

1 The enigmatic curvature of Central Iberia and its puzzling kinematics

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12 Abstract

13 The collision between Gondwana and Laurussia that formed the latest supercontinent,
14 Pangea, occurred during Devonian to Early Permian times and resulted in large-scale orogeny
15 that today transects Europe, northwest Africa and eastern North America. This orogen is
16 characterized by an 'S' shape corrugated geometry in Iberia. The northern curve of the
17 corrugation is the well known and studied Cantabrian (or Ibero-Armorian) Orocline and is
18 convex to the east and towards the hinterland. Largely ignored for decades, the geometry and
19 kinematics of the southern curvature, known as the Central Iberian curve, are still ambiguous
20 and hotly debated. Despite the paucity of data, the enigmatic Central Iberian curvature has
21 inspired a variety of kinematic models that attempt to explain its formation with little consensus.
22 This paper presents the advances and milestones in our understanding of the geometry and
23 kinematics of the Central Iberian curve from the last decade, with particular attention to
24 structural and paleomagnetic studies.

25 When combined, the currently available datasets suggest that the Central Iberian curve
26 did not undergo regional differential vertical-axis rotations during or after the latest stages of the
27 Variscan orogeny, and did not form as the consequence of a single process. Instead, its core is
28 likely a primary curve (i.e. inherited from previous physiographic features of the crust) whereas
29 the curvature in areas outside the core are dominated by folding interference during the
30 Variscan orogeny or more recent Cenozoic (Alpine) tectonics.

31 **Keywords**

32 **Central Iberia Curve, Variscan orogen, Iberia, Cantabrian Orocline, Curved orogens,**

33 **Pangea**

34 1 Introduction

35 Mountain belt systems are the most striking product of plate tectonics. In addition to their
36 astonishing visual effect, marking the locations where ancient and modern plates collided,
37 orogenic belts often preserve a variety of rocks that have the potential to illuminate the entirety
38 of the systems pre- and syn-orogenic history. One of the most striking characteristics of the
39 majority of Earth's orogens are their curvature in plan-view (e.g. van der Voo, 2004; Marshak,
40 2004; Rosenbaum, 2014). The degree of orogenic curvature may range from a few degrees of
41 deflection in structural trend (e.g. Kopet Dag, Iran), to 180° of arc curvature (e.g. Kazakhstan arc
42 and the Carpathians). The kinematics, structural and geodynamic implications of these systems
43 are as varied as their geometries (Marshak, 2004; Weil and Sussman, 2004; Johnston et al.,
44 2013). For example, some orogenic curvatures are hypothesized to be the consequence of
45 physiographic features of the basement that pre-date orogen formation, such as irregular basin
46 architectures or plate margin salients and recesses (e.g. Jura mountains, Hindle et al., 2000),
47 which then control the growth geometry of the ensuing orogen. These systems are known as
48 primary arcs and reflect pre-orogenic geometries and show no significant or systematic vertical-
49 axis rotations along their structural length. On the other hand, oroclines, as classically defined
50 by Carey in 1956, involve systematic differential vertical-axis rotations subsequent to initial
51 orogenic shortening: different sectors of an orogen rotate with variable magnitudes or in
52 opposite directions (e.g. Li et al., 2012). Rotations in Orocines may occur at a range of scales,
53 from thrust emplacement at upper crustal levels (e.g. Izquierdo-Llavall et al., 2018), up to a
54 lithospheric-scale vertical-axis folding (e.g. Li et al., 2018). They can occur as single curves (e.g.
55 Maffione et al., 2009), coupled curves (Johnston, 2001), or in trains of curves (Li and
56 Rosenbaum, 2014). Orocines can form during the main orogenic building event, known as
57 progressive oroclines (Johnston et al., 2013; e.g., the Wyoming salient, Yonkee and Weil, 2010,
58 and Weil et al., 2010) or in a subsequent tectonic pulse, so-called secondary oroclines (Weil
59 and Sussman, 2004). Understanding the kinematics and mechanisms of curvature formation in
60 mountain belts is a critical step to understanding orogenesis in 4D and to evaluate their
61 geodynamic consequences and paleogeographic implications.

