of the CAD. However, there is
2843 not a clear relationship between this Parautochthon and the contiguous Allochthon of the

2844 SAD to the south of the Nort-sur-Erdre Fault, as the strike-slip character of the fault may have
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22845 put into contact pieces originally distant.

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42846 Slices of a Parautochthon comparable to that of the GTMZ in NW Iberia occur in the
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62847 SAD in Vendée, SE of the Bois-de-Cené and underlying Lower Allochthon units (Figs. 4 and
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82848 5a). It includes Lower Ordovician porphyroids, Silurian ampelites and black cherts, and
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102849 synorogenic deposits. But the most striking similarities of this domain are established for the
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122850 Allochthon. There are not significant differences between the Lower Allochthon in Galicia
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142851 (Malpica-Tui and Órdenes complexes) and that of Champtoceaux and Vendée: same
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162852 lithologies, protolith ages and metamorphic evolution. Only the age of high-P metamorphism
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182853 seems to be younger by 10 Ma in France (360 vs. 370 Ma), but dating of high-P rocks by U-
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202854 Pb methods is always challenging because zircon may grow close to the temperature peak, so
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222855 postdating the pressure peak (Fernández-Suárez et al., 2007; Paquette et al., 2017).

232856 As in NW Iberia, the Middle Allochthon in the SAD is formed by ophiolitic
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252857 complexes or units of oceanic affinity. Cambro-Ordovician ophiolites are represented in the
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272858 Audierne Bay Complex, while the Ile-de-Groix and Bois-de-Cené units can be compared with
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292859 the continental-oceanic transitional, upper crustal units of Malpica-Tui and Órdenes
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312860 complexes. High-P metamorphism is in this case coeval (370-360 Ma). Middle Allochthon
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332861 units of Lower Devonian age have not been dated in the SAD, although the age of two
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352862 possible ophiolitic units is unknown. The ophiolitic unit bounding the Champtoceaux
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372863 Complex is a good candidate to compare with Lower Devonian ophiolites in NW Iberia, but
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392864 its protoliths seem 20 Ma younger (380 vs. 400 Ma). Would that be an indication of the
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412865 lifespan of the Devonian Ocean? It is possible, because its closure has only a minimum age
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432866 limit, given by the high-P metamorphism of the Lower Allochthon and Cambro-Ordovician
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452867 Middle Allochthon, dated around 370 Ma.

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472868 Concerning the Upper Allochthon, the Essarts Unit is made of Cambro-Ordovician
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492869 gabbroic and granitic intrusions and nearly contemporaneous high-T metamorphism perhaps
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512870 resulting from magmatic underplating. This is comparable to the intermediate-P Upper
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532871 Allochthon of NW Iberia. However, the eclogitic event suggests correlation with the high-
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552872 P/high-T units. In NW Iberia, no high-P evidences have been found in the intermediate-P
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572873 units, but traces of a Cambro-Ordovician high-T event in the high-P/high-T units are indicated
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592874 by *c.* 500-485 Ma monazites (Fernández-Suárez et al., 2002, 2007). The fact that the Cambro-
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612875 Ordovician and Devonian metamorphic events are so well preserved in the SAD supports the
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632876 contention that intermediate-P/high-T (NW Iberia) or low-P/high-T (Armorican Massif) units

2877 form part of the same coherent terrane with the high-P/high-T units of both massifs, and that
2878 no suture exists between the two groups.

2879 The Audierne Bay Complex includes a few tectonic slices with eclogites and high-P
2880 granulites comparable to the high-P/high-T units in NW Iberia, and also an alternation of
2881 metagreywackes and amphibolites that might present the intermediate-P units of the Upper
2882 Allochthon. Two other units in the SAD representing the uppermost Allochthon (Mauges and
2883 Roc-Cervelle) may have their equivalents in NW Iberia, but correlation is hindered because in
2884 the Upper Allochthon of the GTMZ neither rocks younger than *c.* 510 Ma, nor Cadomian
2885 basement have been found. Metamorphism does not help, as it is Cadomian in the SAD and
2886 Cambro-Ordovician in Iberia. However, the intermediate-P event in the latter (495-482 Ma)
2887 has been dated only in high-T rocks: it cannot be excluded that low-grade Cadomian
2888 metamorphism affected the metasediments.

2889 For the FMC, the Parautochthon consists of Paleozoic, low-grade metasediments and
2890 volcanics of northern Gondwana derivation, as in NW Iberia, where it was also locally
2891 migmatized during late Variscan extension and voluminous granitic plutonism. The Lower
2892 Gneiss Unit can be compared with the Iberian Lower Allochthon by its lithological
2893 association, which includes tholeiitic and alkaline rocks revealing Cambro-Ordovician
2894 extension. Although widely developed Barrovian metamorphism is dated from 360 Ma
2895 onward, previous high-P metamorphism has been identified, and partial melting dated at 375-
2896 365 Ma probably reflects exhumation following subduction. This points to a younger age for
2897 the high-P event in the Lower Gneiss Unit in relation to the subduction-related event in the
2898 Upper Gneiss Unit.

2899 There is a consensus that the two gneissic units are separated by a suture, which is
2900 generally interpreted to occur in the Leptyno-Amphibolitic Complex, where the high-P rocks
2901 abound (Burg and Matte, 1978; Ledru et al., 1989). Vanderhaeghe et al. (2020) individualize
2902 a Middle Allochthon consisting of micaschists associated with an ophiolitic assemblage
2903 preserving low-T eclogitic relicts (up to 2.9 GPa, 660 °C). But it seems included in the
2904 Leptyno-Amphibolitic Complex or at least, the separation from it is unclear. The lithological
2905 association and metamorphic conditions of their Middle Allochthon remind those of the
2906 Cambro-Ordovician oceanic supracrustals in NW Iberia and the Armorican Massif (Ballèvre
2907 et al., 2014). The Lower Devonian ophiolites of the GTMZ Middle Allochthon are also not or
2908 poorly represented in the FMC, where an equivalent unit may exist in the Limousin above the
2909 Lower Gneiss Unit (Girardeau et al., 1986, 1994; Berger et al., 2005).

2910 The best correlation can be established between the Upper Gneiss Unit and the Upper
1
2 2911 Allochthon: same lithologies, igneous protolith ages (Cambro-Ordovician) and metamorphic
3
4 2912 evolution. Parts of the Leptyno-Amphibolitic Complex share with the catazonal Iberian Upper
5
6 2913 Allochthon the Lower Devonian age of high-P metamorphism. No ultrahigh-P metamorphism
7
8 2914 has been described in NW Iberia, but at least 1.9-2.0 GPa were calculated for eclogites and
9
10 2915 ultramafics of the Cabo Ortegal Complex (Mendia, 2000; Mendia and Gil Ibarra, 2005).

11
12 2916 Above the Leptyno-Amphibolitic Complex, but still in the Upper Gneiss Unit,
13
14 2917 migmatitic paragneisses and orthogneisses devoid of high-P relics may represent the high-
15
16 2918 grade upper units of the Upper Allochthon. The uppermost part of the nappe pile in the FMC,
17
18 2919 consisting of low- to medium grade Cambrian greywackes, quartzites and rhyolites, have also
19
20 2920 an equivalent in the Upper Allochthon of the Órdenes and Morais complexes.

21
22 2921 The units of the Central Vosges and the Schwarzwald gneiss complexes have been
23
24 2922 compared with the Leptyno-Amphibolitic Complex of the FMC, particularly the part between
25
26 2923 the Varied and Granulite units (Pin and Vielzeuf, 1988; Fabriès and Latouche, 1988;
27
28 2924 Lardeaux et al., 2014). So, these units might be considered part of an Upper Allochthon
29
30 2925 similar to that of NW Iberia and to the Teplá-Barrandian Zone of the Bohemian Massif.
31
32 2926 However, the young age of high-P metamorphism in the Schwarzwald (*c.* 340 Ma; Kalt et al.,
33
34 2927 1994; Marschall et al., 2003), the fact that granulites possibly represent vertical extrusions at
35
36 2928 mid-crustal levels, and the ultra-potassic magmatism dated at 345-335 Ma rather suggest
37
38 2929 correlation with the Moldanubian Zone of the Bohemian Massif (Schulmann et al., 2002;
39
40 2930 Skrzypek et al., 2014).

41
42 2931 Nevertheless, the Variscan evolution can still be correlated with that of the FMC
43
44 2932 because: i) the ophiolites of the Ligne des Klippes in the VM are equivalent to the Brévenne
45
46 2933 Unit, a back-arc related with the subduction of the Saxothuringian Ocean (Faure et al., 1997;
47
48 2934 Schulmann and Gayer, 2000; Lardeaux et al., 2014); ii) the overlying Visean flysch and the
49
50 2935 Devonian-Carboniferous sediments and volcanics of the metasedimentary belt separating the
51
52 2936 two gneiss complexes in the SWM can be equivalent to the Tuffs Anthracifères, related to the
53
54 2937 subduction of the Rhehercynian Ocean, whose influence was probably responsible for the
55
56 2938 subsequent voluminous granitic plutonism. Affinities of the VM and SWM with both the
57
58 2939 FMC and the Bohemian Massif are not surprising given their position. The differences and
59
60 2940 the possible causes have been discussed by Lardeaux et al. (2014), and will be dealt with
61
62 2941 below in the correlation with the Bohemian Massif.

2942 The massifs of the southern Variscan Belt occur in continuation of, and their units can
1
2 2943 be correlated with the southern FMC. The internal zones are represented in the MTM by the
3
4 2944 Central and Eastern domains which together seem to represent the Upper Allochthon,
5
6 2945 although they might also include a dismembered suture according to Schneider et al. (2014).
7
8 2946 In the Western Domain, one of the thrust sheets bears a rift-related succession with alkaline
9
10 2947 rocks possibly representing the Lower Allochthon, although no high-P has been described. In
11
12 2948 any case, the Middle Allochthon is missing or hidden in the Leptyno-Amphibolitic Complex,
13
14 2949 as in the FMC. The Central Domain presently underlies an inverted early Carboniferous
15
16 2950 synorogenic flysch which is in turn overlain by a Paleozoic sequence including Silurian black
17
18 2951 shales (Bronner and Bellot, 2000). The original stacking order was the opposite, with the
19
20 2952 Eastern Domain above and the Western Domain below. So, the Western Domain may
21
22 2953 represent the Parautochthon and/or the Autochthon.

23
24 2954 In Corsica, most metamorphic outcrops represent a Leptyno-Amphibolitic Complex
25
26 2955 similar to that of the FMC, with eclogites and high-P/high-T granulites, which can be
27
28 2956 assimilated to the Upper Allochthon. The uppermost Allochthon crops out in northern Corsica
29
30 2957 as a low-grade Neoproterozoic and Paleozoic sequence that Rossi et al. (2009) interpreted as
31
32 2958 different from that of the Autochthon in SW Sardinia and the Montagne Noire, and deposited
33
34 2959 in a peri-Gondwanan microcontinent. It may find its equivalents in the uppermost Allochthon
35
36 2960 of NE Iberia (Órdenes and Morais complexes), the SAD (Mauges and Roc-Cervelle units) and
37
38 2961 the FMC (Thiviers-Payzac Unit).

39
40 2962 In Sardinia, the Internal Zone represents the high-P/high-T part of the Upper
41
42 2963 Allochthon, whereas the Internal Nappes of the Nappe Zone may correlate with the Lower
43
44 2964 Allochthon on the basis of its eclogite-facies event in absence of high-P granulites. From a
45
46 2965 structural point of view, the External Nappes constitute a Parautochthon, which is coherent
47
48 2966 with their metamorphic Paleozoic sequence representing in this case only the preorogenic part
49
50 2967 of the succession.

51
52 2968 In the External Massifs of the Alps, the metamorphic evolution of the third domain
53
54 2969 described by Ballèvre et al. (2018), characterized by the high-P eclogitic/granulitic event, may
55
56 2970 be compared to that of the FMC, VM and SWM (von Raumer and Neubauer, 1993).
57
58 2971 Consequently, it is correlated with the Upper Allochthon even if not all available ages of the
59
60 2972 high-P event yield the Lower Devonian age obtained in these and other Variscan massifs. The
61
62 2973 Chamrousse Ophiolite may represent the same ocean whose relicts have been preserved in the
63
64 2974 Middle Allochthon of NW Iberia and in the Audierne Bay Complex of the Armorican Massif.

2975 The ascription of the remaining domains is more problematic. The association of a Devono-
1
2 2976 Carboniferous bimodal suite with conglomerates underlying the ophiolite (second domain),
3
4 2977 suggest a correlation with the Morvan Arc of the FMC. Perhaps also with the migmatitic
5
6 2978 gneisses, early Carboniferous amphibolite and gabbro boudins in the southern SWM and/or
7
8 2979 with the Devono-Carboniferous sediments and volcanics of the metasedimentary belt
9
10 2980 separating the two gneiss complexes in the SWM. Finally, the undated low- to medium-grade
11
12 2981 schists of the first domain could represent the upper units of the Upper Allochthon (the option
13
14 2982 adopted in Fig. 4) because they are considered part of the polymetamorphic group by von
15
16 2983 Raumer et al. (1993), but might also represent the Parautochthon or even the Autochthon.

17 2984 18 19 2985 *7.2.2. Links among Bohemian and western Variscan allochthons (STZ, TBZ, MZ, VM, SWM,* 20 21 2986 *FMC, SAD, GTMZ)*

22
23 2987 The correlation described among the three groups of allochthonous units of the GTMZ
24
25 2988 and the SAD (Ballèvre et al., 2014) can be extended to the STZ allochthons based on the
26
27 2989 lithological association, structural sequence and ages. All the Iberian and Armorican groups
28
29 2990 are well represented in the STZ, although not always the three allochthons occur in one single
30
31 2991 massif.

32
33 2992 In the western branch of the STZ, the Münchberg massif exhibits the most complete
34
35 2993 succession. Its nappe stack recalls that of Cabo Ortegal and Bragança complexes. There, as in
36
37 2994 Münchberg, the Parautochthon consists of a synorogenic flysch with an early Carboniferous
38
39 2995 maximum depositional age, overlain by a relatively distal preorogenic Paleozoic sequence
40
41 2996 (Dias da Silva et al. 2014a; Martínez Catalán et al., 2008, 2016). In Münchberg, the Lower
42
43 2997 Allochthon is not present while in Cabo Ortegal it is only represented by a small unit. The
44
45 2998 only identified oceanic units witnessing the Middle Allochthon are Devonian phyllites,
46
47 2999 prasinites and amphibolites equivalent to the Lower Devonian ophiolites of NW Iberia.
48
49 3000 Finally, the Upper Allochthon is represented mainly by the high-P/high-T units, with only a
50
51 3001 small unit of the intermediate-P group in Cabo Ortegal, and no equivalent in the western STZ.

52 3002 The three allochthons are also well represented along the eastern branch of the STZ,
53
54 3003 from the lower part of the nappe pile south of the Krkonoše Massif to the upper part in the
55
56 3004 Góry Sowie Massif. A small low-grade unit partially synorogenic represents the
57
58 3005 Parautochthon, while the South Krkonoše Unit depicts the lithological association and Late
59
60 3006 Devonian high-P and relatively low-T metamorphism of the Lower Allochthon in Iberia and

3007 the Armorican Massif. To the east, the Kaczawa Complex represents the Middle Allochthon,
1 3008 with the rocks of oceanic affinity being of Cambro-Ordovician, Silurian and Devonian age, of
2
3 3009 which, the first and the later are also identified in NW Iberia (Martínez Catalán et al., 2020).
4
5 3010 Upper Devonian-early Carboniferous flysch with chaotic assemblages and tectonic mélanges
6
7 3011 in the Kaczawa Complex may have an equivalent in the Cabo Ortegal mélange, formed in an
8
9 3012 Upper Devonian subduction zone (Martínez et al., 2012; Arenas and Sánchez Martínez,
10
11 3013 2015). A basaltic unit of probable Late Devonian age (Seston et al., 2000), may represent a
12
13 3014 series of young ophiolites existing in the French Massif Central (Brévenne; Génis; Pin and
14
15 3015 Paquette, 2002; Faure et al., 2009) and the Vosges (Ligne des Klippes; Skrzypek et al., 2012;
16
17 3016 Lardeaux et al., 2014), but not identified in NW Iberia. Further east, the Góry Sowie Massif
18
19 3017 represents the high-P/high-T part of the Upper Allochthon, with identical protolith ages,
20
21 3018 metamorphic evolution and structural position as the high-P/high-T units of the GTMZ in NW
22
23 3019 Iberia and the Münchberg Massif in Germany (Martínez Catalán et al., 2020; Tabaud et al.,
24
25 3020 2020).

26 3021 In the STZ allochthons of Münchberg, Wildenfels, Frankenberg, Krkonoše and Góry
27
28 3022 Sowie massifs, Upper Allochthon units showing an intermediate-P Cambro-Ordovician event
29
30 3023 are absent. In NW Iberia, the Armorican Massif and French Massif Central, these units
31
32 3024 occupy the uppermost position of the nappe pile, so that their absence here can be explained
33
34 3025 by erosion. Moreover, this absence is compensated by the dominant occurrence of low-P
35
36 3026 Cambro-Ordovician metamorphism in the TBZ and MZ, which together represent the Upper
37
38 3027 Allochthon in the Bohemian Massif. Only in the Orlica-Śnieżnik Dome the STZ autochthon is
39
40 3028 surrounded by rocks of the MZ which in the eastern part bear the imprint of Cambro-
41
42 3029 Ordovician medium-P/high-T metamorphism (Staré Město Unit; Štípská et al., 2001).

43 3030 In the west of the TBZ, the basal association of low-grade metasediments, basic rocks,
44
45 3031 acid tuffs and serpentinites under the Mariánské Lázně and Erbenhof-Vohenstrauß units can
46
47 3032 be correlated with the phyllites and prasinites of Münchberg, and represent the Middle
48
49 3033 Allochthon. For the rest of the TBZ, a correlation may be established with the Upper
50
51 3034 Allochthon of the GTMZ (NW Iberia), SAD and FMC.

52 3035 The Mariánské Lázně, Erbenhof-Vohenstrauß units and the Góry Sowie Massif
53
54 3036 underwent a high-P/high-T metamorphic event during the Lower Devonian, followed by
55
56 3037 Middle Devonian exhumation in amphibolite facies. The same happened in the lower units of
57
58 3038 the Upper Allochthon in the GTMA (Cabo Ortegal, Órdenes and Bragança complexes), SAD
59
60 3039 (Essarts and Audierne Bay units) and the FMC (Leptyno-Amphibolitic Complex of the Upper
61

3040 Gneiss Unit). Above, the Teplá Crystalline Complex recorded a history similar to that of the
1
2 3041 upper group of the Upper Allochthon in the Órdenes and Morais complexes, with the
3
4 3042 difference that the Cambro-Ordovician metamorphic event (495-485 Ma) is of low-P in Teplá
5
6 3043 and of intermediate-P in NW Iberia. The Essarts Unit in the SAD did also record an
7
8 3044 Ordovician high-T event before Devonian eclogitization. In NW Iberia and the TBZ,
9
10 3045 Devonian shortening under Barrovian P/T conditions was followed by a low-P event
11
12 3046 connected with extension (Gómez Barreiro et al., 2006; Peřestý et al., 2017). The
13
14 3047 metamorphic contrast between the structurally upper and lower groups of the Upper
15
16 3048 Allochthon (Cambro-Ordovician low or intermediate-P above, Lower Devonian high-P/high-
17
18 3049 T below) is quite strong in all cases, accentuated by extensional shear zones and faults.