62 The Variscan-Alleghanian orogeny resulted in the suturing of Gondwana and Laurussia
63 during Devonian-Carboniferous times, and ultimately led to the formation of the supercontinent,
64 Pangea. This long and sinuous orogen runs for >8000 km along strike and is ca. 1000 km wide,
65 transecting across Europe, to northwest Africa and into eastern North America. The final stages
66 of Pangea amalgamation (e.g. Nance et al., 2010) modified the Western Europe sector of the

67 belt into its characteristic sinuous shape, which today traces at least one, and perhaps four arcs
68 from Poland to Brittany, and then across the Bay of Biscay (Cantabrian Sea) into Iberia, where
69 the system is today truncated by the Betic Alpine orogeny in southeast Iberia (Fig. 1; e.g. Weil
70 et al., 2013). The southern truncation of the Variscan in Europe hinders a precise correlation
71 with equivalent age outcrops in NW Africa.

72 Within the Iberian Peninsula, the orogen is characterized by two large-scale curves (Fig.
73 2): (1) to the north is the well studied and nearly 180° secondary orocline, the Cantabrian (a.k.a.
74 Ibero-Armorian) Orocline, which buckled a segment of the Variscan belt from ~315 to ~290 Ma
75 (e.g. Weil et al., 2019 and references therein); and (2) to the south is a curve with disputed
76 magnitude and kinematics, ~~and~~ is usually referred to as the Central Iberian curve/orocline or
77 Castilian bend (Martínez-Catalán et al., 2015). Though there remains tremendous uncertainty
78 on the geometry and kinematics of the Central Iberia~~curve~~ curve, multiple hypotheses exist as to its
79 nature, and disagreements continue on its importance ~~in the tectonic evolution of Europe~~ during
80 the waning stages of Paleozoic global supercontinent construction. The diversity of ~~author's~~
81 interpretations ~~of~~ the Central Iberian curve range from a nonexistent structure (Dias et al.,
82 2016), to being one of the most important pieces ~~to our understand~~ of the late Carboniferous
83 and Permian geodynamics of the Iberian Variscan system (e.g. Martínez-Catalán et al., 2011;
84 2014).

85 This paper reviews the most recent advances on the geometry and kinematics of the
86 Central Iberia~~curve~~ curve, synthesizing what we know and what we don't, and ending with a
87 discussion of the main unsolved issues. We hope that this paper fosters novel studies that will
88 lead to a better understanding of when and which mechanisms acted in the aftermath of the
89 Variscan-Alleghanian orogeny.

90 2 The long and winding orogen

91 The Variscan (~~Europe NW Africa~~) Alleghanian (~~North America~~,  Orogeny is a continental-
92 scale tectonic system (1000 km wide and 8000 km long) that sutured Gondwana and Laurussia
93 together, forming the supercontinent Pangea (e.g. Domeier and Torsvik, 2014; Edel et al., 2018;
94 Pastor-Galán et al., 2019a). The fragments of this system are now dispersed over three
95 continents, Europe, Africa and North America due to the Mesozoic break-up of Pangea (Buiter
96 and Torsvik, 2014; Keppie, 2015). This orogen formed as a consequence of a long and
97 protracted tectonic history that involved several ~~different~~ events, from initial convergence (ca.
98 420 Ma; e.g. Franke et al., 2017), to the consumption of multiple putative oceanic tracts and/or
99 basins that existed between Gondwana and Laurussia (ca. 280 Ma; e.g. Kirsch et al., 2012).

100 The Variscan-Alleghanian orogen itself represents the closing of at least one major ocean, the
101 Rheic (e.g. Nance et al., 2010), whose axial ridge likely failed or subducted at ca. 395 Ma along
102 its paleo-northern margin (e.g. Woodcock et al., 2007; Gutiérrez-Alonso et al., 2008a). Perhaps
103 the orogeny involved other large oceans (Stampfli and Borel, 2002; Franke et al., 2017; 2019),
104 but most surely involved several minor seaways and basins that existed between Gondwana,
105 Laurussia, and several intervening micro-continents (e.g. Azor et al., 2008, Dallmeyer et al.,
106 1997; Kroner and Romer, 2013; Díez-Fernández et al., 2016; Pérez-Cáceres et al., 2017). The
107 final continent-continent collision began after closure of all oceans and intervening seaways.
108 The commencement of this deformation was ~~diachronistic~~ and became progressively younger
109 westwards (in present-day coordinates): with Devonian continent-continent collisions along the
110 eastern boundary, progressing to earliest Permian ages in the westernmost sector (McWilliams
111 et al., 2013; Chopin et al., 2014; López-Carmona et al., 2014; Franke et al., 2017).

112 The present-day geometry of the Variscan-Alleghanian systems has a contorted trace
113 (Fig. 1). In Europe, from east to west, the trend starts with a prominent curve around the
114 Bohemian massif (e.g. Tait et al., 1996), followed by a deflection in the Ardennes-Brabant (e.g.
115 Zegers et al., 2003). In Brittany the outer curvature of the Cantabrian or Ibero-Armorican
116 orocline begins (e.g. van der Voo et al., 1997), and wraps nearly 180° around across the Bay of
117 Biscay as it turns in NW Iberia. The Central Iberian curve marks the final concave to the west
118 curve (in present-day coordinates) and is the focus of this paper (e.g. Aerden, 2004; Martínez
119 Catalán, 2011; Shaw et al., 2012). The orogen continues in North America where, from north to
120 south, it has salients and recesses that undulate back and forth from Atlantic Maritime Canada
121 (e.g. O'Brien, 2012) down along the Pennsylvanian and Alabama curves (e.g. Thomas, 1977).

122 Interpretation on the origin of these curvatures varies widely. The curvatures in North
123 America are argued to be the result of a preexisting irregular margin of Laurentia due to the
124 break-up of the Rodinia supercontinent, which resulted in the formation of orogenic salients and
125 recesses during subsequent Appalachian collision (e.g. Rankin, 1976; Thomas, 1977, 2004). In
126 this case, vertical-axis rotations affected only the upper crustal levels during orogenesis (e.g.
127 Marshak, 1988; Bayona et al., 2003; Hnat and van der Pluijm, 2011). In Europe, the Bohemian
128 and Ardennes-Brabant massif curvatures have poor kinematic constraints. In the Bohemian
129 Massif, some suggest secondary rotations that formed an orocline (Tait et al., 1996), while
130 others suggest little to no vertical-axis rotations and a primary arc (Chopin et al., 2012). The
131 Ardennes-Brabant Massif record some vertical-axis-rotations (e.g. Molina-Garza and
132 Zeijderveld, 1996), but it is unclear if these are a response to progresive or secondary oroclinal
133 bending, or whether rotations only affected the upper crust. The most outstanding example of

134 Variscan-Alleghanian orogen curvature is exposed in the Iberian Massif, with the Cantabrian
135 Orocline and the coupled curvature of Central Iberia.