19 3050 A thick Neoproterozoic succession with Cadomian arc-related and Cambro-
20
21 3051 Ordovician rift-related magmatism characterizes the lower part of the TBZ (Dörr et al. 2002;
22
23 3052 Drost et al. 2004) whose uppermost unit is a rather complete and weakly metamorphosed
24
25 3053 Paleozoic succession preserved at the core of the Praha or Barrandian Syncline. The
26
27 3054 metamorphic grade decreases upward from the basal Mariánské Lázně and Erben-
28
29 3055 Vohenstrauß units through the Teplá Crystalline Complex (Peřestý et al., 2017) and a thick
30
31 3056 Neoproterozoic succession, until the core of the Praha Syncline.

32 3057 The Neoproterozoic sequence of the TBZ is compared with that of the Mauges Unit in
33
34 3058 the SAD, while in the GTMZ, the age of the high-grade metasediments is unconstrained. For
35
36 3059 the Paleozoic succession, several species of Middle Ordovician trilobites of the uppermost
37
38 3060 Allochthon in the SAD are also found in the TBZ, while only one is found in the Iberian
39
40 3061 Autochthon (Ballèvre, 2016), thus supporting separation from mainland Gondwana at
41
42 3062 Cambro-Ordovician times. In the FMC, the Paleozoic is represented by the Cambrian
43
44 3063 Thiviers-Payzac Unit. In the GTMZ, the Paleozoic succession is terrigenous and monotonous
45
46 3064 in the uppermost units of the Órdenes and Morais complexes, and its age is constrained to the
47
48 3065 middle-upper Cambrian by detrital zircons and intrusive magmatism (Fernández-Suárez et al.,
49
50 3066 2003; Fuenlabrada et al., 2010; Arenas et al., 2016). The latter, dated at 510-490 Ma, is
51
52 3067 bimodal, with arc affinities in the upper units (Andonaegui et al., 2002, 2012) and extension-
53
54 3068 related in the lower units of the Upper Allochthon (Galán and Marcos, 1997).

55 3069 A link of the TBZ with the Upper Allochthon of the northern small Variscan massifs
56
57 3070 has been proposed for the Central Vosges and Schwarzwald gneiss complexes (Pin and
58
59 3071 Vielzeuf, 1988; Fabriès and Latouche, 1988; Lardeaux et al., 2014), but also discussed, due
60
61 3072 in part to the younger ages of high-P metamorphism there (Schulmann et al., 2002; Skrzypek

3073 et al., 2014). However, correlation is straightforward with the southern small massifs: the
1
2 3074 Leptyno-Amphibolitic Complex is recognized in the Central and Eastern domains of the
3
4 3075 Maures, Corsica and the Internal Zone of Sardinia. That correlation can be extended at least to
5
6 3076 the third domain described in the External Massifs of the Alps by Ballèvre et al. (2018),
7
8 3077 characterized by a polymetamorphic evolution including eclogites or high-P granulites. Very
9
10 3078 interesting is the uppermost allochthonous unit of northern Corsica, with Neoproterozoic
11
12 3079 terrigenous sediments, interbedded metabasalts and a rather complete Paleozoic sequence,
13
14 3080 with a possible equivalent in the Mauges Unit of the SAD and perhaps also in the upper parts
15
16 3081 of the TBZ.

17 3082 Concerning the MZ, we consider it as a distinct element but still forming part of the
18
19 3083 Upper Allochthon. A back-arc setting has been proposed to explain its Silurian and Devonian
20
21 3084 volcano-sedimentary sequences, while the fore-arc could be represented by the TBZ and by
22
23 3085 the tectonic assemblage resulting from subduction of the Saxothuringian Ocean (Schulmann
24
25 3086 et al., 2009).

26 3087 Due to differences in the evolution between the Massif Central and Bohemian massifs,
27
28 3088 Lardeaux et al. (2014) proposed a division of the Moldanubian Zone between western and
29
30 3089 eastern parts, establishing as a possible limit the Bray Fault. That leaves the VM and SWM in
31
32 3090 the eastern Moldanubian, even when the authors interpret the northernmost FMC and the
33
34 3091 Southern Vosges as part of the same unit, as suggested by Jung and Roques (1952). However,
35
36 3092 similarities of the Central Vosges and the Schwarzwald gneiss complexes with the
37
38 3093 Moldanubian exist, namely the relatively young age of high-P metamorphism, the possible
39
40 3094 vertical extrusions of granulites to mid-crustal levels and the early-middle Carboniferous
41
42 3095 ultra-potassic magmatism (Kalt et al., 1994; Schulmann et al., 2002; Marschall et al., 2003;
43
44 3096 Skrzypek et al., 2014).

45 3097 It seems that the VM and SWM share characteristics and events with both the
46
47 3098 Bohemian and French Central massifs, in agreement with their relative geographic positions.
48
49 3099 Taken together, their tectonic, metamorphic and magmatic evolution seems to change
50
51 3100 gradually from east to west. For the Variscan magmatism, the change is abrupt between the
52
53 3101 eastern and western parts of the FMC, an additional argument against the role of the Bray
54
55 3102 Fault. These changes are related with the subduction of the Saxothuringian and
56
57 3103 Rhenohercynian zones under the Allochthon which to the west (SAD and GTMZ) apparently
58
59 3104 did not occur.

3105

1
23106 7. 3. *Mid-Variscan suture and Mid-Variscan Ocean*

3
43107 The Upper Allochthon was a peri-Gondwanan terrane and part of it was a Cambro-
5
63108 Ordovician magmatic arc. The possibility that it was formed by more than one terrane cannot
7
83109 be excluded. However, the ocean separating the Upper Allochthon from the Gondwanan
9
103110 Autochthon seems unique, as it exhibits a similar history in the Middle Allochthon of the
11
123111 whole Variscan Belt. This favors that the overlying Upper Allochthon was also a single
13
143112 terrane, even if it consisted of different paleogeographic environments, including fore arc,
15
163113 back-arc, thinned continental crust, shallow continental platform and emerged land.

17
183114 The intervening ocean formed part of the Rheic realm, but was not the Rheic in a strict
19
203115 sense because the latter was defined as the ocean separating Laurussia and Gondwana, and
21
223116 Laurussia includes three continental masses of which Avalonia is the one in contact with the
23
243117 Variscan Belt to the south of the Midlands and Brabant massifs (Nance and Linnemann,
25
263118 2008). So, the Rheic suture is that of (or hidden by) the Rhenohercynian Zone and extends
27
283119 westward to the South Portuguese Zone. It has also been claimed as the suture between the
29
303120 Moldanubian and Moravo-Silesian zones (Mazur et al., 2010; Jastrzębski et al., 2013). The
31
323121 ocean represented by the Middle Allochthon can be called using the names of the massifs
33
343122 where it is identified, as for instance, the Galicia-Brittany-Massif Central Ocean of Matte
35
363123 (2001, 2007). But if the same suture continues up to the Bohemian Massif, Saxothuringian
37
383124 and Sudetian should be added. The resulting name would be exceedingly long.

39
403125 A short name would be more suitable. But for instance, Saxothuringian Ocean is of
41
423126 doubtful validity for the whole belt as the STZ was strictly defined in the Bohemian Massif
43
443127 and northernmost Schwarzwald and Vosges (Kossmat, 1927). Similarly, South Armorican
45
463128 Ocean (Paris and Robardet, 1994) refers only to one of the large Variscan massifs. A better
47
483129 name for the ocean represented by the Middle Allochthon was Medio-European (Ledru et al.,
49
503130 1989; Cartier and Faure, 2004; Faure et al., 2009; Lardeaux et al., 2014), used in contributions
51
523131 where the authors consider only one ocean in addition to the Rheic/Rhenohercynian.

53
543132 The common feature of the suture represented by the Middle Allochthon is that it
55
563133 exists in all the large massifs strictly representing the Variscan Belt, *i.e.* the large central and
57
583134 western European massifs. And the suture occurs roughly in the middle of the belt. Mid-
59
603135 Variscan suture would then be a better term, the disappeared sea could be referred to as the
61
62
63
64
65

3136 Mid-Variscan Ocean, and the whole ensemble of terranes forming the allochthonous
1
23137 accretionary prism may be called the Mid-Variscan Allochthon.

3138 5 63139 **8. Restoring late Variscan deformation**

8 93140 *8.1. Scope and revision of age data*

10
113141 Crustal/lithospheric thickening characterizes the early Variscan (Devonian) and first
12
133142 stages of Variscan deformation (early Carboniferous) in Europe and Morocco. It is associated
14
153143 with convergence: early Variscan high-P metamorphism related with subduction and
16
173144 Barrovian metamorphism associated with collision. The late Variscan stages (late
18
193145 Carboniferous) recorded the extensional collapse of the orogen, as well as important wrench
20
213146 movements and orocline development. Extension is characterized by crustal thinning and low-
223147 P/high-T metamorphism, while wrenching did not cause significant changes in crustal
23
243148 thickness, neither a characteristic metamorphic imprint except retrogradation recorded locally.
25
263149 Nevertheless, transcurrence affected the orogenic crust by dislocations that hinder the original
27
283150 correlation among units and terranes, and also obscure the links between local and global
29
303151 processes.

31
323152 A way to cope with the effects of transcurrence is carrying out a retrodeformation of
33
343153 wrench displacements and arcs. Ideally, it would show the position of the different massifs
35
363154 and zones before late Variscan movements, and inverting their sequence, the successive steps
37
383155 would show a part of the temporal evolution of the belt. However, going back by steps that
39
403156 represent real time snapshots would require more precise geochronology than available at
41
423157 time. As well, the amount of displacement or strain should be known for strike-slip faults and
433158 shear zones, and for the domains among them, and this is not the case for most of them.

44
453159 Also important are the problems of partition of deformation to the scale of the belt,
46
473160 with more than one component involved. It is known that extension in the internal zones was
48
493161 contemporaneous with contraction in the more external ones (Faure et al.; 2009; Martínez
50
513162 Catalán et al., 2009), and that strike-slip shear zones were coeval with extensional and/or
52
533163 compressional structures (Martínez Catalán et al., 2014; Edel et al., 2018). Moreover, each
54
553164 particular shear zone may include different amounts of transpression, and may rotate during
563165 its activity in relation with oroclines or within wider shear zones whose limits may be
57
583166 imprecise.

3167 For these reasons, restoring wrenching deformation to the scale of the belt will be
1
2 3168 necessarily approximate, retrodeformation can only be partial and the successive steps might
3
4 3169 not represent the situation at a given time interval. The approach followed is to restore the
5
6 3170 main strike-slip shear zones and faults in steps marked by cross-cutting relationships to the
7
8 3171 map scale. The younger structures are restored first, and lately, the late Variscan internal
9
10 3172 deformation of a large part of the Variscan Belt is restored considering it as a broad dextral
11
12 3173 ductile shear zone based on the relationship of ductile shear zones with late folds (Iglesias
13
14 3174 Ponce de León and Choukroune, 1980; Gumiaux, 2003; Gumiaux et al., 2004) and the fact
15
16 3175 that they are always present in the Allochthon and the surrounding Autochthon. The
17
18 3176 restoration of arcuate geometries (oroclines) is very imprecise, but it is one of the goals of this
19
20 3177 contribution, and also one that yields the more interesting results.

21
22 3178 Age data, although not always precise, provide a time frame to constrain Variscan
23
24 3179 wrench-type deformation. In the Massif Central, Armorican and Maures-Tanneron Massifs,
25
26 3180 plutonism related with Variscan convergence and collision started at the Middle-Upper
27
28 3181 Devonian and extended until the early Permian. Much of that magmatism was synkinematic
29
30 3182 with the activity of transcurrent shear zones, the oldest of which are dated at c. 370 Ma in the
31
32 3183 South Armorican Domain and the Massif Central (Rolin et al., 2009). These authors link the
33
34 3184 370-315 Ma wrenching activity with indentation of continental microplates during Variscan
35
36 3185 collisional tectonics. Shear zones coeval with the emplacement of old plutons were sealed by
37
38 3186 younger ones, and were frequently reactivated later. Rolin et al. (2009) distinguished two
39
40 3187 families, the older of which is mostly dextral, striking N100°-130° E, and formed during the
41
42 3188 Famennian-early Viséan (370-340 Ma). The younger is middle Viséan-Namurian in age (340-
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44 3189 315 Ma) and includes dextral shear zones striking N150°-160° E and N130° E, and conjugate
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46 3190 sinistral shear zones striking N0°-90° E.

47
48 3191 The older family of Rolin et al. (2009) affects the allochthonous units and locally its
49
50 3192 relative autochthon, and can be related to the emplacement of the Mid-Variscan Allochthon,
51
52 3193 which in NW Iberia ended at 345-340 Ma (Dallmeyer et al., 1997). The younger family is
53
54 3194 represented in the other the Variscan massifs, but mostly clustering at the end of the middle
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56 3195 Viséan-Namurian time interval or even later. Ductile deformation in the Elbe Shear Zone was
57
58 3196 dated by Hofmann et al. (2009) at 337-323 Ma (considering the maximum time span allowed
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60 3197 by the error bars). The old age is based on the sigmoidal shape of a deformed granitic pluton,
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62 3198 while the young age corresponds to an undeformed granite close to the EFZ and cross-cutting
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64 3199 previous faults and cleavage. A wider time interval of 350-315 Ma is provided by Žák et al.
65

3200 (2014) also based on relations with granitoids. In the Armorican Massif, granitoids
1 3201 synkinematic with dextral shearing may span a 330-300 Ma interval in the northern part of the
2 3202 SAD (Bernard-Griffiths et al., 1985b; Gumiaux et al., 2004), and are somewhat younger in
3 3203 the south (310-290 Ma; Carron et al., 1994). The Bray Fault is dated around 320 Ma, but the
4 3204 evidences are imprecise, based on Ar-Ar dating of biotite in a drilled mylonitic orthogneiss
5 3205 (Matte et al., 1986), and on inversion of normal faults in the Culm Basin, not directly on the
6 3206 Bray Fault itself (Leveridge and Hartley, 2006).

13 3207 The dextral PTSZ of western Iberia, which for half of its length limits the GTMZ
14 3208 Allochthon, continues by the ductile dextral NEF and SASZ-N of the Armorican Massif, also
15 3209 limiting the Mid-Variscan Allochthon there (Figs. 1, 2 and 4). In Iberia, ages of
16 3210 metamorphism and synkinematic granites indicate that late folding related with dextral
17 3211 transcurrence occurred at 315-305 Ma (Dallmeyer et al., 1997; Rubio Pascual et al., 2013a;
18 3212 Martínez Catalán et al., 2014), while late dextral shear zones are dated at 310-305 Ma
19 3213 (Gutiérrez-Alonso et al., 2015) and even continued until 290 Ma in the Pyrenees (Cochelin et
20 3214 al., 2017). The PTSZ, NEF and SASZ-N were displaced during dextral transpression occurred
21 3215 in the Pyrenees at 310-290 Ma, coeval with lateral flow, dome emplacement and upper crust
22 3216 thickening (Cochelin et al., 2017).

31 3217 May be that the Ar-Ar dating in the CIZ reflects cooling ages a few million years
32 3218 younger than shear zone motion. Also, isotopic dating of undeformed granitoids does not
33 3219 necessarily yield a minimum age of fault activity, as later deformation can be concentrated
34 3220 outside the igneous massifs. But still, significant deformation in the Elbe Shear Zone is older
35 3221 than the mentioned Iberian equivalents. Perhaps, the Elbe zone was deformed by SW-NE
36 3222 dextral shearing older than 310 Ma, which also affected NW Iberia and the Armorican Massif,
37 3223 and was later reoriented to NW-SE during a younger shearing event as the one dated at *c.*
38 3224 310-290 Ma in Iberia and France.

46 3225 Transpression in the Pyrenees was probably associated to the NPF, one of a group of
47 3226 large NW-SE dextral faults and associated shear zones which also includes the SMF, SIF, BrF
48 3227 and EFZ (Fig. 1). They are oblique to the general SW-NE trend of the belt and overprint all
49 3228 previous structures. Consequently, they are thought to represent the latest significant
50 3229 movements and must be restored before the folds and shear zones they cut across. Some of
51 3230 these faults cannot be studied, because they are hidden or were reactivated as plate boundaries
52 3231 later, as the SIF between Iberia and Africa, or the NPF. Others, like the Elbe Fault Zone and
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3232 Bray Fault have been dated, but their age contradicts the structural evidence of being the
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23233 latest significant Variscan structures.

3
43234 Sinistral motion is of relatively minor importance in the Variscan Belt, except in
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63235 southern Iberia. There, it is post Upper Namurian (Bashkirian), according to relationship with
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83236 Carboniferous sedimentation (Simancas et al., 2001; Expósito et al., 2002; Martínez Poyatos,
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103237 2002). This is somewhat older (*c.* 320-310 Ma) than dated motion in the PTSZ and SAD. But
113238 even older dextral deformation has been reported in the latter (330 Ma; Bernard-Griffiths et
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133239 al., 1985b; Guineberteau et al., 1987) and in the Elbe Shear Zone (Hofmann et al., 2009),
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153240 while in the Massif Central, both dextral and sinistral movements yielded ages of 345-320
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173241 (Gay et al., 1981; Costa et al., 1993; Rolin et al., 2009) and 320-300 Ma (Ledru et al., 1994b).

193242 8.2. Step by step restoration

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243244 The process is illustrated in Fig. 11. It starts from a modified version of the late
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263245 Carboniferous reconstruction of Figure 1, where the late-Variscan movement of the large
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283246 NW-SE faults has been removed (Fig. 11a). While the real geometry and displacement of the
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303247 late Variscan NPF in Figure 1 is rather hypothetical, its reconstruction by matching the GTMZ
313248 and SAD is easy and dependable. Moreover, two clear magnetic anomalies that were probably
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333249 in continuation to each other, one in the eastern part of Órdenes Complex (Martínez Catalán
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353250 et al., 2012) and the other Baie d’Audierne (asterisks in Figs. 1 and 11a), help fitting both
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373251 sides of the Bay of Biscay. For the rest of the large NW-SE Variscan faults, a simple
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393252 translation has been applied to match the Variscan zones at each side.

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413253 This first step shows that the Ibero-Armorican Arc (IAA) is much tighter for the
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433254 zonation of the Iberian and Armorican massifs than for a group of dextral shear zones which
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453255 seem to occur in continuity to each other. The next step (Fig. 11b) consists of removing part
463256 of the IAA rotating these massifs in opposite directions to get the Porto-Tomar Shear Zone
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483257 (PTSZ) lined up with other significant dextral structures, the Nort-sur-Erdre Fault (NEF) and
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503258 the Lalaye-Lubine and Baden-Baden shear zones (LLSZ-BBSZ).

51
523259 The key for restoration is the similitude of the Paleozoic sequence between Buçaco
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543260 (western Portugal) and Crozon (western Brittany), interpreted as reflecting proximity before
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563261 the late Variscan strike-slip tectonics and the opening of the Bay of Biscay (Robardet et al.,
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583262 1990; Young, 1990). These localities are separated in Figure 11a by *c.* 800 km, while in
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603263 Figure 11b, the distance is reduced to *c.* 500 km due to the partial restoration of the IAA,

3264 which at this stage had a lesser amplitude. That slip has been removed in Figure 11c, as well
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2 3265 as most of the curvature of the Bohemian Arc. The two sinistral shear zones of southern Iberia
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4 3266 (BCSZ and SISZ) are continued to the north displacing the OMZ which would continue to the
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6 3267 south of England filling a space under the waters of the Bristol Channel which connects the
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8 3268 OMZ and STZ.