136 2.1 Two of us: The Variscan orogen in Iberia

137 The western half of the Iberian Peninsula constitutes the Iberian massif, one of the
138 largest exposures of the Variscan orogen and the only place that contains an almost continuous
139 cross section of the orogen (Fig. 2; e.g. Lotze 1945, Julivert 1974, Pérez-Estaún et al., 1991;
140 Ayarza et al., 1998; Simancas et al., 2003; Ribeiro et al. 2007, Martínez Catalán et al., 2014,
141 2019). The majority of the Iberian Massif contains Gondwanan affinity rocks (e.g. Murphy et al.,
142 2008; Pastor-Galán et al., 2013a; Gutiérrez-Marco et al., 2017) and likely represents a proximal
143 piece of the Gondwana margin until its final amalgamation with Pangea (e.g. Pastor-Galán et
144 al., 2013b). Owing to the stratigraphic, structural and petrological styles, the Iberian Massif has
145 been traditionally divided into six tectonostratigraphic zones (Fig. 2; Lozze, 1945; Julivert, 1971):
146 (1) Cantabrian Zone represents a Gondwanan thin-skinned foreland fold-and-thrust belt. It has
147 overall low-grade internal deformation and metamorphism, and represents shortening that
148 occurred during Mississippian times (e.g. Marcos and Pulgar, 1982; Pérez Estaún et al., 1988;
149 Gutiérrez-Alonso 1996; Alonso et al., 2009; Pastor-Galán et al., 2009; 2013b). (2) The West-
150 Asturian Leonese Zone represents a metamorphic fold-and-thrust belt with barrovian
151 metamorphism that collapsed coevally with thrust emplacement onto the Cantabrian Zone (e.g.
152 Martínez-Catalán et al., 1992; Alcock et al., 2009; Martínez-Catalán et al., 2014). (3) The
153 Central Iberian Zone represents the Gondwanan hinterland with Barrovian and Buchan
154 metamorphism and is intruded by igneous rocks of various ages (e.g. Macaya et al., 1991; Díez
155 Balda, 1995; Gutiérrez-Alonso et al., 2018). (4) The Ossa-Morena Zone represents the most
156 distal zone of the Gondwana platform, and is characterized by a metamorphic fold-and-thrust
157 belt with dominantly sinistral displacement (e.g. Robardet and Gutiérrez-Marco, 2004; Quesada,
158 2006). (5) The Galicia-Tras-os-Montes Zone represents a far travelled allochthonous terrane
159 that contains high pressure units and relics of oceanic-like crust (e.g. López-Carmona et al.,
160 2014; Martínez-Catalán et al., 2019). (6) The South Portuguese Zone represents a foreland
161 fold-and-thrust belt with little internal deformation and metamorphism with Avalonian affinity and
162 a strong left-lateral component of shear (e.g. Pereira et al., 2012; Pérez-Cáceres et al., 2016;
163 Oliveira et al., 2019). Geographically, the external zones of the Gondwana margin are nested to
164 the north into the core of the Cantabrian Orocline, whereas the hinterland zones are to the west
165 and center of the massif (Fig. 2; e.g. Díaz Balda, 1995; Azor et al., 2019). The southwestern-
166 most extent of Iberia contains a putative suture of the Rheic ocean, as well as a piece of the

167 Laurussian margin fold-and-thrust belt, today preserved in the South Portuguese Zone (e.g.
168 Pereira et al., 2012, 2017; Oliveira et al., 2019).

169 The Gondwanan autho~~cton~~ stratigraphy (Cantabrian, West Asturian-Leonese, Central
170 Iberian and Ossa Morena Zones) consist of a Neoproterozoic arc and back-arc basin (e.g.
171 Fernández-suárez et al., 2014), which evolved to a rift-to-drift Cambrian to Early Ordovician
172 sequence and then to an Ordovician to Late Devonian passive margin basin sequence (e.g.
173 Sánchez-García et al., 2019; Gutiérrez-Marco et al., 2019; Gutiérrez-Alonso et al., submitted).
174 Overall the system transitioned from a relatively isolated Early Cambrian continental, to a
175 restricted marine basin, to development of an open marine platform that was locally punctuated
176 by magmatism (e.g. Gutiérrez-Alonso et al., 2008b; Palero-Fernández, 2015). The Ossa
177 Morena zone represents the outermost platform, followed by an intermediate platform
178 characterized by an asymmetric horst (Central Iberian Zone) and graben (West-Asturian
179 Leonese Zone), which ends in the innermost shelf environment of the Cantabrian zone (Fig. 3;
180 e.g. Gutiérrez-Marco et al., 2019). The differences between the West Asturian-Leonese and
181 Central Iberian Zone are mainly deeper vs. shallower sedimentary facies (respectively) and a
182 local Lower Ordovician unconformity in the Central Iberian Zone (Toledanian, e.g. Álvaro et al.,
183 2018) that places Lower Ordovician strata atop pre-Cambrian to Cambrian rocks (Fig. 3; e.g.
184 Gutiérrez-Marco et al., 2019). The Central Iberian Zone is divided into two domains: (1) The Ollo
185 de Sapo domain, which contains abundant Lower Ordovician calc-alkaline magmatism (e.g.
186 Díez Montes, 2006; Gutiérrez-Marco et al., 2019); and (2) the ‘Schistose—~~greywacke~~ Domain’
187 characterized by a predominance of outcrops of Neoproterozoic to Lower Cambrian
188 sedimentary rocks (e.g. Gutiérrez-Marco et al., 2019 and references therein).

189 The Galicia Tras-os-Montes Zone (Farias et al., 1987) is a complex structural stack
190 including a basal schistose unit (Parautochthon; Dias da Silva et al., in press) structurally
191 overlain by mafic rocks with an oceanic-like signature and other far-traveled rocks under high-
192 pressure metamorphism (e.g. López-Carmona et al., 2014; Martínez-Catalán et al., 2019). The
193 oceanic rocks of this zone are classically interpreted as a Rheic Ocean suture (e.g. Martínez
194 Catalán et al., 2009). Recent interpretations support its origin as a minor oceanic basin or
195 seaway within the realm of Gondwana (e.g. Pin et al., 2002; Arenas et al., 2016).

196 The South Portuguese Zone constitutes the Laurussian foreland fold-and-thrust belt in
197 the Iberian Variscides (e.g. Pereira et al., 2012; Pérez-Cáceres et al., 2017). It contains three
198 units: (1) the Pulo de Lobo, a low grade metamorphic accretionary prism with clastic
199 sedimentary rocks and basalts with MORB signature (e.g. Azor et al., 2019; Pérez-Cáceres et
200 al., this volume); (2) The Iberian Pyrite Belt, which is a world class volcanogenic massive sulfide

deposit formed between 390 and 330 Ma (e.g. Oliveira et al., 2019a; 2019b); and (3) the Baixo Alentejo Flysch, which is located to the southwest and is a syn-orogenic composite turbiditic sequence with ages from ~330 to ~310 Ma (Oliveira et al., 2019b). The boundary between the South Portuguese and Ossa Morena zones is a sinistral shear zone (so-called Beja-Acebúches, Quesada and Dallmeyer., 1994; Pérez-Cáceres et al., 2016) that contains a strongly deformed amphibolitic belt with oceanic affinity (Munha et al., 1986; Munha, 1989; Quesada et al., 2019). This belt potentially represents dismembered relics of the Rheic ocean and/or a subsidiary seaway that opened during a Variscan transtension event in SW Iberia (e.g. Pérez-Cáceres et al., 2015; Quesada et al., 2019).

Finally, Paleozoic rocks occur sporadically within the Alpine Betic chain. Their lithological monotony, paucity of fossils, and the intensity of deformation and metamorphism during Alpine orogeny, make recognizing the original features of the different successions challenging (e.g. Martín-Algarra et al., 2019). Some faunal and detrital zircon studies suggest that the Paleozoic outcrops in the Betics may be similar to the most seaward realms of the Gondwanan platform (i.e., the Cantabrian Zone; e.g. Rodríguez-Cañero et al., 2018; Jabaloy-Sánchez et al., 2018). Following the latest plate reconstructions of the Mediterranean during Meso-Cenozoic times, the Paleozoic units of the Betic-Rif chain may have been located proximal to the present-day position of the Balearic Islands (van Hinsbergen et al., 2020).