9
10 3269 These two sinistral shear zones, which overprint the southern Iberian suture and the
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12 3270 contact between the OMZ and CIZ, are an anomaly on the general dextral kinematics of late
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14 3271 Variscan tectonics. Probably, they developed between two indentors, one to the west, related
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16 3272 to the Grand Banks of Newfoundland and responsible for the hidden arc between Iberia and
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18 3273 Morocco (Simancas et al., 2005, 2009), and the other to the east, along the north Gondwana
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20 3274 margin, which would be at the origin of the Ibero-Armorican Arc. Removing the slip of the
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22 3275 BCSZ and SISZ yields a still relatively complicated map (Fig. 11d) where the Mid-Variscan
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24 3276 Allochthon looks as a triangular indentor splitting the CIZ in two branches which continue to
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26 3277 the north by the CAD and NAD of the Armorican Massif and to the south by the
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28 3278 Mediterranean province of the Autochthon.

29
30 3279 The CAD is fully bounded by dextral shear zones, and the NAD probably too. We
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32 3280 interpret the two domains as strike-slip duplexes, of which the first is more deformed than the
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34 3281 latter (Le Gall, 1999; Gumiaux et al., 2004). Figure 11c shows that the two key localities of
35
36 3282 Buçaco and Crozon could have been adjacent to each other yet before restoring sinistral
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38 3283 shearing, but Figure 11d shows that the NAD duplex has not been restored yet. This is done in
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40 3284 Figure 11e, which places the NAD close to the Moroccan CB, in agreement with the
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42 3285 correlation established between them and with the OVD of southern Iberia. Note that in the
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44 3286 restoration, the OVD forms part of the NAD (Fig. 11d) before being displaced by the BCSZ.
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46 3287 The translation of the little deformed NAD would have sheared the CAD, a deformation that
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48 3288 has been restored in Figure 11e.

49
50 3289 If that interpretation of the CAD and NAD is correct, joining Buçaco and Crozon does
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52 3290 not account for the full dextral slip along the main shear zone or zones, because a previous
53
54 3291 slip is necessary to move the NAD to the north of the CAD. The restoration shows that this
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56 3292 must have occurred before motion in the BCSZ and SISZ, because they came from the west
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58 3293 of these two sinistral shear zones and rest presently to the east. But the BCSZ, and most
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60 3294 probably the SISZ were later cut across and dragged by the dextral PTSZ.

3295 In addition to late Variscan shear zones and faults, the whole Variscan Belt was
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23296 affected by relatively tight late Variscan folds with a well developed axial planar cleavage,
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43297 with the exception of the NAD. In the CIZ for instance, late folds are similar in wavelength
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63298 and amplitude to those of the CAD, and coeval slaty cleavage is widely developed in pelitic
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83299 formations. Late folds and associated slaty cleavage are commonly recognized structures also
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103300 in the STZ and the TBZ. In areas where metamorphism had reached the high-T greenschist
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123301 facies and higher, late folds are also common, and the associated fabric is a well-developed
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143302 crenulation cleavage.

153303 On the grounds of plate tectonics evolution (dealt with in the next section), we
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173304 consider that late folding reflects distributed dextral shearing, while strike-slip shear zones
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193305 and faults reflect more localized deformation of the same generalized dextral shearing. In the
20
213306 CAD, a shear value $\gamma = 1.5$, equivalent to a shear angle of 56.3° , has been calculated from the
22
233307 angular relationship between the shearing direction and the mean attitude of the associated
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253308 cleavage (Gumiaux, 2003; Gumiaux et al., 2004). This implies 50 % shortening perpendicular
26
273309 to the cleavage if deformation took place only by simple shear.

283310 A 10-25% shortening is necessary to develop an incipient cleavage (Ramsay and
29
303311 Huber, 1983), 30 % until it becomes apparent (Siddans, 1972), and 50 % is considered a
31
323312 starting value for a well developed slaty cleavage (Wood, 1974). Because late folds are
33
343313 widespread and their axial planar cleavage is generally well developed, Gumiaux's shear
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363314 value is used for the last step: restoring distributed dextral shearing along the Variscan Belt
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383315 (Fig. 11f).

39
403316 The new picture shows a Mid-Variscan Allochthon that is wedge-shaped in map view,
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423317 but the wedge is open and realistic, although not necessarily so much open. A right or acute
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443318 angle shape may have resulted only from mechanisms related to the emplacement.
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463319 Famennian-Lower Carboniferous dextral shearing at the northern limb of the wedge (Rolin et
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483320 al., 2009) may have contributed to make it more acute during its emplacement. Also, in terms
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503321 of the critically tapered wedge models (Davis et al., 1983), the thicker is a thrust sheet relative
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523322 to thinner adjacent parts, the more probabilities it has of spreading above its relative
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543323 autochthon forming salients (Boyer, 1995). The Mid-Variscan Allochthon was really thick,
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563324 especially by the contribution of the Upper Allochthon, which in turn was a microcontinent of
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583325 limited lateral extent. This may have created a wedge shaped, even tongue shaped salient
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603326 simply during the emplacement. So, the actual shape of the Allochthon before late Variscan

3327 folding may be that of Figure 11f if the later, generalized ductile dextral shear had a value $\gamma =$
1 3328 1.5, and more acute if that value was lower.

3 43329 The last step of restoration represents a stage just before the onset of generalized late-
5 63330 Variscan dextral shearing, when late folds had not developed and early folds wrap around the
7 83331 indenter delineating an open arc which later became the CIA-EMA. An approximate position
9 103332 of the CAD, NAD and CB is indicated. At this stage, the Allochthon had been emplaced and
11 123333 the SPZ and RHZ were developing thrust belts, which occurred after 340 Ma. Previous
13 143334 wrench deformation dominantly dextral, described in the SAD, FMZ and the Bohemian
15 163335 Massif, was probably related to the emplacement of the Allochthon. A comment on available
17 183336 ages for transcurrent deformation is given in the previous subsection, and an attempt to fit the
19 203337 different steps of late Variscan deformation in a time frame is done in the next section.

21 22 23 24 3339 **9. Plate tectonics and Variscan evolution**

25 26 27 3340 *9.1. Sources and approach*

28 29 30 31 32 33 34 35 36 37 38 39 40 41 42 43 44 45 46 47 48 49 50 51 52 53 54 55 56 57 58 59 60 61 62 63 64 65
3341 The Pangea configuration proposed by Wegener (1929) for the late Carboniferous,
3342 commonly known as Pangea A is generally accepted before the opening of the Atlantic
3343 Ocean. However, inconsistency between previous paleomagnetic poles of Laurentia and
3344 Gondwana led Irving (1977) to propose an alternative Pangea B model where Laurentia
3345 occupied a more western position, implying large dextral shearing to explain the transition
3346 from models B to A. Relative dextral displacement between Laurussia and Gondwana
3347 continued in principle from the late Carboniferous to the Permian and perhaps up to the
3348 Lower Triassic (Muttoni et al., 2009; Gallo et al., 2017).

3349 Various plate reconstructions have been attempted in the last decades and are
3350 particularly useful in combination with software utilities allowing the user to introduce
3351 changes in plate kinematics, such as GMap and GPLates. However, while abundant for the
3352 Mesozoic and Cenozoic, reconstructions are scarce and show significant differences for the
3353 Paleozoic. As a way to validate the partial retrodeformation of the Variscan Belt, it has been
3354 checked against one of such a models: the Domeier and Torsvik (2014) reconstruction for the
3355 late Paleozoic (410-250 Ma). Figure 12 was built using GPLates
3356 (<https://sourceforge.net/projects/gplates/files/gplates/2.2/>, accessed 5/5/2021). For information
3357 concerning GPLates, see Boyden et al. (2011).

3358 The reconstruction is based on Domeier and Torsvik (2014) and uses the following
1
2 3359 reconstruction geometries and tree of Matthews et al. (2016):

3
4 3360 Matthews_etal_GPC_2016_ContinentalPolygons.gpmlz

5
6 3361 Matthews_etal_GPC_2016_Coastlines.gpmlz

7
8 3362 Matthews_etal_GPC_2016_410-0Ma_GK07.rot. The latter allows reconstruction from
9
10 3363 420 Ma. For the Cambro-Ordovician maps (500-470 Ma), the reconstruction tree
11
12 3364 Master2010_II.rot and animations 500.A97 and 470.A97 of GMap2015 were used
13
14 3365 (<http://www.iggl.no/resources.html>, accessed 5/5/2021).

15
16 3366 The choice of Domeier and Torsvik (2014) reconstruction is because it is mostly based
17
18 3367 on paleomagnetism, although this method does not allow establishing paleolongitudes. The
19
20 3368 authors acknowledge that paleomagnetic poles are not abundant for the time interval of their
21
22 3369 reconstruction, and they also use eruptions in Large Igneous Provinces and kimberlite
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24 3370 emplacements in relation with a mantle reference frame for provisionally establishing
25
26 3371 paleolongitudes, as well as additional geological and paleontological data. Moreover, the
27
28 3372 reconstruction of Domeier and Torsvik (2014) counts on the work done by Mathews et al.
29
30 3373 (2016) in the files ready to use for GPlates. We prefer this reconstruction also because it fits
31
32 3374 better our knowledge about the Variscan evolution in comparison with, for instance, one of
33
34 3375 the most popular alternatives, that of Stampfli et al. (2013).

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36 3376 Domeier and Torsvik (2014) stated that a number of Gondwana paleomagnetic data
37
38 3377 reveal a relatively stationary position about the South Pole near Angola during the Middle and
39
40 3378 Upper Devonian. Meanwhile, Devonian paleomagnetic data for the main Variscan massifs,
41
42 3379 although scarce, indicate that they remained at constant latitude. Then, to explain early
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44 3380 Carboniferous collision with Baltica, they move them thousands of kilometers from northern
45
46 3381 Africa to Laurentia during the Devonian and turn them back to northern Africa during the
47
48 3382 Carboniferous, with the Iberian Massif returning to practically to the same position but after
49
50 3383 having experienced a counterclockwise rotation of 90°. And all that happened without
51
52 3384 changes in the shape of the massifs, which behaved as rigid elements.

53
54 3385 According to stratigraphic similarities between the Autochthon of the Iberian Massif
55
56 3386 and Morocco, the existence of a suture between them had not been considered (Michard et al.,
57
58 3387 2008, 2010a). However, high-P metamorphism and new ages of high-T metamorphism in the
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60 3388 Rif Domain and Betics point to a convergent plate boundary being active there toward the end
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62 3389 of the Variscan cycle. Even so, correlation among zones of both sides suggests that relative
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64 3390 movements did not modified essentially the relative position of Iberia and Africa. Based on
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3391 that correlation and in the evidence that strong deformation characterizes the Variscan
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2 3392 Massifs, we will keep the model of Domeier and Torsvik (2014) for the major plates,
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4 3393 inserting our own model regarding more specifically the evolution of the Variscan Belt.

5 6 3394 7 8 3395 *9.2. Reconstruction*

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10 3396 The main continental masses and previously described significant elements of the
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12 3397 Variscan Belt are shown in Figure 12 in orthographic projection for different ages. For the
13
14 3398 two Cambro-Ordovician reconstructions, a projection centered in the South Pole has been
15
16 3399 used, while for the tothers, covering the 400-290 Ma time interval in 10 stages, a fixed
17
18 3400 geographic grid shows the absolute and relative motions of the elements (“features” in
19
20 3401 GPlates’ slang). Features unrelated to Variscan evolution, such as northern and eastern
21
22 3402 territories (Siberia, China, etc.) are omitted. For a better understanding of the Variscan
23
24 3403 evolution, Figure 13 shows cross-section snapshots of continents, terranes and plate
25
26 3404 boundaries corresponding to the time frames of Figure 12. The sections vary for each frame in
27
28 3405 that figure to show different aspects of the tectonic evolution until 320 Ma.

29 30 3406 31 32 3407 *9.2.1. Terrane individualization*

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34 3408 The reconstruction starts at 500 Ma, when future Avalonia was still part of Gondwana
35
36 3409 and a wide passive continental margin developed to the north and NW of present Africa. A
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38 3410 piece of the outermost margin of NW Gondwana was being pulled apart by slab roll back of a
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40 3411 subduction zone which created a magmatic arc in the side facing the Iapetus Ocean. This was
41
42 3412 probably the same responsible for the Avalonia volcanic arc (Pickering and Smith, 1995,
43
44 3413 1997), which continued further to the south along more than 2000 km in present South
45
46 3414 America forming the Famatinia-Puna Ordovician arc (Ducea et al., 2017). The detached piece
47
48 3415 formed later the Mid-Variscan Upper Allochthon. It was a microcontinent whose
49
50 3416 individualization may have started at 520-510 Ma and where magmatic and metamorphic
51
52 3417 peaks were reached around 500 Ma.

53 3418 Placing the future Upper Allochthon microcontinent in the western part of present
54
55 3419 northern Africa is based on the absence of a *c.* 1 Ga Mesoproterozoic detrital zircon
56
57 3420 population (Fernández-Suárez et al., 2003), compared with their abundance in the central and
58
59 3421 northern Iberian Autochthon (Fernández-Suárez et al., 1999, 2000; Martínez Catalán et al.,
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61 3422 2004b; Gómez Barreiro et al., 2007) and the Central Armorican Domain (Dabard et al., 2021;

Stephan et al., 2019). The zircon age spectra of the Upper Allochthon are compatible with provenance from the West African Craton, while those of the Autochthon in Iberia suggest a more eastern paleoposition, closer to the Saharan Metacraton and the adjacent Arabian-Nubian Shield. Note the proximity at 500 Ma of the future Upper Allochthon with the Saxothuringian and Ossa-Morena Zones, where the Mesoproterozoic zircon population is also absent or very rare (Linnemann et al., 2008; Martínez Catalán et al., 2020). Moreover, faunistic similarities between the Teplá-Barrandian and Iberian Autochthon include Ordovician species found in the Ossa-Morena Zone (Gutiérrez-Marco et al., 1999b).

The microcontinent has been placed to the north of, and partly outboard from future Avalonia because its separation from Gondwana predates that of Avalonia, which occurred after 480 Ma (Pickering and Smith, 1997; Stampfli et al., 2013). Its location is conditioned by the paleoposition of Avalonia, Baltica and Gondwana in the reconstruction of GMap2015 for 500 Ma (Fig. 12). If the position of Baltica was more to the east or north, the terrane may have been located in the eastern continuation of Avalonia instead of partly outboard of it.

For the sake of simplicity, we have considered Avalonia as a unique feature instead of including Avalon-Acadia and England-Brabant separately. Moreover, we have included the Rhenohercynian and Moravo-Silesian zones in Avalonia and prolonged it to the NE to explain its probable continuation to the Balkans and Turkey. However it is possible that the latter were different entities.

At 470 Ma, separation of Avalonia had been completed with the opening of the Rheic Ocean, while the location of the future Upper Allochthon is unconstrained, but perhaps never too far from Gondwana. Figure 13 shows with a question mark the possible position of northern Avalonia at that time assuming that the reconstruction at 500 Ma is correct.

Meanwhile, the northern and NW marine platform of Gondwana continued to extend, as rift-related bimodal and alkaline magmatism took place until at least 460 Ma in the future Lower Allochthon (Santos Zalduegui et al., 1995; Díez Fernández et al., 2012b), the Parautochthon (Dias da Silva et al., 2016) and the boundary between de Ossa-Morena and Central Iberian zones (Galindo, 1989; Ochsner, 1993).

9.2.2. Early Variscan convergence

By the Lower Devonian, a wide water realm existed north of Gondwana. Domeier and Torsvik (2014) divided it in two oceanic domains, the Rheic to the west and the Paleotethys to

3455 the east, with one and three mid-ocean ridges respectively, and both separated by a subduction
1
2 3456 zone dipping to the east. In their text, the authors place the opening of the Paleotethys in the
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4 3457 Lower Devonian, but in their figures, this is an older, very large ocean at 410 Ma. In Figure
5
6 3458 12, this hypothetical large Paleotethys of Domeier and Torsvik (2014) is labeled at 400 Ma
7
8 3459 between brackets with a question mark. Actually, opening of the Paleotethys occurred at the
9
10 3460 Eifelian (390 Ma) according to Stampfli et al. (2013), which suggest that it did not exist or
11
12 3461 was rather narrow at 400 Ma. A large region in present northern Africa recorded an Upper
13
14 3462 Devonian-early Carboniferous thermal uplift and rifting (Frizon de Lamotte et al., 2013)
15
16 3463 probably representing the inception of the true Paleotethys.

17 3464 At 400 Ma, the oceanic lithosphere between Gondwana and the future Upper
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19 3465 Allochthon is that of the Mid-Variscan Ocean preserved in the Middle Allochthon. It was
20
21 3466 partly the lithosphere of the Ordovician Rheic Ocean and partly new, created by the relative
22
23 3467 movements between Gondwana and the microcontinent during the Silurian and Lower
24
25 3468 Devonian. Details are unknown, but suprasubduction signatures have been found in some
26
27 3469 oceanic rocks of these ages, suggesting that intra-oceanic subduction zones and perhaps
28
29 3470 incipient volcanic arcs developed close to the microcontinent (Díaz García et al., 1999a; Pin
30
31 3471 et al., 2002, 2006; Sánchez Martínez et al., 2007b, 2011; Arenas et al., 2014).

32 3472 The first early Variscan events were recorded in the Upper Allochthon during the
33
34 3473 Lower Devonian: high-P/high-T metamorphism reaching eclogite facies and dated at (415?)
35
36 3474 400-380 Ma is identified in the lower group of all Variscan Massifs where the Upper
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38 3475 Allochthon occurs (Schäfer et al., 1993b; Santos Zalduegui et al., 1996; Franke, 2000;
39
40 3476 Lardeaux et al., 2001; Ordóñez Casado et al., 2001; Kryza and Fanning, 2007; Godard, 2009;
41
42 3477 Koglin et al., 2018; Collett et al., 2018; Martínez Catalán et al., 2020). This event reflects
43
44 3478 subduction of the Upper Allochthon lower group under the units presently forming the upper
45
46 3479 group (Martínez Catalán et al., 2009). If this subduction zone was the same as that proposed
47
48 3480 by Domeier and Torsvik (2014), its polarity should be the opposite, (*i.e.*, to the NW; Fig. 13)
49
50 3481 because the part of the microcontinent subducted at *c.* 400 Ma was the rifted margin of the
51
52 3482 microcontinent, facing Gondwana, while the upper plate was the active Cambro-Ordovician
53
54 3483 margin, facing the Rheic Ocean (Fig. 12, 400 Ma).

54 3484 That early subduction initiated the building of an accretionary wedge by incorporation
55
56 3485 of new elements. The Mid-Variscan Ocean was closed by oceanic subduction, underthrusting
57
58 3486 and imbrication by 380 Ma. This was followed by continental subduction of the extended
59
60 3487 external part of the northern Gondwana margin. Exhumation of the previously subducted part

3488 of the microcontinent took place between 395-360 Ma, while oceanic imbricates formed the
1 3489 Middle Allochthon and high-P metamorphism affected the more external part of the northern
2
3 3490 Gondwana margin at 375-360 Ma, which later was incorporated to the upper plate as the
4
5 3491 Lower Allochthon during the collisional stage. An oblique subduction with an eastward
6
7 3492 component is suggested in Figure 12 (compare 400 and 370 Ma images) based on kinematic
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9 3493 criteria observed in the Lower Allochthon of NW Iberia (Malpica-Tui Complex; Díez
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11 3494 Fernández and Martínez Catalán, 2012). Once the subduction of the Lower Allochthon was
12
13 3495 blocked by underthrusting of thicker Gondwanan crust, the convergence regime changed from
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15 3496 subductive to collisional.