The Variscan orogen in Iberia shows multiple deformation, metamorphic, and magmatic events (e.g. Martínez-Catalán et al., 2014; Azor et al., 2019; Fig. 2) that evolved diachronously from the suture towards the external zones (Dalmeyer et al., 1997): (1) An initial continent-continent collision began ca. 370-365 Ma, which produced ~~high pressure~~ metamorphism (e.g. Lopez-Carmona et al. 2014). (2) Between 360 and 330 Ma a protracted shortening phase occurred, frequently divided into main phases C1 and C2, that were accompanied by Barrovian type metamorphism (e.g. Dias da Silva et al., in press) and plutonism at ~340 Ma (e.g. Gutiérrez-Alonso et al., 2018). (3) An extensional collapse event, so-called E1, occurred at ~333-317 Ma, which formed core-complexes and granitic domes in the Central Iberian and West Asturian-Leonese zones (Fig. 2C; e.g. Alcock et al., 2009; Díez-Fernández and Pereira, 2016; López-Moro et al., 2018). This event is coeval and genetically linked to the formation of the foreland fold-and-thrust-belt of the Cantabrian Zone (e.g. Gutiérrez-Alonso, 1996). (4) A late Carboniferous shortening event (C3) occurred ca. 315-290 Ma and is argued to have resulted in the formation of the Cantabrian Orocline and was accompanied by the intrusion of mantle derived granitoids (Fig. 2C; e.g. Gutiérrez-Alonso et al., 2011a, 2011b; Pastor-Galán et al., 2012a). (5) A final early Permian extensional event (E2), mostly found in the Central Iberian

235 Zone, resulted in the formation of core complexes and regional doming (Dias da Silva et al., in
236 press). (6) A final shortening event (C4), possibly coeval with E2, resulted in widespread brittle
237 deformation (e.g. Azor et al., 2019; Fernández-Lozano et al., 2019).

238 In SW Iberia, the aforementioned Variscan deformation events are characterized by a
239 dominant sinistral component, which contrasts with the general dextral component recognized in
240 most other regions of the orogen (e.g. Martínez Catalán et al., 2011; Gutiérrez-Alonso et al.,
241 2015). Early collisional structures (C1) formed NE-vergent recumbent folds in the southernmost
242 Central Iberian Zone and SW-vergent folds and thrusts in the Ossa Morena and South
243 Portuguese zones. This phase continued with a transtensional event that heterogeneously
244 extended the continental lithosphere (e.g. Pérez-Cáceres et al. 2015). Coevally, an important
245 extension-related magmatic event happened, perhaps assisted by a plume-type mantle
246 (Simancas et al. 2006) or a slab break-off (Pin et al. 2008). After this transtensional event,
247 significant left-lateral transpression occurred forming the extensive shear zones to the north and
248 south of Ossa Morena Zone (Fig. 2B), which accommodated the majority of the transcurrent
249 motion. However, left-lateral displacements are observed all along the Ossa Morena and South
250 Portuguese zones. Pérez-Cáceres et al. (2016) estimated over 1000 km of collisional
251 convergence in SW Iberia, most of which corresponds with left-lateral displacements parallel to
252 terrane boundaries.

253 3 Synthesis on the Geometry and Kinematics of the Cantabrian 254 Orocline

255 Understanding the geometry, kinematic evolution and mechanics of curved mountain
256 systems is crucial to developing paleogeographic and tectonic reconstructions (e.g. Marshak,
257 2004; Van der Voo, 2004; Li et al., 2012; van Hinsbergen et al., 2020). Introduced by Carey
258 (1955 p.257), an orocline (from Greek ὄπος, mountain, and κλίνω, bend) is "...an orogenic
259 system, which has been flexed in plan to a horse-shoe or elbow shape." Although sometimes
260 used in the literature as a geometric description of any orogenic curvature, herein orocline is
261 strictly used as a the term for map-scale bends that underwent vertical-axis rotations (Weil and
262 Sussman, 2004; Johnston et al., 2013; Pastor-Galán et al., 2017a). The kinematic classification
263 of curved mountain belts (Weil and Susman, 2004; Johnston et al, 2013) distinguishes two end
264 members: (1) Primary orogenic curves, which describe those systems in which curvature is a
265 primary feature of the orogen and formed without significant or systematic vertical-axis rotations,
266 and (2) Secondary oroclines, where orogenic curvature was acquired due to vertical-axis
267 rotations subsequent to primary orogenic building. Those systems whose curvature is the

268 product of vertical-axis rotation during the primary orogenic pulse and/or only a portion of the
269 observed curvature is secondary are progressive oroclines.

270 The orocline test (or strike test), evaluates the relationship between changes in regional
271 structural trend (relative to a reference trend for an orogen) and the orientations of a given
272 geologic fabric element or magnetization (relative to a reference direction). In terms of
273 evaluating developmental kinematics, the most relevant geologic marker is paleomagnetic
274 declination, which can be used to quantitatively evaluate total and systematic rotations as a
275 function of along-strike variability. Once acquired, data is plotted on Cartesian coordinate axes
276 with the strike (S) of the orogen (relative to a reference) along the horizontal axis, and the fabric
277 azimuth (F, relative to a reference) along the vertical axis. The test originally used a basic least-
278 squares (OLS) regression (Schwartz and Van der Voo, 1983) to estimate the slope (coded m in
279 formulas), ideally between 0 and 1, which then is interpreted with respect to vertical-axis
280 kinematics. More recently, Yonkee and Weil (2010b) and Pastor-Galán et al. (2017a) introduced
281 more robust statistics to estimate the slope and its uncertainty, considering and propagating
282 errors of the input data. Primary orogenic bends show no change of paleomagnetic declination
283 orientations with varying structural trend, and therefore the slope is expected to be 0. In
284 progressive oroclines, the declination variation records some fraction of the total observed
285 orogenic strike variability, and thus the slope would range between 0 and 1, depending on the
286 amount of primary curvature. Secondary oroclines are those in which the paleomagnetic vectors
287 record 100% of the rotation, yielding slopes of 1, meaning that the orogenic system must have
288 started as a roughly linear system that then underwent secondary vertical-axis rotations until its
289 present-day curvature was acquired. The slope obtained with the orocline test can only be
290 confidently interpreted when the chronology of fabric formation is well known.

291 The trend of the Variscan belt in Iberia follows a sinuous “S” shape that is especially
292 prominent in the northwest region of the Iberian Peninsula, and then becomes more subtle due
293 to the predominance of younger cover sequences in the central and eastern regions of the
294 peninsula (Fig. 1 and 2). This dramatic geometry has stimulated a century long scientific debate
295 as to its origin (e.g. Suess, 1892; Staub, 1926; Martínez Catalán et al., 2015). To the north and
296 convex to the west is the Cantabrian Orocline, and to the center-south and convex to the east is
297 the Central Iberian curve. The overall trend of the Cantabrian Orocline starts in Brittany (France)
298 and southern England and then curves through the Bay of Biscay and then south into central
299 north Iberia (Fig. 1, 2 and 4). The Cantabrian Orocline (also known as Ibero-Armorian Orocline/
300 Arc, Asturian Arc or Cantabrian-Asturias Arc) is arguably the first curved orogen that was
301 scientifically described, recognized by the change in structural trend of mapped thrusts and fold