16
17 3497 Meanwhile, the western part of the Rheic Ocean recorded renewed activity. An arc
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19 3498 developed at the late Silurian-Lower Devonian (425-392 Ma; Zeh and Will, 2010) in eastern
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21 3499 Avalonia, and relics of it are preserved in the Mid German Crystalline High and the Northern
22 3500 Phyllite Zone, at the boundary between the Rhenish and Bohemian massifs. Laurussia had
23
24 3501 been amalgamated by 420 Ma (Ludlow; Stampfli et al., 2013), and a ribbon microcontinent
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26 3502 and/or island arc was pulled apart from Laurussia (Hanseatic terrane of Stampfli et al., 2013),
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28 3503 opening the Rhenohercynian Ocean. Its existence spans ~410-380 Ma for Franke (2000),
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30 3504 ~410-350 Ma for Zeh and Gerdes (2010) and ~370-320 Ma for Stampfli et al. (2013). We
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32 3505 have chosen the interval given by Zeh and Gerdes (2010) and Zeh and Will (2010) for central
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34 3506 Europe, adapting their reconstruction, shown in a series of cross sections, to ours (Figs. 12
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36 3507 and 13). The result is an unexpectedly wide Rhenohercynian Ocean between Avalonia and the
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38 3508 Silurian-Devonian arc, whose reliability depends on the reconstruction by Domeier and
39
40 3509 Torsvik (2014). But it finds support in the fact that present west-dipping subduction zones are
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42 3510 characterized by large back-arc spreading (Doglioni et al., 2007).

43 3511 The reconstruction is coherent with the evolution of SW Iberia at the contact between
44
45 3512 the OMZ and SPZ. There, an ophiolite dated at 480 Ma represents the original lithosphere of
46
47 3513 the Rheic Ocean (Fonseca et al., 1999; Pedro et al., 2010). But it is included in a high-P
48
49 3514 mélange forming part of an orogenic wedge and representing the closure of an ocean by SW-
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51 3515 directed subduction at *c.* 400 Ma (Rubio Pascual et al., 2013b). An arc and back-arc pair,
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53 3516 another high-P unit and a NE-verging accretionary wedge developed later and complex
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55 3517 continent/ocean interactions, including an extensional event, occurred before a thin-skinned
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57 3518 thrust belt developed in the South Portuguese Zone between 340-310 Ma (Simancas et al.,
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59 3519 2004). Note that because the *c.* 90° late-Variscan counterclockwise rotation of SW Iberia,
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61 3520 present SW-directed subduction and NE-verging thrusting correspond, before 340 Ma, to
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3521 NW-directed and SE-verging respectively, fitting the reconstruction of Fig. 12. The opening of
1 3522 the Rhenohercynian or Lizard-Giessen Ocean occurred in the latest Silurian-Lower Devonian
2
3 3523 (425-392 Ma; Zeh and Will, 2010; 405-390 Ma; Clark et al., 1998; Woodcock et al., 2007),
4
5 3524 causing the closure of the Rheic Ocean by the Upper Devonian-early Carboniferous. Later,
6
7 3525 the closure of the Rhenohercynian Ocean is indicated by eclogites of 350 Ma, and the
8
9 3526 collision and underthrusting of Avalonia started probably around *c.* 345 Ma in the South
10
11 3527 Portuguese Zone (Simancas, 2004), the Rhenohercynian Zone (Oncken et al., 2000; Zeh and
12
13 3528 Will, 2010), and the Moravo-Silesian Zone (Tomek et al., 2019).

14 3529 15 16 17 3530 *9.2.3. Variscan collision*

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19
20 3531 The collision of the accretionary wedge with Gondwana emplaced the Mid-Variscan
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22 3532 Allochthon and was accompanied by the incorporation of the Parautochthon and of
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24 3533 generalized compressional deformation. Recumbent folds formed in NW Iberia, and the first
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26 3534 folds in the Central Iberian Zone and the Moroccan Eastern Meseta started developing in the
27
28 3535 earliest Carboniferous. The Saxothuringian Zone was covered by the Allochthon and partly
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30 3536 subducted beneath. But there is no evidence that the same occurred to the west, in the Ossa-
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32 3537 Morena Zone or in Morocco, where compression was probably related with collision of the
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34 3538 Silurian-Devonian arc of Avalonian derivation. This is supported by zircon age spectra in
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36 3539 Upper Devonian synorogenic flysch in the southern Ossa-Morena Zone and the northern part
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38 3540 of Eastern Meseta, which points to an Avalonian source (Pereira et al., 2012; Accotto et al.,
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40 3541 2020, 2021). That source may have been a piece of the Silurian-Devonian arc obducted at *c.*
41
42 3542 360 Ma not far from the flysch occurrences (Fig. 12), while arc collision may explain the
43
44 3543 recumbent folds in the central part of the Ossa-Morena Zone and the subduction of the
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46 3544 Cubito-Moura Unit in the south and of the Central Unit in the north. Both the latter occurred
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48 3545 at the Upper Devonian (380-360 Ma) and are coeval with high-P metamorphism of the
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50 3546 eclogitic unit in the Pays de Léon and in the Mid German Crystalline High.

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52 3547 The collisions of the accretionary prism and of the Silurian-Devonian arc were
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54 3548 apparently unrelated with push by Laurussia, the large continent formed by Laurentia, Baltica
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56 3549 and Avalonia, as it was far away at that time (370-360 Ma) according to the reconstruction of
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58 3550 Domeier and Torsvik (2014) that we have used. A fact supporting this hypothesis is that no
59
60 3551 other allochthon was emplaced on top of the Upper Allochthon, which excludes obduction
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62 3552 and/or overthrusting by Laurussia.

3553 According to the reconstruction provided by Matthews et al. (2016) for GPlates,
1
2 3554 Laurussia moved to the east in relation to Gondwana some 50-55° in longitude between 360-
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4 3555 340 Ma. Then both migrated together to the NE between 340-330 Ma, which was followed at
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6 3556 330-320 Ma by a new relative eastward displacement of Laurussia of roughly 15° in
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8 3557 longitude. The first relative dextral translation explains the closure of the Rhenohercynian
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10 3558 Ocean as well as much of the activity mentioned in the previous paragraph. But it would have
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12 3559 been insufficient for provoking a collision with the Variscan Belt at *c.* 345 Ma if the latter had
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14 3560 remained attached to northern Gondwana. An additional motion of the Variscan Belt was
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16 3561 needed, which in our model is solved by a counterclockwise rotation due to the opening of a
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18 3562 small ocean to the south.

19 3563 Such ocean is a good candidate for being named the Paleotethys. Its birth coincides
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21 3564 with an Upper Devonian-early Carboniferous extensional event recorded in southern Iberia
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23 3565 and northern Africa (Simancas et al., 2003; Frizon de Lamotte et al., 2013; Rubio Pascual et
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25 3566 al., 2013b; Lahfid et al., 2019). In the model, the Paleotethys began to open at 365 Ma and
26
27 3567 reached its maximum width at 345 Ma, when the Rhenohercynian Ocean had closed (Zeh and
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29 3568 Gerdes, 2010) and a thrust belt initiated its development in the South Portuguese,
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31 3569 Rhenohercynian and Moravo-Silesian zones. No oceanic lithosphere was apparently created
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33 3570 in Morocco (Michard et al., 2008, 2010a; Simancas et al., 2009).

34 3571 The progressive closure of the Mid-Variscan Ocean, coeval with the initial rifting
35
36 3572 stage that will lead to the opening of the Paleotethys, provides an explanation for
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38 3573 simultaneous contraction and extension in the Variscan Belt (Fig. 12, 370-350 Ma interval)
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40 3574 which is an alternative to the mantle plume hypothesis (Simancas et al., 2003, 2006), although
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42 3575 both are not mutually exclusive.

43 3576 The middle Carboniferous Variscan suture in the Mauritanides (Caby and Kienast,
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45 3577 2009), might be that of the Rheic, Rhenohercynian or Paleotethys oceans, but also due to the
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47 3578 collision of a peri-Gondwanan terrane approximately located where plate reconstructions
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49 3579 place the Mesozoic Florida-Suwannee terrane. The reconstructions of Domeier and Torsvik
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51 3580 (2014) and Matthews et al. (2016) do not include such terrane, and keep Florida-Suwannee
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53 3581 fixed to Gondwana during the Paleozoic. Consequently, the same is done here, being aware
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55 3582 that pulling apart and docking of some terrane could have occurred in NW Africa and new
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57 3583 data might provide a precise correlation with European Variscan sutures.

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3585 9.2.4. Late Variscan deformation

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23586 The final emplacement of the Iberian Allochthon has been dated at 345-340 Ma,
3 including thrusting onto the Parautochthon and of the whole ensemble onto the Autochthon
43587 (Dallmeyer et al., 1997; Martínez Catalán et al., 2007, 2009). This can be related to the
5
63588 closure of the Rhehercynian Ocean and final collision with Laurussia. The continental
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83589 subduction of the Saxothuringian Zone is dated between 370-340 Ma (Franke and Stein,
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103590 2000; Žáčková et al., 2010; Konopásek et al., 2019), although the peak of high-P
113591 metamorphism occurred at 360-350 Ma (Schmädicke et al., 1995; Rötzler and Plessen, 2010;
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133592 Závada et al., 2021). Continuation of convergence until 340 Ma suggests a relationship with
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153593 Laurussian push, which may also be the cause of propagation of early Variscan folds and
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173594 thrust faults toward the foreland, and of the initial curvature of the fold axes that ultimately
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193595 led to the Central Iberian Arc and possibly the Eastern Meseta Arc (Figs. 1, 2 and 8), creating
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213596 a configuration similar to that of Figure 11f. The reconstruction at 350-330 Ma indicates that
223597 most of the northern boundary of the Mid-Variscan Allochthon came into contact with
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243598 Avalonia and supports the contention that the Brunovistulian block derives from, or formed
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263599 part of that ribbon continent.
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303601 Subsequent relative dextral displacement of Laurussia and Gondwana was responsible
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323602 for the late Variscan deformation restored in Figure 11, whose successive steps are
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343603 approximately reflected in Figure 12 between 330 and 290 Ma. They show that the suture
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363604 related with the emplacement of the Mid-Variscan Allochthon derives from an ocean situated
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383605 to the NW, but after strong dextral shearing involving rotational deformation, the root of the
393606 suture was moved to a N-S attitude in the east of the Variscan Belt.
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423607 Placing the root in the center of the Bohemian Massif makes sense, because: i) this is
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443608 the only place where part of the Moho may be that of the Allochthon (Fig. 7b); ii) no Middle
453609 or Lower Allochthon units have been described (Fig. 6). The latter is what can be expected in
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473610 the root zone of a large thrust fault, while in the periphery, detached fragments of its relative
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493611 autochthon can be commonly found as imbricates. The probable continuation of the root zone
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513612 to the south is an area where the Variscan basement is mostly covered, and probably includes
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533613 a terrane or terranes different from that of the Mid-Variscan Upper Allochthon.
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553614 In that interpretation, the Mediterranean province of the Variscan Belt does not
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573615 represent an individual terrane, but simply the eastern part of the extended northern
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593616 Gondwana platform, which recorded a somewhat different evolution (younger Ordovician
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3617 magmatism, Sardinian unconformity, along margin facies changes). The transition from the
1
2 3618 Iberian and central Europe province to the Mediterranean province occurs between the two
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4 3619 flanks of the Ibero-Armorican Arc (IAA; Fig. 1). But the reconstruction shows that it occurs
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6 3620 also between the two sides of the more or less acute angle of the tongue-shaped Mid-Variscan
7
8 3621 Allochthon (Fig. 11e). The microcontinent whose collision formed the Mid-Variscan
9
10 3622 Allochthon impinged the northern Gondwana platform at a place where some lateral
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12 3623 variations existed previously, but changes are progressive and occurred along hundreds of
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14 3624 km: the separation from the Central Iberian Zone under the Cabo Ortegal Complex in NW
15
16 3625 Iberia and the western Pyrenees was at least of 500 km, and from there to the Montagne Noire
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18 3626 another 300 km must be added.

19 3627 The Balkanides and Pontides, include significant pieces of the Variscan Belt. They
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21 3628 presently surround the Black Sea Upper Cretaceous back-arc basin, and in the reconstruction
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23 3629 of Matthews et al. (2016) for GPlates, they occur in the eastern continuation of the Variscan
24
25 3630 Mediterranean province. Okay and Topuz (2017) compare the evolution of the Balkanides
26
27 3631 and Pontides with that of central Europe during the Carboniferous, finding close similarities
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29 3632 for the metamorphic evolution and granite magmatism, and propose continuity among them at
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31 3633 least from the Upper Devonian. Conversely, the future Anatolides and Taurides, where no
32
33 3634 Variscan deformation is recorded, remained attached to Gondwana (*i.e.*, the southern margin
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35 3635 of Paleotethys) during the whole upper Paleozoic.

36 3636 In the reconstruction, the small Paleotethys closed at 320 Ma following north-directed
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38 3637 subduction. This is in agreement with the two classical provinces established in the Caucasus
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40 3638 and Turkey for the late Paleozoic (Adamia et al., 1987, 2011): the northern province
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42 3639 representing an active margin and the southern one a passive margin. A third province exist
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44 3640 farther to the north, from Moesia to the Istanbul Zone, interpreted as the southern margin of
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46 3641 Laurussia during the late Paleozoic (Okay and Topuz, 2017). This corresponds to the eastern
47
48 3642 part of the Avalonia terrane, which joined the Pontides much later, during the Upper Triassic.

49 3643 The space between the Mediterranean province and the Anatolides and Taurides was
50
51 3644 occupied, at least partially, by Gondwanan continental crust: the Alpine terranes of Figure 12,
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53 3645 which were later incorporated to the Alpine chains. These include for instance Adria,
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55 3646 Calabria, the Dinaride platform, Greece, and the Peloponnese. According to Matthews et al.
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57 3647 (2016), they stayed tied to Gondwana until the Alpine collision, while Stampfli et al. (2013)
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59 3648 include them in the continuation of the Variscan Belt to the east, separated at the Middle
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3649 Devonian by the opening of the Paleotethys and attached to the south of Laurasia during the
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2 3650 Upper Carboniferous.

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4 3651 The IAA and probably the hidden arc between Iberia and Morocco developed at the
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6 3652 late stages of collision. For the first of them, an age of 305-295 Ma has been proposed based
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8 3653 on paleomagnetic data (Weil, 2006; Weil et al., 2010), while structural data fixed the interval
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10 3654 at 310-299 Ma (Pastor-Galán et al., 2011). Although a buckling model has been proposed by
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12 3655 Weil et al. (2010) to explain the origin of the IAA, we consider more reasonable the
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14 3656 indentation hypothesis proposed by Matte and Ribeiro (1975) and Matte (1986, 1991). The
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16 3657 reason is that structures such as the very large recumbent folds and thrust sheets of the West
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18 3658 Asturian-Leonese and Cantabrian zones occur only at the arc's core, and fade away to the
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20 3659 southern limb in the WALZ and its continuation to La Demanda and Iberian Chains, and to
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22 3660 the northern limb in the Basque Massifs of western Pyrenees. The large horizontal translations
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24 3661 at the hinge zone of the arc are more compatible with underthrusting of some relatively rigid
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26 3662 indentor than with a tangential longitudinal strain at the core of a large buckle fold.

27 3663 It can be argued that the folds and most of the thrust faults predate the development of
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29 3664 the IAA, which is true. But possibly some of these structures were reactivated during the
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31 3665 formation of the arc. This may be the case of the recumbent folds and basal thrust fault of the
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33 3666 Mondoñedo Nappe (Fig. 3a), where the two largest recumbent folds share a reverse limb of
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35 3667 20-30 km (Matte, 1968; Bastida et al., 1986; Martínez Catalán et al., 2003) and Ar-Ar ages of
36
37 3668 cleavage yielded abnormally young values (300-280 Ma; Dallmeyer et al., 1997). In the CZ,
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39 3669 piggy-back and out-of-sequence thrusting remained active until 300 Ma (Pérez-Estaún et al.,
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41 3670 1988, 1991; Alonso et al., 2009; Merino-Tomé et al., 2009), so overlapping the time span of
42
43 3671 arc development. On the other side, the tangential translations at the hinge of the arc should
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45 3672 be accompanied by wrench movements at the limbs. For the northern limb, coeval (310-290
46
47 3673 Ma) dextral transurrence in the southern branch of the South Armorican Shear Zone and the
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49 3674 Pyrenees (Carron et al., 1994; Cochelin et al., 2017) may account for that displacement. The
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51 3675 counterpart in the southern limb may include the sinistral Juzbado-Penalva Shear Zone (Figs.
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53 3676 1 and 2).

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55 3677 For the hidden arc between Iberia and Morocco, indentation by the Grand Banks of
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57 3678 Newfoundland has been proposed by Simancas et al. (2005, 2009). A more detailed model by
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59 3679 Simancas et al. (2013) considers that curvatures of the Central Iberian Arc (CIA) and the
60
61 3680 hidden arc were originally the same, nucleated around a paleogeographic salient in Avalonia
62
63 3681 and amplified, cut across and superposed during the Variscan collision.

3682 In our model, both arcs are different and of different age, as explained in the section
1 devoted to retrodeformation (Fig. 11) and shown in Figure 12. The CIA was nucleated by the
2 3683 emplacement of the Mid-Variscan Allochthon and closed by dextral shearing, partly during
3 3684 emplacement and partly during late-Variscan transcurrence. The hidden arc is linked to the
4 3685 activity of a wide sinistral shear zone whose motion took place during a short interval,
5 3686 intercalated in the generally dextral transcurrent regime dominant during the late
6 3687 Carboniferous. The Grand Banks indenter moved during the late Carboniferous, although the
7 3688 existence of initial smooth salients in Avalonia are not excluded, neither possible
8 3689 reactivations of older structures producing indentors there. That motion probably overlapped
9 3690 in time with the development of the IAA, and can be related with indentors in the two major
10 3691 plates confining the Variscan Belt: a northern indenter developed in Avalonia and a southern
11 3692 one in NE Gondwana, probably a promontory formed by deformation of the future Alpine
12 3693 terranes. That motion, coeval with continuity of dextral shearing, remained active during the
13 3694 early Permian.
14 3695

15 3696 Pereira et al. (2015a, 2017b) interpreted late Variscan (315-280 Ma) calc-alkaline
16 3697 granitoids of Iberia as related to subduction of the Paleotethys linked to the development of
17 3698 the IAA, and unrelated to the Variscan collision. However, the same granitoids have been
18 3699 interpreted as resulting from partial melting of the lower crust with variable mantelic heat and
19 3700 magmatic input and also variable contribution of orogenic crustal thickening (Castiñeiras et
20 3701 al., 2008; Carracedo et al., 2009; Orejana et al., 2012; Villaseca et al., 2017). Even assuming a
21 3702 crustal origin with variable mantle contribution, Gutiérrez-Alonso et al. (2011) defended
22 3703 delamination of thickened mantle lithosphere during the closure of the IAA orocline as the
23 3704 source of heat necessary for massive granite production. And Alcock et al. (2015) calculate
24 3705 that heat production due to crustal thickening could had contributed significantly to heat
25 3706 input for NW Iberian granite production, without discarding an additional source due to
26 3707 asthenospheric upwelling.
27

28 3708 Our model does not back the Paleotethys oceanic subduction hypothesis of Pereira et
29 3709 al. (2015a, 2017b) for late Variscan granitoids, because south of the Iberian Peninsula, the
30 3710 small Paleotethys closed at 320 Ma, and in Matthews et al. (2016) reconstruction, no oceanic
31 3711 crust existed south of Iberia between 325-280 Ma. This does not mean that the model by
32 3712 Pereira et al. (2015a, 2017b) should be discarded, because the reconstruction may be
33 3713 incorrect. Meanwhile, we find that our model of the Variscan Belt implies a late Variscan
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3714 narrowing and associated crustal thickening which rather supports the alternative of crustal
3715 thickening and melting with variable amounts of mantle involvement.

3716 Figure 13 shows the late Variscan evolution of the belt in relation with the relative
3717 position of Laurussia and Gondwana. Successive positions of these plates are based on the
3718 model by Domeier and Torsvik (2014) and the reconstruction of Matthews et al. (2016). The
3719 Variscan Belt became progressively squeezed between the major plates as shown by the angle
3720 formed by two reference lines, which decreases between 340 and 290 Ma. This imposed an
3721 internal deformation similar to trishear (Erslev, 1991), a kinematic model of fault-propagation
3722 folding where the decrease in displacement along a fault induces heterogeneous shear in a
3723 triangular zone radiating from the tip line (Hardy and Allmendinger, 2011). The trishear
3724 model was initially applied to the triangular area in front of a propagating thrust fault (Erslev,
3725 1991; Zehnder and Allmendinger, 2000; Cristallini and Allmendinger, 2011) and was
3726 extended to normal faults and 3D models by Cardozo (2008).

3727 The late Variscan evolution of the belt in relation with the relative position of the
3728 major, supposedly rigid plates, was conditioned by the eastward and southward translation
3729 and clockwise rotation of Laurussia relative to Gondwana and by internal heterogeneities
3730 inside the mobile zone. The eastward relative translation of Laurussia between 340 and 320
3731 Ma, calculated for a longitudinal displacement of *c.* 15° (continuous and dashed red lines at
3732 340 Ma), is around 1660 km at the Equator and 1440 km at 30° south latitude. This explains
3733 the strong dextral component of late Variscan deformation. Relative northward translation of
3734 Gondwana and *c.* 12° clockwise rotation are responsible for the closure of the small
3735 Paleotethys Ocean, indentation leading to orocline development (Ibero-Armorican Arc and a
3736 hidden arc between Iberia and Morocco), and to late, NW-SE dextral oblique faulting. In
3737 addition, early to middle Permian conjugate strike-slip brittle faulting was put in relation with
3738 such a megashear and the NW-SE oblique faulting by Arthaud and Matte (1977).

3740 **10. Conclusions**

3741 The non-cylindrical geometry of the Variscan Belt derives from the lack of continuity
3742 of one of its zones, the Mid-Variscan Allochthon, while other zones essentially maintain their
3743 continuity along the overall belt from the Bohemian Massif to Morocco. The geometry of the
3744 Mid-Variscan Allochthon, together with that of the whole ensemble, is partially restored for
3745 the late Variscan deformation in a series of steps that yield a picture of the geometry at the

3746 end of the early Variscan stage. Based on that picture, the Mid-Variscan Allochthon is
1 3747 explained by overthrusting of a microcontinent, a peri-Gondwanan terrane detached from a
2 3748 region close to NW Africa and collided with more eastern parts of northern Gondwana
3 3749 platform.
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7 3750 What is preserved of the microcontinent presently forms the Upper Allochthon, but its
8 3751 collision involved the closure of an intervening Mid-Variscan Ocean whose lithosphere partly
9 3752 belongs to the initial (Cambro-Ordovician) Rheic Ocean and partly was created later, during
10 3753 the Silurian and Devonian. Some of these oceanic lithospheres were incorporated to an
11 3754 accretionary prism during closure of the Mid-Variscan Ocean, and were later overthrust onto
12 3755 the northern Gondwana platform, presently forming the Middle Allochthon. Outer parts of
13 3756 that continental platform were also incorporated to the accretionary prism and later emplaced
14 3757 to more internal parts of it, forming the Lower Allochthon and the Parautochthon.
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22 3758 The available knowledge of pre-Variscan and early Variscan evolution and the results
23 3759 of late Variscan retrodeformation have been integrated in a calibrated plate-tectonics
24 3760 reconstruction covering the late Cambrian to early Permian, which is shown with a focus on
25 3761 the Variscan deformation. The resulting model explains the non-cylindrical geometry by: i)
26 3762 the small size of the Upper Allochthon microcontinent compared with the extent of the
27 3763 deformed northern Gondwana and Avalonia plates and ii) large intracontinental dextral
28 3764 shearing caused by the relative eastward displacement of Laurussia relative to Gondwana. The
29 3765 latter also explains the rotation of the Mid-Variscan Allochthon root zone from an original
30 3766 position to the NW of the belt to its proposed present location in the central and southern parts
31 3767 of the Bohemian Massif.
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41 3768 Southward translation and moderate clockwise rotation of Laurussia relative to
42 3769 Gondwana explain indentation leading to formation of the Ibero-Armorican Arc and a hidden
43 3770 arc between Iberia and Morocco. The open Massif Central Arc may be a side effect of the
44 3771 Ibero-Armorican Arc. However, the Central Iberian Arc is primarily related to the indentation
45 3772 of the Mid-Variscan Allochthon, although it was strongly amplified by late-Variscan dextral
46 3773 shearing. The Bohemian Arc is also related to dextral shearing, resolved there in the Elbe
47 3774 Shear Zone and the Moldanubian Thrust.
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16
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Figure captions

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 51 6710 **Fig. 1.** Sketch of the Variscan belt in an early Permian reconstruction that postdates ductile
 52
 53 6711 and brittle transcurrence. **Zones (Z):** CZ, Cantabrian; CIZ, Central Iberian (OVD, Obejo-
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 55 6712 Valsequillo Domain); GTMZ, Galicia-Trás-os-Montes; MSZ, Moravo-Silesian; MZ,
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 57 6713 Moldanubian; OMZ, Ossa-Morena; RHZ, Rhenohercynian; SPZ, South Portuguese; STZ,
 58
 59 6714 Saxo-Thuringian; TBZ, Teplá-Barrandian; WALZ, West Asturian-Leonese. Armorican
 60 6715 Domains (AD): CAD, Central; NAD, North; SAD, South; MM: Moroccan Meseta (AA, Anti-

6716 Atlas; CB, Coastal Block; EM, Eastern Meseta; SB, Sehoul Block; WM, Western Meseta).
 1
 26717 **Shear zones (SZ), faults (F) and fault zones (FZ):** BBSZ, Baden-Baden; BCSZ, Badajoz-
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 46718 Córdoba; BrF: Bray; EFZ, Elbe; JPSZ, Juzbado-Penalva; LLSZ, Lalaye-Lubine; MMFZ,
 5
 66719 Middle Meseta; MT, Moldanubian Thrust; NASZ, North Armorican; NEF, Nort-sur-Erdre;
 7
 86720 NPF, North Pyrenean; PASZ, Posada-Asinara; PTSZ, Porto-Tomar; SAF, South Atlas; SASZ,
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 106721 South Armorican (N and S, northern and southern branches); SHF, Sillon Houiller; SIF, South
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 126722 Iberia; SISZ, Southern Iberian; SMF, South Meseta; TTZ, Teisseyre-Tornquist Zone; VF,
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 146723 Variscan Front; WMSZ, Western Meseta. **Arcs (A):** BA, Bohemian; CIA, Central Iberian;
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 166724 EMA, Eastern Meseta; IAA, Ibero-Armorican; MCA, Massif Central. **Other:** BAU, Beja-
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 186725 Acebuches Unit; BM, Basque Massifs; Bu, Buçaco; Co, Corsica; Cr, Crozon. FMC, French
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 206726 Massif Central; LC, Lizard Complex; MTM, Maures-Tanneron Massif; MGCH, Mid-German
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 226727 Crystalline High; MN, Montagne Noire; Sa, Sardinia; SWM, Schwarzwald Massif; VM,
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 246728 Vosges Massif.

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 266729
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 286730 **Fig. 2. (a)** Geological map of the Iberian Peninsula showing the Variscan massifs and the
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 306731 zones of the Iberian Massif, continued into the Pyrenees, Iberian Chain and Catalonia Coast
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 326732 Ranges. Based on the Geological map of Spain and Portugal to scale 1:1000 000 (IGME-
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 346733 LNEG, 2015) and in Martínez Catalán (2012). The most important structures are shown,
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 366734 including early and late folds, thrust faults, and the main ductile strike-slip shear zones. **(b)**
 37
 386735 Detail of the geology of the Galicia allochthonous complexes showing the distribution of the
 39
 406736 three Allochthons and the Parautochthon. **(c)** Detail of the geology of the Portuguese
 41
 426737 allochthonous complexes. **Zones (Z):** CZ, Cantabrian; CIZ, Central Iberian; GTMZ, Galicia-
 43
 446738 Trás-os-Montes; OMZ, Ossa-Morena; SPZ, South Portuguese; WALZ, West Asturian-
 45
 466739 Leonese. **Domains of the CIZ (D):** DRF, of Recumbent Folds; DUF, of Upright Folds; OSD,
 47
 486740 Ollo de Sapo; OVD, Obejo-Valsequillo; SGD, Schist-Greywacke. **Allochthonous**
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 506741 **complexes:** B, Bragança; CO, Cabo Ortegal; M, Morais; MT, Malpica-Tui; O, Órdenes.
 51
 526742 **Other units:** BAU, Beja-Acebuches; CU, Central Unit; EAMB, Évora-Aracena Metamorphic
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 546743 Belt; NSSB, Northern Sierra de Sevilla Batholith; PL, Pulo do Lobo Unit. **Shear zones (SZ)**
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 566744 **and faults (F):** BCSZ, Badajoz-Córdoba; JPSZ, Juzbado-Penalva; NPF, North Pyrenean;
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 586745 PTSZ, Porto-Tomar; SISZ, Southern Iberian. **Arcs (A):** CIA, Central Iberian; IAA, Ibero-
 59
 606746 Armorican. **Pyrenees:** AZ, Axial Zone; NM, Northern Massifs; Basque Massifs: CV, Cinco
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 626747 Villas; QR, Quinto Real; UB, Ursuya-Baygura. **Balearic Islands:** Ma, Majorca; Me,

Menorca; Ib, Ibiza. **Allochthonous groups:** LA, Lower Allochthon; MA, Middle Allochthon; PA, Parautochthon; UA, Upper Allochthon (IP, Intermediate-P; HP-HT, High-P/High-T).

Fig. 3. Cross sections representative of the structure of the Iberian Massif including the Variscan sutures. **(a)** Northwest Iberian section A-A', modified from Martínez Catalán et al. (2014). **(b)** Central Iberian section B-B', based on Martínez Poyatos et al. (2012). **(c)** Southwest Iberian section C-C', based on Simancas et al. (2003) and partly on Rubio Pascual et al. (2013b). The sections are built from field geology and partly using deep seismic reflection profiles. For section A-A', these are ESCIN-1 (Pérez-Estaún et al., 1994, 1995), ESCIN-3.2 (Álvarez-Marrón, 1995, 1996) and ESCIN-3.3 (Martínez Catalán et al., 1995; Ayarza et al., 1998). The two latter were acquired offshore, while section A-A' derives from onshore geology. In the parts of section A-A' roughly corresponding to these two profiles, (GTMZ, CIZ and WALZ), the seismic reflections have been placed approximately. Section B-B' roughly follows the seismic line ALCUDIA (Martínez Poyatos et al., 2012). Section C-C' follows the seismic line IBERSEIS (Simancas et al., 2003). Whrench faults are continued at depth avoiding to cross low-dipping reflections. Colors corresponding to zones or groups are superposed to frames or grey shades indicating lithologies or units. **Zones (Z):** CZ, Cantabrian; CIZ, Central Iberian; GTMZ, Galicia-Trás-os-Montes; OMZ, Ossa-Morena; SPZ, South Portuguese; WALZ, West Asturian-Leonese. **Domains of the CIZ (D):** DRF, of Recumbent Folds; DUF, of Upright Folds; OSD, Ollo de Sapo; OVD, Obejo-Valsequillo; SGD, Schist-Greywacke. **Other units:** BAU, Beja-Acebuches; EAMB, Évora-Aracena Metamorphic Belt; IRB, Iberian Reflective Body. **Other:** Upper Allochthon: IP, Intermediate-P; HP-HT, High-P/High-T. See Figure 2 for location of the sections.

Fig. 4. Geological map of the French and related Variscan Massifs. Based on the Geological map of France to scale 1:1 000 000 (BRGM, 1996), and in Faure et al. (2009), Ballèvre et al. (2014), and Lardeaux et al. (2014). **Zones (Z):** MZ, Moldanubian; RHZ, Rhenohercynian; STZ, Saxothuringian; WALZ, West Asturian-Leonese; **Armorican Domains (AD):** CAD, Central; NAD, North; SAD, South. **Shear zones (SZ) and faults (F):** BBSZ, Baden-Baden; BrF: Bray; GCF, Granville-Cancalle; LLSZ, Lalaye-Lubine; NASZ, North Armorican; NEF, Nort-sur-Erdre; NPF, North Pyrenean; SASZ, South Armorican (N and S, northern and southern branches); SHF, Sillon Houiller. **Arcs (A):** MCA, Massif Central. **Pyrenees:** AZ,

6780 Axial zone; NM, Northern massifs; Basque Massifs: CV, Cinco Villas; QR, Quinto Real; UB,
 1
 2 6781 Ursuya-Baygura. **Other:** Au, Audierne Bay Complex; BC, Bois-de-Cené Unit; Bv, Brévenne
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 4 6782 Unit; Cé, Cévennes; Ch, Champtoceaux Complex; ChO, Chamrousse Ophiolite; ChS,
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 6 6783 Chantonay Syncline; Co, Corsica; EmA, External Massifs of the Alps (AaM, Aar; ArM,
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 8 6784 Argentera; BeM, Belledonne; MBM, Mont Blanc); Es, Essarts Unit; GN, Giessen Nappe; IG,
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 10 6785 Ile-de-Groix; LC, Lizard Complex; LiO, Limousin ophiolite; Ln, Léon; Ly, Lyonnais; MTM,
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 12 6786 Maures-Tanneron Massif; MGCH, Mid-German Crystalline High; MN, Montagne Noire; Mo,
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 14 6787 Morvan Arc; Ms, Mauges Unit; Nj, Najac Massif; Od, Odenwald Massif; Ru, Ruhla Massif;
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 16 6788 SG, St. Georges-sur-Loire Unit; Sp, Spessart Massif; SWM, Schwarzwald Massif; TA, Tuffs
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 18 6789 Anthracifères; TP, Thiviers-Payzac Unit; VD, Velay Dome; VM, Vosges Massif.

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 21 6791 **Fig. 5.** Cross sections representative of the structure of the French Variscan massifs. **(a)**
 22
 23 6792 Armorican section A-A'', including the Mid-Variscan suture in the southern part. **(b)** Western
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 25 6793 Massif Central section B-B'. **(c)** Eastern Massif Central section C-C''. The sections are built
 26
 27 6794 from field geology, based on the Geological map of France to scale 1:1 000 000 (BRGM,
 28
 29 6795 1996), and on geological sections by Ledru et al. (1994b), Faure et al. (2009), Ballèvre et al.
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 31 6796 (2014), and Lardeaux et al. (2014). Section A-A'' follows the trace of seismic reflection
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 33 6797 profiles ARMOR 2 South and 2 North (Bitri et al., 2003, 2010), AR2, AR1 and SWAT 10
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 35 6798 (Bitri et al., 2001), whose interpretations have been included. For the Central Armorican
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 37 6800 Domain, reflections from the Brittany Wide Angle Refection Profile (BWARF; Matte and
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 39 6801 Hirn, 1988) separated c. 100 km to the east, have been projected into the section. Whrench
 40
 41 6802 faults are continued at depth avoiding to cross low-dipping reflections. Colors corresponding
 42
 43 6803 to zones or groups are superposed to frames or grey shades indicating lithologies or units.
 44
 45 6804 **Armorican Domains (AD):** CAD, Central; NAD, North; SAD, South. **Shear zones (SZ)** and
 46
 47 6805 **faults (F):** CF, Combrailles; GCF, Granville-Cancalle; NASZ, North Armorican; NEF, Nort-
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 49 6806 sur-Erdre; OF, Ouzilly; SASZ, South Armorican (N and S, northern and southern branches);
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 51 6807 SLF, South Limousin. **Other:** FMC, French Massif Central; MN, Montagne Noire. Upper
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 53 6808 Allochthon: IP, Intermediate-P; HP-HT, High-P/High-T. See Figure 4 for location of the
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 55 6809 sections.

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 57 6810 **Fig. 6.** Geological map of the Bohemian Massif. Based on the Geological maps of Germany
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 59 6811 to scale 1:1 000 000 (BGR, 1993), and the Czech Republic to scale 1:500 000 (Czech
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Geological Survey, 2007), and in Mazur and Aleksandrowski (2001), Franke and
 Żelaźniewicz (2000, 2002), Schulmann et al. (2009, 2014), Zeh and Will (2010), Mazur et al.
 (2015) and Martínez Catalán et al. (2019). **Zones (Z):** MZ, Moldanubian; MSZ, Moravo-
 Silesian; NPhZ, Northern Phyllite; RHZ, Rhenohercynian; SPhZ, Southern Phyllite; STZ,
 Saxothuringian; TBZ, Teplá-Barrandian. **Shear zones (SZ), faults (F) and fault zones (FZ):**
 DF, Danube; EFZ, Elbe; FF, Franconian; ISF, Intra-Sudetic; MT, Moldanubian Thrust; OFZ,
 Odra; PF, Pfahl; SBF, Sudetic Boundary. **Arcs (A):** BA, Bohemian. **Massifs and units:** BL,
 Blanský les; Dr, Drosendorf; EB, Erzgebirge; EV, Erbenhof-Vohenstrauß; F, Frankenberg;
 Gf, Gföhl; GS, Góry Sowie; HM, Harz Mountains; Ka, Kaczawa Complex; KH, Kutna Hora;
 Kr, Krkonoše; LB, Lausitz Block; Mü, Münchberg; ML, Mariánské Lázně; NMe, Nové
 Město; OS, Orlica-Šniežnik Dome; PS, Praha Syncline; SG, Saxonian Granulitebirge; SK,
 South Krkonoše; SM, Staré Město; W, Wildenfels. **Other:** CBPC, Central Bohemian Plutonic
 Complex; ISB, Intra-Sudetic Basin; MGCH, Mid-German Crystalline High; Ru, Ruhla; SBB,
 South Bohemian Batholith; Sv, Svatka window; Th, Thaya window.

Fig. 7. Representative sections across the Bohemian Massif showing the crustal structure. **(a)**
 Northern section A-A", based on Krawczyk et al. (2000), Schulmann and Gayer (2000),
 Chopin et al. (2012) and Tomek et al. (2019). **(b)** Southern section B-B", based on Franke
 (2000), Schulmann et al. (2005) and Guy et al. (2011). The sections are partly based on deep
 seismic reflection profiles GRANU 9501 and 9502 (DEKORP, 1999; Krawczyk et al., 2000),
 and 9HR (Chlupáčová and Švancara, 1994). The seismic profiles are more or less close to, but
 do not coincide with the geological sections, and the reflections have been projected
 approximately on them. Colors corresponding to zones or groups are superposed to frames or
 grey shades indicating lithologies or units. **Zones (Z):** MZ, Moldanubian; MSZ, Moravo-
 Silesian; NPhZ, Northern Phyllite; RHZ, Rhenohercynian; STZ, Saxothuringian; TBZ, Teplá-
 Barrandian. See Figure 6 for location of the sections.

Fig. 8. Geological map of Variscan Morocco. Based on the Geological map of Europe and
 Adjacent Areas to scale 1:5 000 000 (BGR, 2005), Michard et al. (2010a) and Ouanaimi et al.
 (2016). **Zones (Z):** AA, Anti-Atlas; CB, Coastal Block; EM, Eastern Meseta; SB, Schoul
 Block; WM, Western Meseta. **Shear zones (SZ), faults (F) and fault zones (FZ):** MMFZ,
 Middle Meseta; RTFZ, Rabat-Tiflet; SAF, South Atlas; SMF, South Meseta; WMSZ,

6844 Western Meseta. **Arcs (A):** EMA, Eastern Meseta. **Other:** NZ, Nappe Zone; SMFr, South
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2 6845 Meseta Front.

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4 6846
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6 6847 **Fig. 9.** Correlation of zones, groups and units and of early Variscan and Variscan orogenic
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8 6848 magmatism along the Variscan Belt. Only the main massifs are included, while small ones
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10 6849 can generally be assimilated to nearby large ones. Colors and abbreviations are the same used
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12 6850 in maps and sections.

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14 6851
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16 6852 **Fig. 10.** Timing of early Variscan and Variscan events of the large massifs. The time interval
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18 6853 spans 425-275 Ma. Small massifs are assumed to have similar evolutions as the massifs to
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20 6854 which they are related in the text, and many of them are included in the figure. Bar colors and
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22 6855 abbreviations are the same used in the maps, sections and text. The events related to
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24 6856 subduction of the Rheic and Rhenohercynian oceans are given separated from those
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26 6857 considered related with the building of the Mid-Variscan accretionary complex forming the
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28 6858 Allochthon, their emplacement on the northern Gondwana Autochthon and the subsequent
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30 6859 evolution of the ensemble.

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32 6860
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34 6861 **Fig. 11.** Partial retrodeformation of brittle and ductile late orogenic deformation of the
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36 6862 Variscan Belt. The start point is the late Carboniferous reconstruction shown in Figure 1.
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38 6863 Black continuous lines represent faults not restored yet, while dashed lines represent faults
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40 6864 whose translation was removed in the last step. **(a)** After removing displacement of the main
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42 6865 late Variscan NW-SE faults (SIZ, NPF, BrF and EFZ). **(b)** The IAA has been partially
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44 6866 unfolded. **(c)** Displacement of a large dextral shear zone equivalent to the PTSZ, NEF, LLSZ
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46 6867 and BBSZ has been partially removed, and the IAA is shown at an incipient stage, as well as
47
48 6868 the BA. **(d)** Sinistral motion of the BCSZ and SISZ has been removed. **(e)** Early dextral
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50 6869 motion along the NASZ and SASZ has been removed. **(f)** Internal deformation associated to
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52 6870 ductile dextral megashear has been restored in the Allochthon, the CAD, EM, WM and the
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54 6871 Mediterranean province, while the little deformed CB and NAD are tentatively placed to the
55
56 6872 south. Ductile restoration is indicated by homogeneous, sinistral simple shear with a shear
57
58 6873 angle of 56.3° , equivalent to a shear value $\gamma = 1.5$. **Zones (Z):** CZ, Cantabrian; CIZ, Central
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60 6874 Iberian (OVD, Obejo-Valsequillo Domain); GTMZ, Galicia-Trás-os-Montes; MSZ, Moravo-
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62 6875 Silesian; MZ, Moldanubian; OMZ, Ossa-Morena; RHZ, Rhenohercynian; SPZ, South

6876 Portuguese; STZ, Saxothuringian; TBZ, Teplá-Barrandian; WALZ, West Asturian-Leonese.
 1
 2 6877 Armorican Domains (AD): CAD, Central; NAD, North; SAD, South; MM: Moroccan Meseta
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 4 6878 (AA, Anti-Atlas; CB, Coastal Block; EM, Eastern Meseta; WM, Western Meseta). **Shear**
 5 6879 **zones (SZ), faults (F) and fault zones (FZ):** BBSZ, Baden-Baden; BCSZ, Badajoz-Córdoba;
 6
 7 6880 BrF: Bray; EFZ, Elbe; LLSZ, Lalaye-Lubine; NEF, Nort-sur-Erdre; NPF, North Pyrenean;
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 9 6881 PTSZ, Porto-Tomar; SIF, South Iberia; SISZ, Southern Iberian; SMF, South Meseta. **Arcs**
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 11 6882 **(A):** BA, Bohemian; CIA, Central Iberian; EMA, Eastern Meseta; IAA, Ibero-Armorican;
 12
 13 6883 MCA, Massif Central. **Other:** Bu, Buçaco; Cr, Crozon.

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 15 6884
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 17 6885 **Fig. 12.** Evolution of the Variscan Belt in the context of Paleozoic plate dynamics. The main
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 19 6886 continental masses and the elements forming the Variscan Belt are shown in orthographic
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 21 6887 projection for different ages. Northern and eastern territories (Siberia, China, Indonesia) are
 22
 23 6888 omitted. For the late Cambrian and Lower Ordovician, a projection centered in the South Pole
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 25 6889 has been used. The geographic grid is drawn every 15°, with a thicker line for the Equator,
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 27 6890 and is fixed for the 500-470 and 400-290 Ma intervals. This figure was built using animations
 28
 29 6891 500.A97 and 470.A97 of GMap2015, the reconstruction geometries of Matthews et al. (2016)
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 31 6892 for GPlates, and the partial retrodeformation carried out in this work (Fig. 11). Blue lines
 32
 33 6893 indicate the location of sections in Figure 13 (dotted parts not shown). Note that the Mid-
 34
 35 6894 Variscan Allochthon represents the future Upper Allochthon until 400 Ma. Between 390 and
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 37 6895 360 Ma it progressively incorporated the Middle and Lower Allochthons and by 340 Ma also
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 39 6896 the Parautochthon.

40 6897
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 42 6898 **Fig. 13.** Proposed plate dynamics of the Variscan Belt depicted using cross-sections, the
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 44 6899 location of which is indicated by continuous blue lines in Figure 12. The sections vary for
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 46 6900 each time frame. They are located to cut across the Mid-Variscan Allochthon except at 370
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 48 6901 Ma, designed to highlight the role of the Silurian-Devonian arc. The late Variscan evolution
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 50 6902 related with transcurrent dextral shearing is not shown. In this figure, the three allochthons
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 52 6903 and the Parautochthon are depicted with the colors used in previous figures (maps, sections
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 54 6904 and correlations) to shown their progressive incorporation to the accretionary wedge and its
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 56 6905 thrusting onto the Autochthon during Variscan collision.

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Fig. 14. Late evolution of the Variscan Belt shown in three steps in relation with the relative position of Laurussia and Gondwana using the reconstruction of Matthews et al. (2016). The geographic grid is drawn every 15°, with a thicker line for the Equator. The Variscan Belt became progressively squeezed between the two mayor plates, as shown by the angle formed by two reference red lines. This angle decreases between 340 and 290 Ma imposing an internal deformation in the triangular area in between, comparable to the trishear kinematic model (Erslev, 1991). The dashed line at 340 Ma approximately represents the position of the parallel continuous line 20 Ma later if only translation had occurred. The curved dashed line at 290 Ma represents a realistic trace of the continuous straight line drawn in Laurussia at 340 Ma. Legend as in Figure 12.

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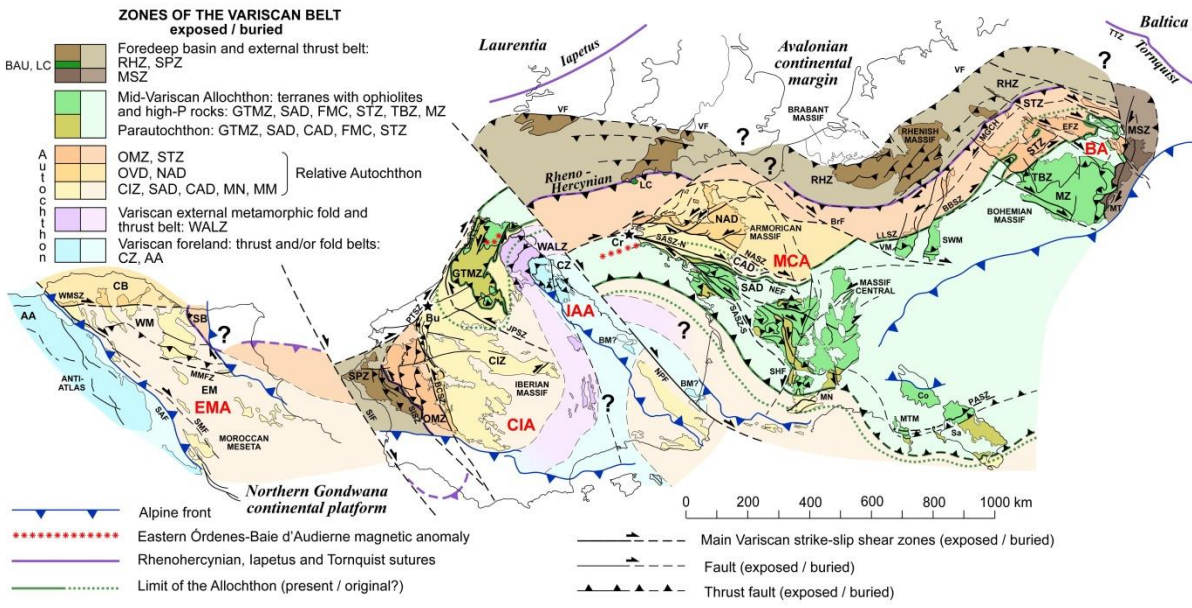


Fig. 1

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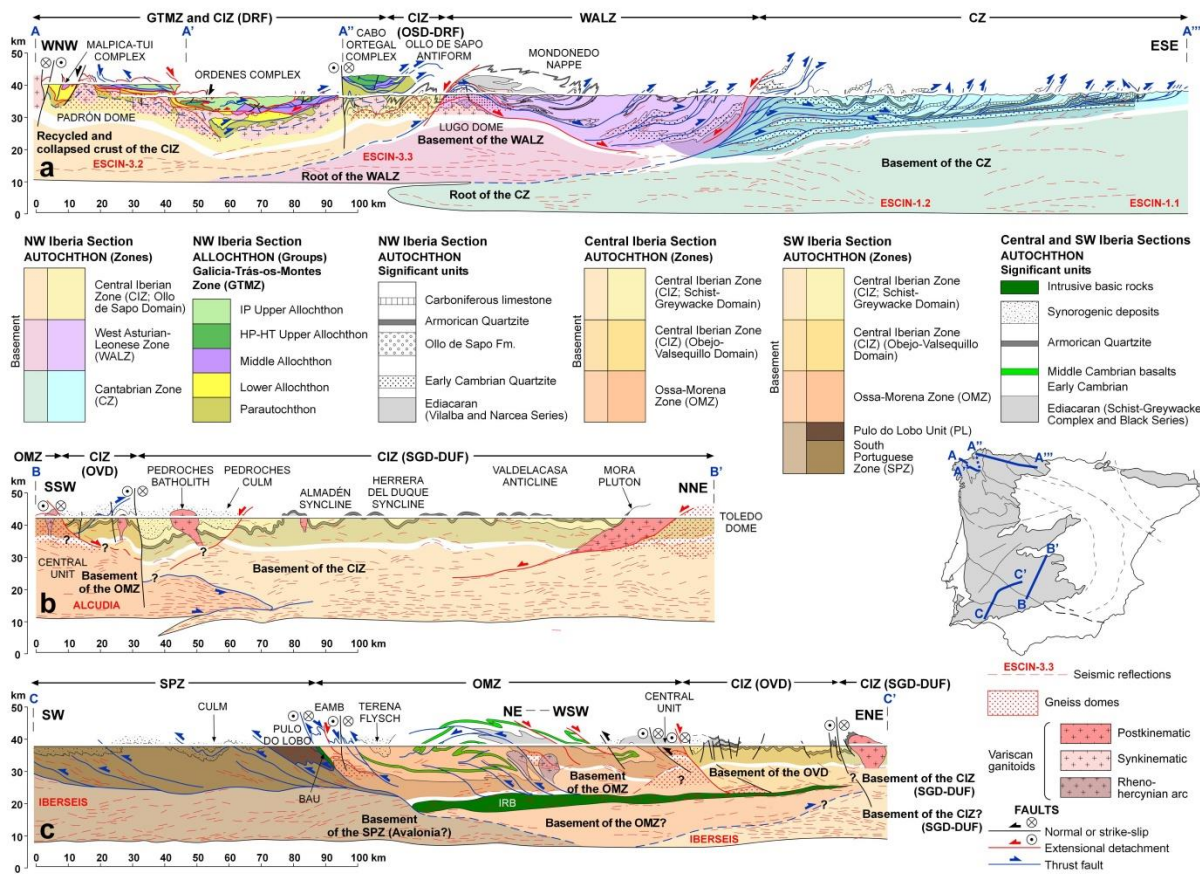
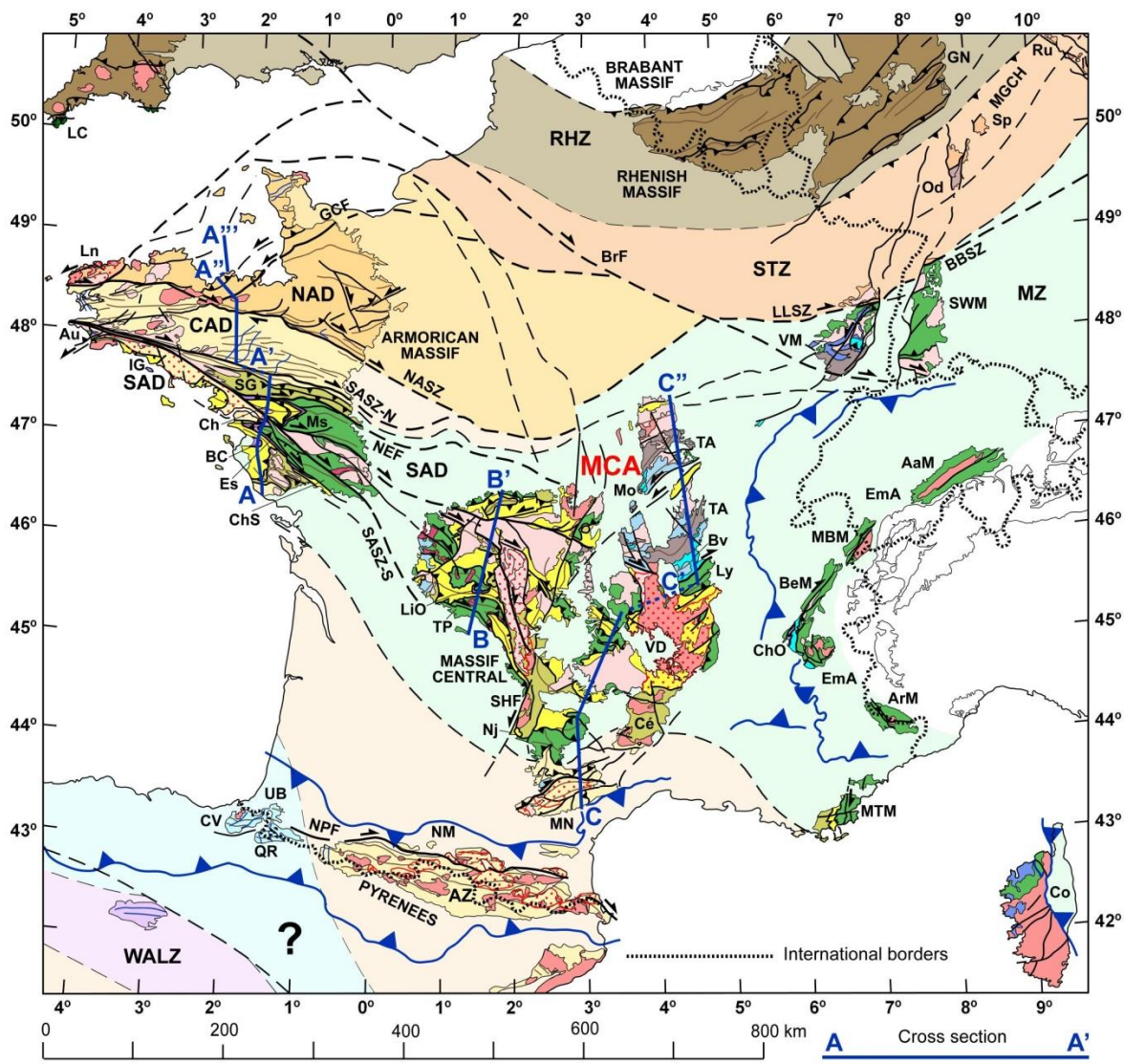


Fig. 3

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Early folds
 Thrust fault
 Main strike-slip shear zones
 Late folds
 Other faults
 Alpine fronts

ZONES OF THE FRENCH AND RELATED MASSIFS (exposed / buried)

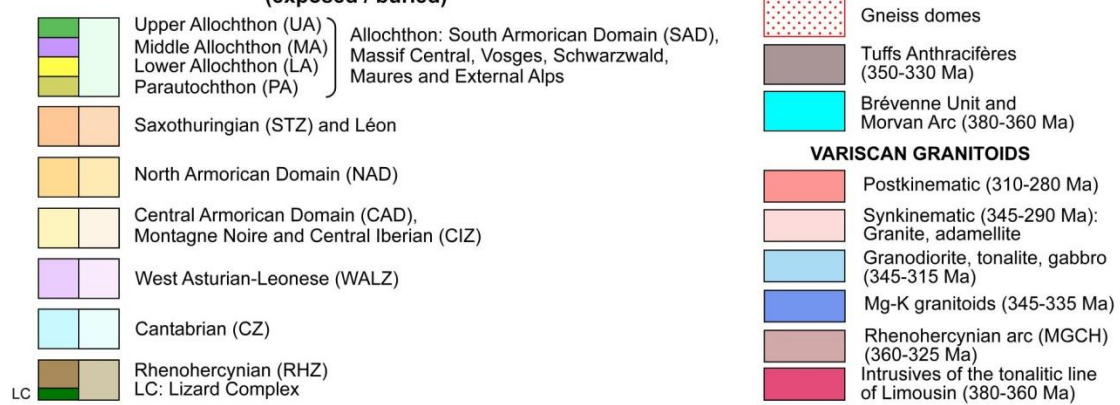


Fig. 4

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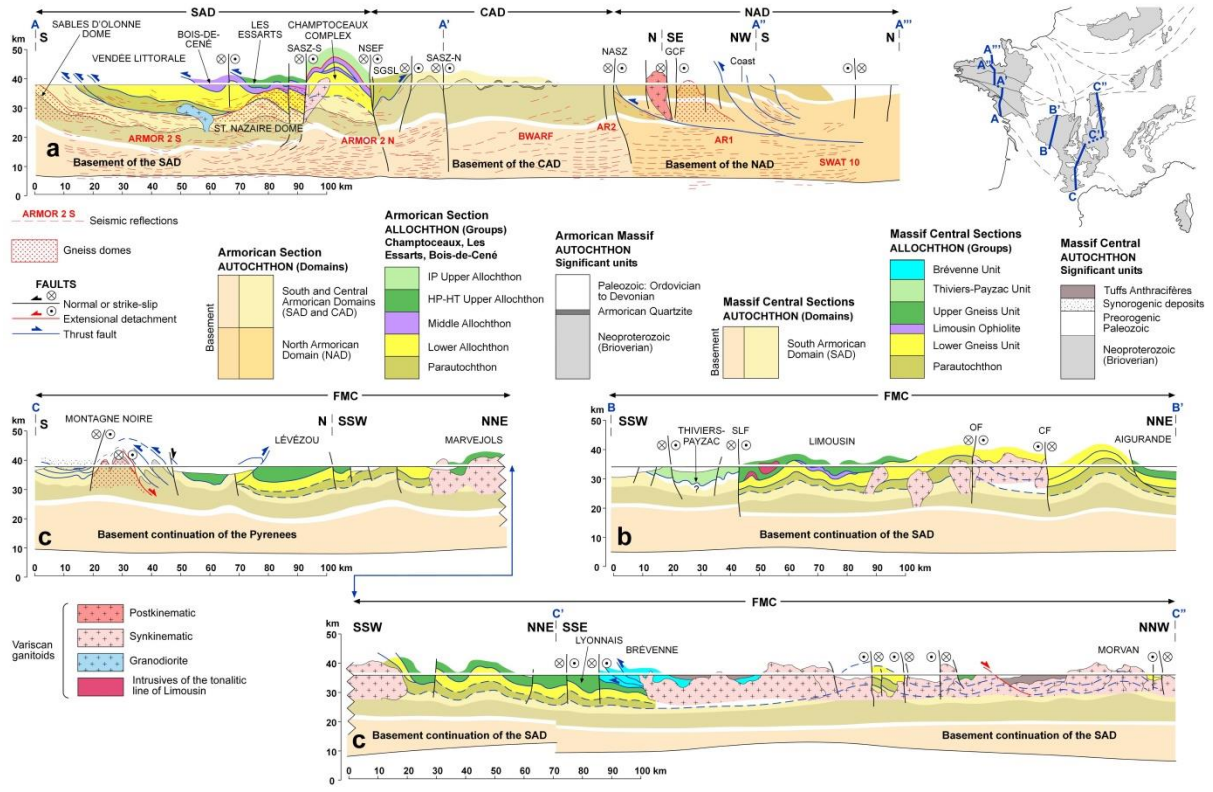


Fig. 5

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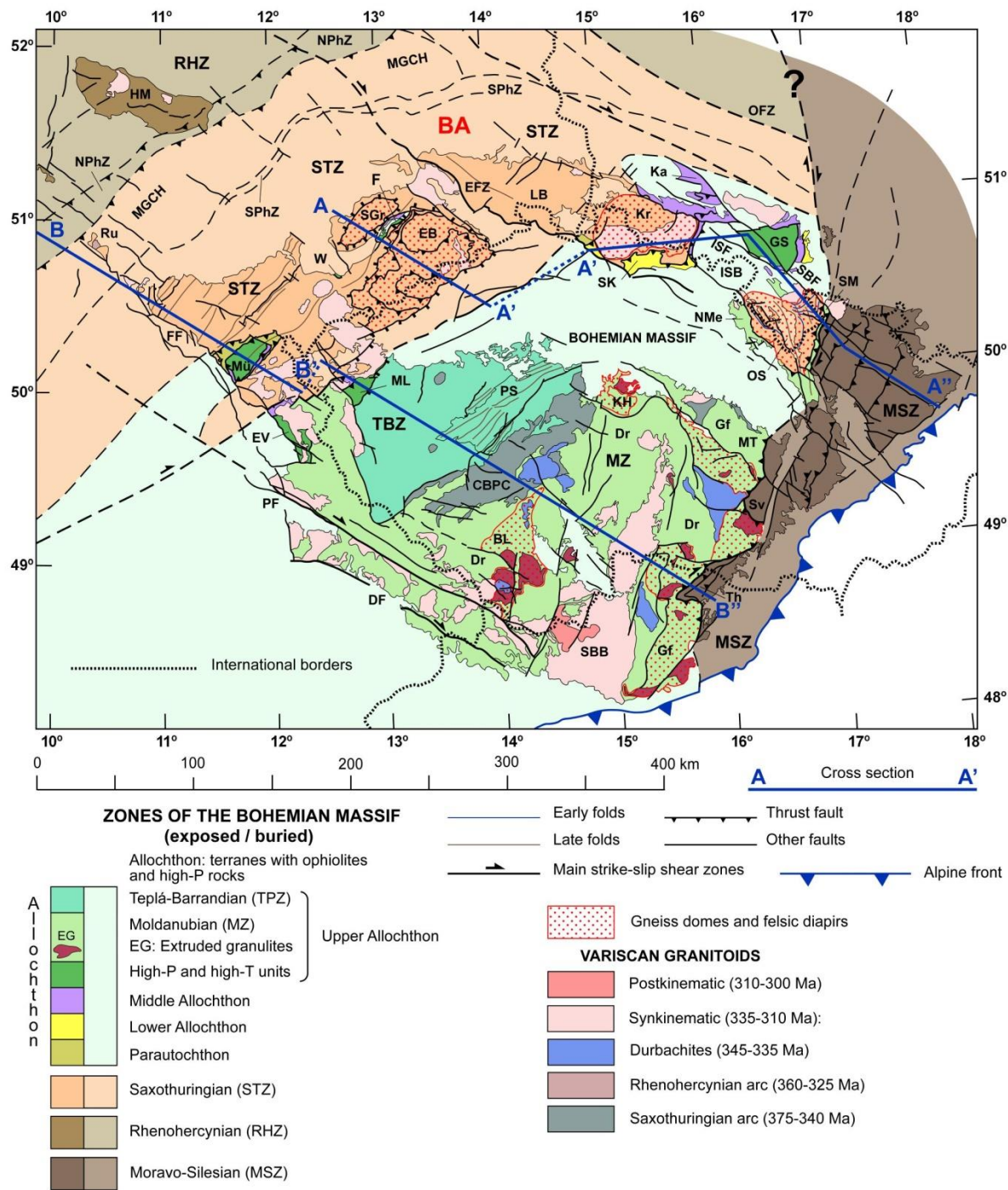


Fig. 6

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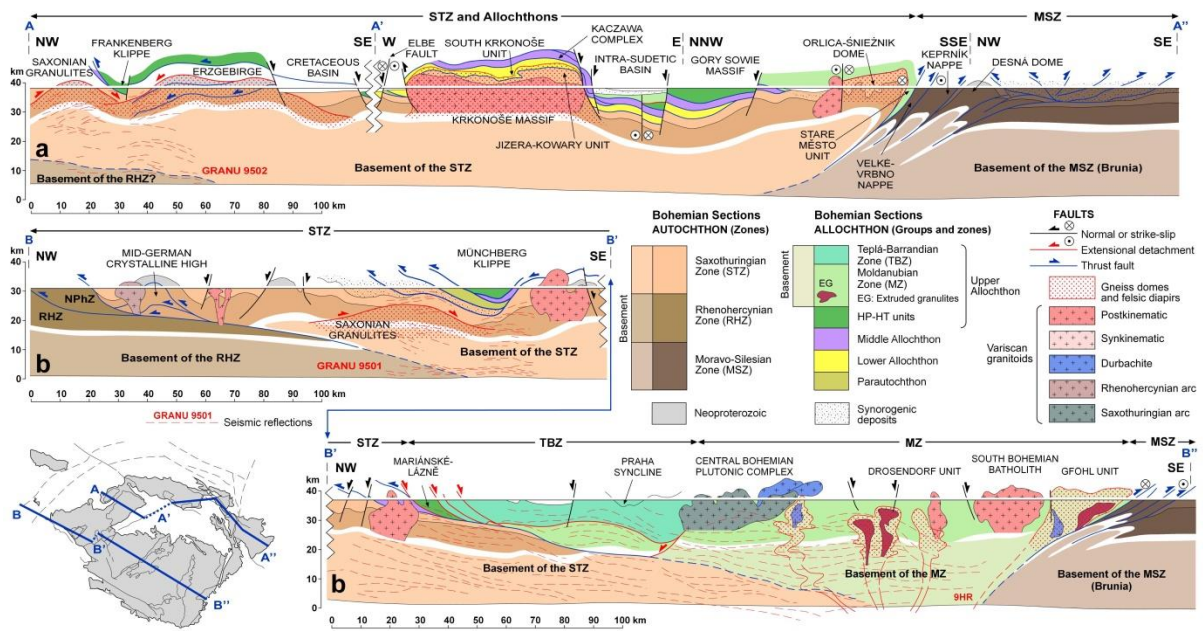


Fig. 7

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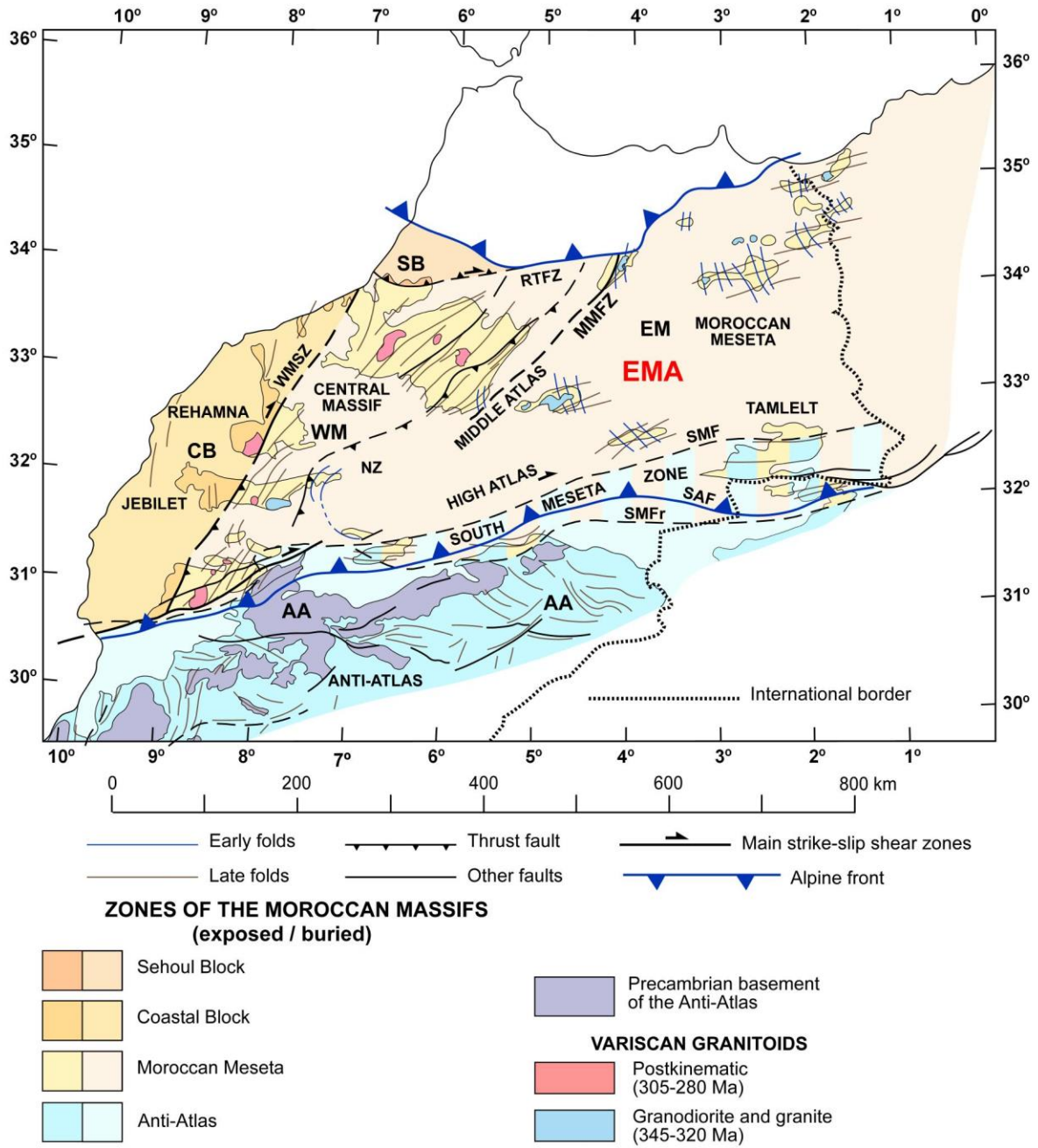


Fig. 8

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Northern Morocco	Iberian Massif	Armorican Massif	French Massif Central	Bohemian Massif
	Galicía-Trás-os-Montes Zone IP Upper Allochthon	South Armorican Domain IP Upper Allochthon	Upper Gneiss Unit IP and LP metasediments	Teplá-Barrandian Zone Moldanubian Zone
	Galicía-Trás-os-Montes Zone HP-HT Upper Allochthon	South Armorican Domain HP-HT Upper Allochthon	Upper Gneiss Unit Leptyno-Amphibolitic Complex	Teplá-Barrandian Zone HP-HT Units Saxothuringian Zone Allochthonous HP-HT Units
	Galicía-Trás-os-Montes Zone Lower Devonian Middle Allochthon	South Armorican Domain Lower Devonian Middle Allochthon	Middle Allochthon Limousin Ophiolite	Saxothuringian Zone Prasinite Units Sudetic Silurian-Devonian Ophiolitic U.
	Galicía-Trás-os-Montes Zone Cambro-Ordovician Middle Allochthon	South Armorican Domain Cambro-Ordovician Middle Allochthon	Middle Allochthon Leptyno-Amphibolitic Complex	Saxothuringian Zone Sudetic Cambro-Ordovician Oceanic U.
	Galicía-Trás-os-Montes Zone Lower Allochthon	South Armorican Domain Lower Allochthon	Lower Gneiss Unit	Saxothuringian Zone Sudetic Lower Allochthon
	Galicía-Trás-os-Montes Zone Parautochthon	South Armorican D. Vendée littorale Central Armorican Domain St. Georges-sur-Loire Unit	Parautochthon	Saxothuringian Zone Münchberg Parautochthon
Anti-Atlas	Cantabrian Zone			
Moroccan Meseta Eastern and Western Mesetas	West Asturian-Leonese Zone Central Iberian Zone	South Armor. Domain Autochthon Central Armorican Domain	Montagne Noire	
Coastal Block	Obejo-Valsequillo Domain	North Armorican Domain		
	Central Unit	Pays de Léon		Saxothuringian Zone Autochthon
Sehoul Block	Ossa-Morena Zone			
	South Portuguese Zone			Rhenohercynian Zone Moravo-Silesian Zone?

VARISCAN OROGENIC MAGMATISM	Northern Morocco	Iberian Massif	Armorican Massif	French Massif Central	Bohemian Massif
Postkinematic granitoids: heat input and melts partly derived from thinned mantle lithosphere	305-280 Ma WM, CB	310-285 Ma WALZ, CIZ, OMZ, GTMZ	310-285 Ma NAD, CAD, SAD	310-280 Ma FMC	310-300 Ma MZ
Synkinematic granitoids: related with crustal thickening, LP-HT metamorphism and extension Pink: peraluminous, S-type Blue: metaluminous, I-type	345-320 Ma EM, WM	322-305 Ma WALZ, CIZ, GTMZ	340-290 Ma NAD, CAD, SAD	345-310 Ma FMC, VM	335-310 Ma STZ, TBZ, MZ
	335-320 Ma EM, WM	347-320 Ma WALZ, CIZ, GTMZ		345-315 Ma N of FMC, VM	
Mg-K rich magmatism: partial melting of subducted lower crust or enriched subcontinental upper mantle		~340 Ma GTMZ		345-332 Ma VM	345-335 Ma MZ
Rhenohercynian arc: tholeiitic and calc-alkaline subduction of Rhenohercynian oceanic lithosphere		355-335 Ma OMZ, SPZ		350-330 Ma FMC, VM	370-325 Ma MGCH
Saxothuringian Arc: Subduction of Saxothuringian/ Mid-Variscan oceanic lithosphere			375-370 Ma SAD	380-360 Ma NW of FMC	380-360 Ma NE of FMC
Silurian-Devonian arc: Subduction of the Rheic Ocean	~370 Ma SB	~400 Ma SPZ	~390 Ma NAD		425-390 Ma MGCH

Fig. 9

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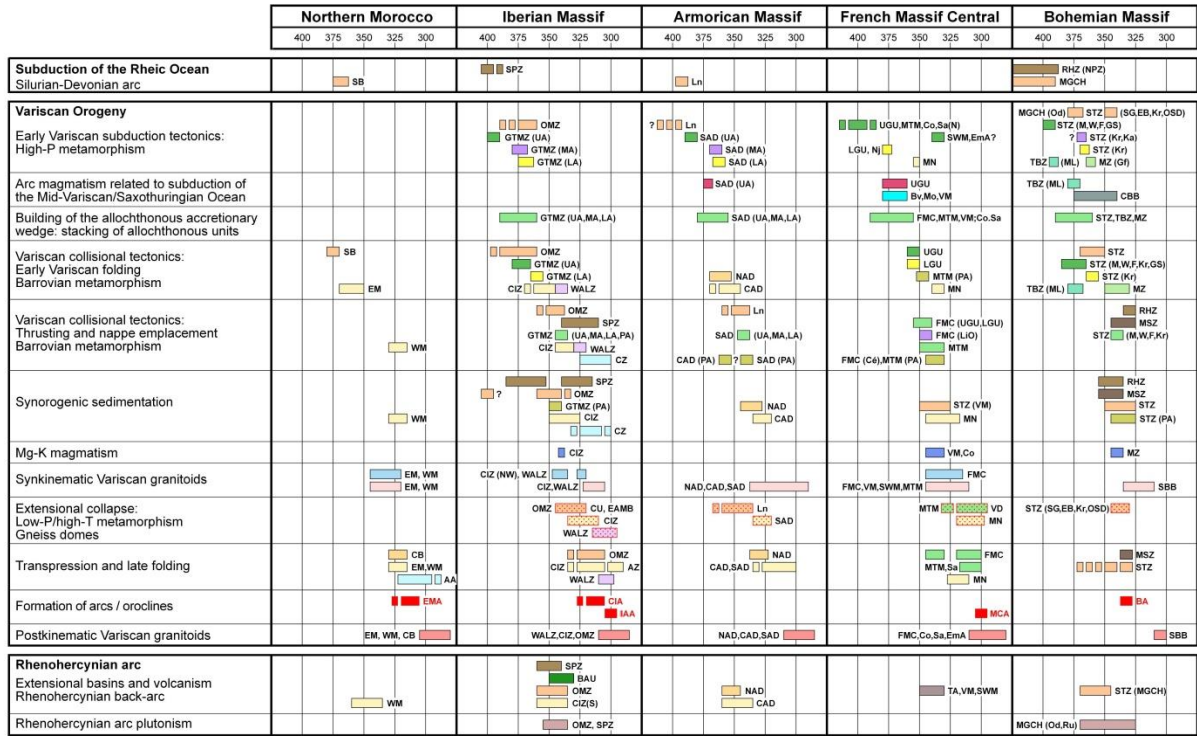


Fig. 10

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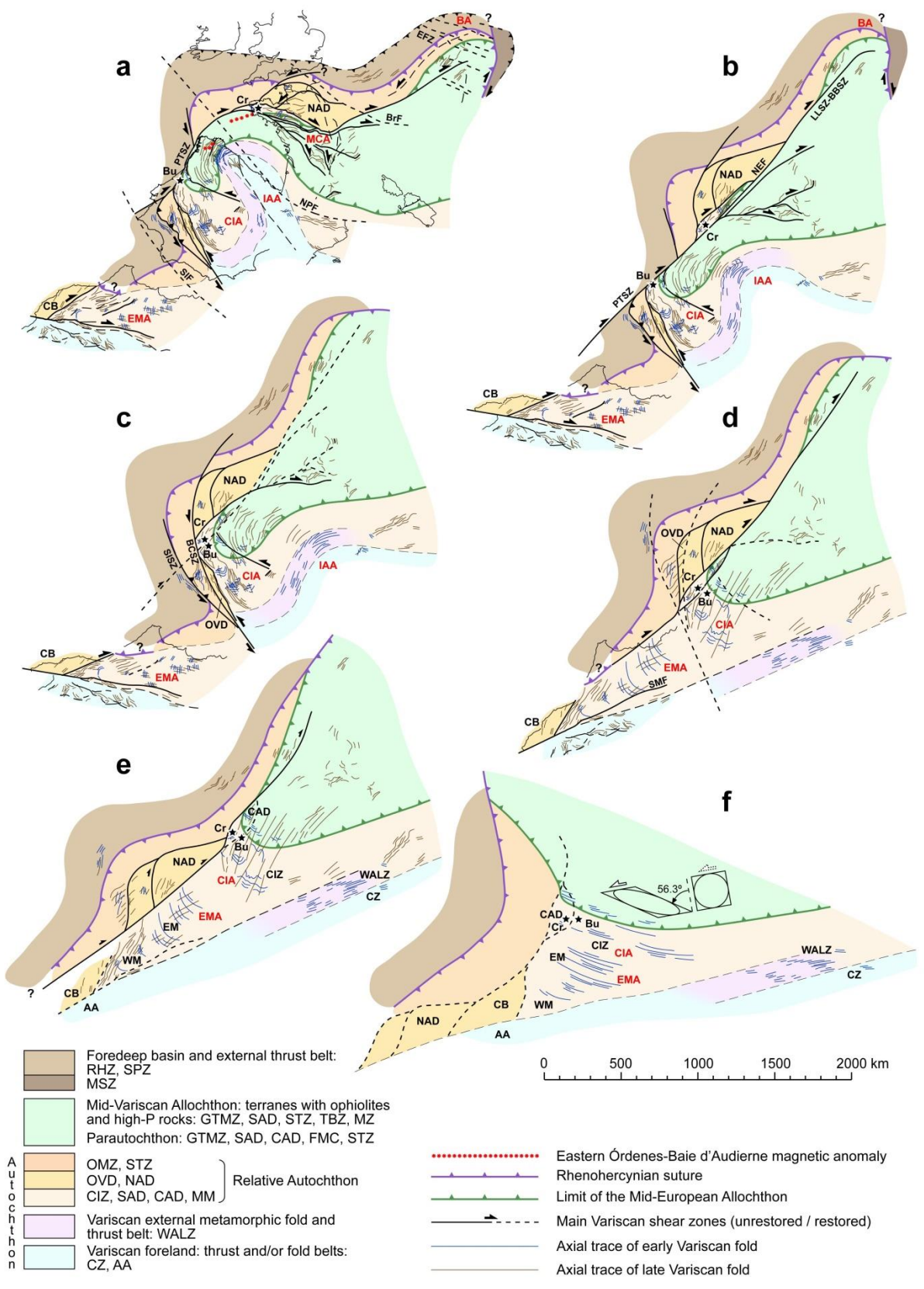


Fig. 11

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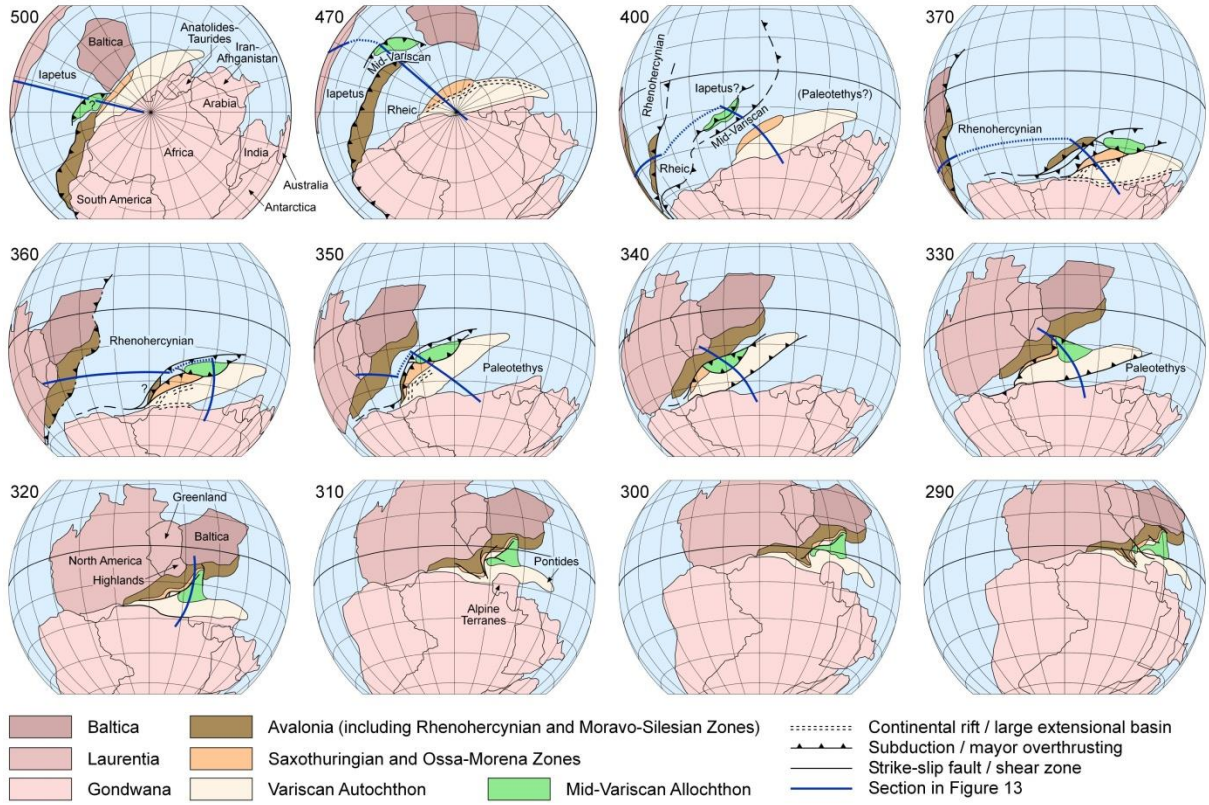


Fig. 12

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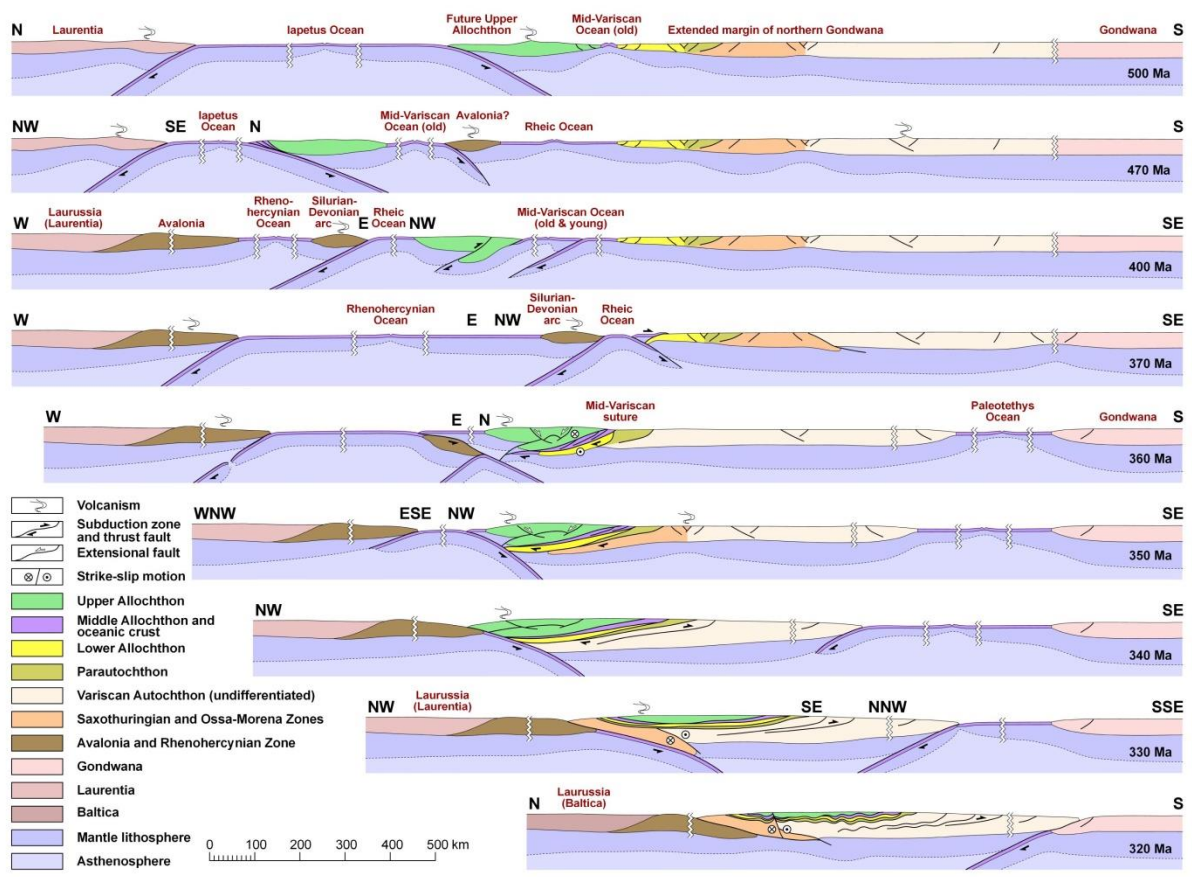


Fig. 13

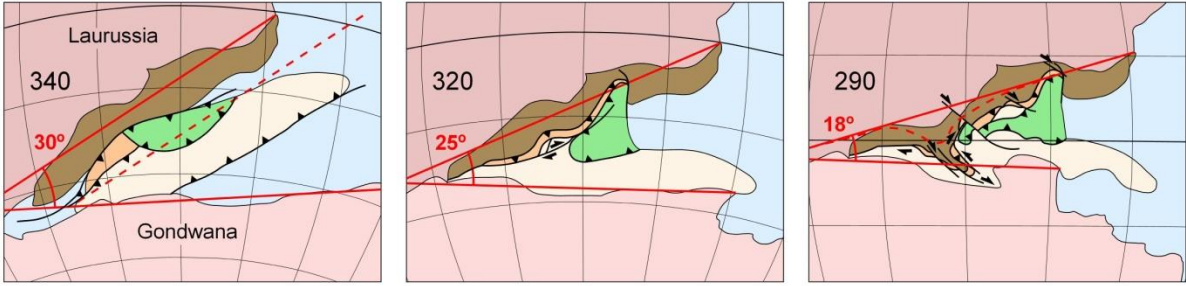


Fig. 14

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Abbreviations

The list includes in alphabetical order the abbreviations used in the text and figures.

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23	AA	Anti-Atlas (Morocco)	EB	Erzgebirge Massif (Bohemian Massif)
24	AaM	Aar Massif (Alps)	EFZ	Elbe Fault Zone (Bohemian Massif)
25	AMM	Argentera Massif (Alps)	EM	Eastern Meseta (Morocco)
26	Au	Audierne Bay Complex (Armorican Massif)	Ema	External Massifs of the Alps
27	AZ	Axial Zone (Pyrenees)	EMA	Eastern Meseta Arc (Morocco)
28	B	Bragança (NW Iberia allochthonous complex)	Es	Essarts Unit (Armorican Massif)
29	BA	Bohemian Arc	EV	Erbendorf-Vohenstrauß Unit (Bohemian Massif)
30	BAU	Beja-Acebuches Unit (Iberian Massif)	F	Frankenberg Klippe (Bohemian Massif)
31	BBSZ	Baden-Baden Shear Zone (Schwarzwald Massif)	FF	Franconian Fault (Bohemian Massif)
32	BC	Bois-de-Cené Unit (Armorican Massif)	FMC	French Massif Central
33	BCSZ	Badajoz-Córdoba Shear Zone (Iberian Massif)	GCF	Granville-Cancalle Fault (Armorican Massif)
34	BeM	Belleme Massif (Alps)	Gf	Gföhl Unit (Bohemian Massif)
35	BL	Blansky les Unit (Bohemian Massif)	GN	Griessen Nappe (Rhenish Massif)
36	BM	Basque Massifs	GS	Göry Sowte Massif (Bohemian Massif)
37	BfF	Bray Fault	GTMZ	Galicia-Trás-os-Montes Zone (Iberian Massif)
38	Bu	Buçaco (Portugal)	HM	Harz Mountains (Germany)
39	Bv	Brévenne Unit (French Massif Central)	HP-HT	High-P/High-T Upper Allochthon (European massifs)
40	CAD	Central Armorican Domain	IAA	Ibero-Armorican Arc
41	CB	Coastal Block (Morocco)	Ib	Ibiza (Balearic Islands)
42	CBPC	Central Bohemian Plutonic Complex	IG	Ile-de-Groix (Armorican Massif)
43	Cé	Cévennes (French Massif Central)	IP	Intermediate-P Upper Allochthon (European massifs)
44	CF	Combrailles Fault (French Massif Central)	IRB	Iberian Reflective Body
45	Ch	Champtoceaux Complex (Armorican Massif)	ISB	Intra-Sudetic Basin (Bohemian Massif)
46	ChO	Chamrousse Ophiolite (Alps)	ISF	Intra-Sudetic Fault (Bohemian Massif)
47	CHS	Chantonnay Syncline (Armorican Massif)	JPSZ	Juzbado-Penalva Shear Zone (Iberian Massif)
48	CIA	Central Iberian Arc	Ka	Kaczawa Complex (Bohemian Massif)
49	CIZ	Central Iberian Zone	KH	Kuthna Hora Unit (Bohemian Massif)
50	Co	Corsica	Kr	Krkonoše Massif (Bohemian Massif)
51	CO	Cabo Ortegal (NW Iberia allochthonous complex)	LA	Lower Allochthon (European massifs)
52	Cr	Crozon (Armorican Massif)	LB	Lausitz Block (Bohemian Massif)
53	CU	Central Unit (Iberian Massif)	LC	Lizard Complex (England)
54	CV	Cinco Villas (Basque Massif, Pyrenees)	LiO	Limousin ophiolite (French Massif Central)
55	CZ	Cantabrian Zone (Iberian Massif)	LLSZ	Lalaye-Lubine Shear Zone (Vosges Massif)
56	DF	Danube Fault (Bohemian Massif)	Ln	Léon (Armorican Massif)
57	Dr	Drosendorf Unit (Bohemian Massif)	Ly	Lyonnais (French Massif Central)
58	DRF	Domain of Recumbent Folds (Iberian Massif)	M	Morais (NW Iberia allochthonous complex)
59	DUF	Domain of Upright Folds (Iberian Massif)	Ma	Majorca (Balearic Islands)
60	EAMB	Évora-Araccena Metamorphic Belt (Iberian Massif)	MA	Middle Allochthon (European massifs)
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18	MBM	Mont Blanc Massif (Alps)	RHZ	Rhenoherynian Zone
19	MCA	Massif Central Arc	RTFZ	Rabat-Tiflet Fault Zone (Morocco)
20	Me	Menorca (Balearic Islands)	Ru	Ruhla Massif (Germany)
21	MECS	Maures-Estérel-Corsica-Sardinia block	Sa	Sardinia
22	MGCH	Mid-German Crystalline High	SAD	South Armorican Domain
23	ML	Mariánské Lázně Unit (Bohemian Massif)	SAF	South Atlas Fault (Morocco)
24	MM	Moroccan Meseta	SASZ-N	South Armorican Shear Zone, Northern branch
25	MMFZ	Middle Meseta Fault Zone (Morocco)	SASZ-S	South Armorican Shear Zone, Southern branch
26	MN	Montagne Noire (French Massif Central)	SB	Sehoul Block (Morocco)
27	Mo	Morvan Arc (French Massif Central)	SBB	South Bohemian Batholith
28	Ms	Mauges Unit (Armorican Massif)	SBF	Sudetic Boundary Fault (Bohemian Massif)
29	MSZ	Moravo-Silesian Zone (Bohemian Massif)	SG	St. Georges-sur-Loire Unit (Armorican Massif)
30	MT	Moldanubian Thrust (Bohemian Massif)	SGD	Schist-Greywacke Domain (Iberian Massif)
31	MTM	Maures-Tanneron Massif (Southern France)	SGr	Saxonian Granulitebirge (Bohemian Massif)
32	MTu	Malpica-Tui (NW Iberia allochthonous complex)	SHF	Sillon Houiller (French Massif Central)
33	Mü	Münchberg Klippe (Bohemian Massif)	SIF	South Iberia Fault
34	MZ	Moldanubian Zone (Bohemian Massif)	SISZ	Southern Iberian Shear Zone
35	NAD	North Armorican Domain	SK	South Křkonosé Unit (Bohemian Massif)
36	NASZ	North Armorican Shear Zone	SLF	South Limousin Fault (French Massif Central)
37	NEF	Nort-sur-Erdre Fault (Armorican Massif)	SM	Staré Město Unit (Bohemian Massif)
38	Nj	Najac Massif (French Massif Central)	SME	South Meseta Fault (Morocco)
39	NM	Northern Massifs (Pyrenees)	SMEFr	South Meseta Front (Morocco)
40	NMe	Nové Město Unit (Bohemian Massif)	Sp	Spessart Massif (Germany)
41	NPF	North Pyrenean Fault	SPhZ	Southern Phyllite Zone (Germany)
42	NPhZ	Northern Phyllite Zone (Germany)	SPZ	South Portuguese Zone
43	NSSB	Northern Sierra de Sevilla Batholith (Iberian Massif)	STZ	Saxothuringian Zone
44	NZ	Nappe Zone (Morocco)	Sv	Svatka window (Bohemian Massif)
45	O	Ordenes (NW Iberia allochthonous complex)	SWM	Schwarzwalde Massif (Germany)
46	Od	Odenwald Massif (Germany)	TA	Tufts Anthracifères (French Massif Central)
47	OF	Ouzilly Fault (French Massif Central)	TBZ	Teplá-Barrandian Zone (Bohemian Massif)
48	OFZ	Odra Fault Zone (Bohemian Massif)	Th	Thaya window (Bohemian Massif)
49	OMZ	Ossa-Morena Zone (Iberian Massif)	TP	Thiviers-Payzac Unit (French Massif Central)
50	OS	Orlica-Šnieżnik Dome (Bohemian Massif)	TTZ	Teisseyre-Tornquist Zone (Baltica)
51	OSD	Olló de Sapo Domain (Iberian Massif)	UA	Upper Allochthon (European massifs)
52	OVD	Obejo-Valsequillo Domain (Iberian Massif)	UB	Ursuya-Baygura (Basque Massif, Pyrenees)
53	PA	Parautochthon (NW Iberian Massif)	VD	Velay Dome (French Massif Central)
54	PASZ	Posada-Asinara Shear Zone (Sardinia)	VF	Variscan Front
55	PF	Pfahl Fault (Bohemian Massif)	VM	Vosges Massif (France)
56	PL	Pulo do Lobo Unit (Iberian Massif)	W	Wildentfels Klippe (Bohemian Massif)
57	PS	Praha Syncline (Bohemian Massif)	WALZ	West Asturian-Leonese Zone (Iberian Massif)
58	PTSZ	Porto-Tomar Shear Zone (Iberian Massif)	WM	Western Meseta (Morocco)
59	QR	Quinto Real (Basque Massif, Pyrenees)	WMSZ	Western Meseta Shear Zone (Morocco)
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