302 axes (Schultz, 1858, Barrois, 1882, Suess, 1892). The Cantabrian Orocline traces an arc with a
303 curvature close to 180° within the central Cantabrian Zone (the Gondwanan foreland in Iberia,
304 fig. 2), and opens to approximately 150° as one moves to the outer arc reaches (Fig. 1). At the
305 crustal-scale, the Cantabrian Orocline represents a ~~first order~~ vertical-axis buckle fold in plan-
306 view that refolds pre-existing Variscan structures (e.g. Julivert and Marcos, 1973; Weil et al.,
307 2001). The inner arc of the orocline, or the Cantabrian Zone is characterized by tectonic
308 transport towards the core of the orocline, i.e., the orocline has a contractional core, where low
309 finite strain values and locally developed cleavage occur (Pérez-Estaún et al., 1988; Gutiérrez-
310 Alonso, 1996; Pastor-Galán et al., 2009). Within the inner core a variety of structures record
311 non-coaxial strain, which produced complex interference folds and rotated thrust sheets (e.g.
312 Julivert and Marcos, 1973; Julivert and Arboleya, 1984; Pérez-Estaún et al., 1988; Aller and
313 Gallastegui, 1995; Weil, 2006, 2013; Pastor-Galán et al., 2012b; Shaw et al., 2015; 2016a; Del
314 Greco et al., 2016). In contrast, the outer arc shows a ca. 150° interlimb angle vertical-axis fold
315 that was accommodated by significant shearing, both dextral and, in lesser amounts, sinistral
316 penecontemporaneous to vertical-axis rotation (Gutiérrez-Alonso et al., 2015). Weil et al. (2013,
317 2019) extensively review the geometry of the Cantabrian Orocline.

318 All kinematic data studied so far support a model in which the Cantabrian Orocline
319 formed due to secondary vertical-axis rotation in a period of time younger than 315 Ma and
320 older than 290 Ma. Overall, the southern limb of the orocline rotated counterclockwise (CCW)
321 and the northern limb clockwise (CW; Fig. 4). Orocline formation happened subsequent to the
322 main shortening phases of the orogen (C1 and C2) and late-stage orogenic collapse (E1), and
323 therefore, it is an ideal example of a secondary orocline in the strictest sense. Development of
324 the Cantabrian Orocline requires the existence of a roughly linear orogenic belt during early
325 Variscan closure of the Rheic Ocean (with a roughly N-S orientation in present-day
326 coordinates), which was subsequently bent in plan-view into an orocline during late stages of
327 Pangea amalgamation. Such interpretation is grounded in paleomagnetic studies (e.g. Hirt et
328 al., 1992; Parés et al. 1994; Stewart, 1995; van de Voo et al., 1997; Weil, 2006; Weil et al.,
329 2000; 2001; 2010), along with important contributions from structural (e.g. Gutiérrez-Alonso
330 1992; Kollmeier et al., 2000; Merino-Tomé et al., 2009; Pastor-Galán et al., 2011; 2014; Shaw et
331 al., 2015) and geochronological studies (e.g., Tohver et al., 2008; Gutiérrez-Alonso et al., 2015).
332 Weil et al. (2013) provides a comprehensive review on the kinematic constraints, ~~updated in~~
333 ~~2017a by Pastor Galán et al., and in 2019 by Weil et al.~~

334 4 The intriguing geometry of the Central Iberian curve

335 The more southern Central Iberian curve has a similar magnitude, but opposite
336 curvature compared to the Cantabrian Orocline (Fig. 1 and 2B). This structure has been referred
337 to as the Central Iberian curve, arc, bend or orocline. In this paper we use 'Central Iberian
338 curve'. The other aforementioned terms involve still unknown parameters or are misleading:
339 orocline imply kinematics (Weil and Sussman, 2004); bend refers to a mechanism of formation
340 (e.g. Fossen, 2016); and arc could be ambiguous, since the term is commonly used for volcanic
341 chains. This curvature was first described by Staub (1926) and was termed the Castilian bend.
342 Continental drift pioneers paid some attention to Staub's description (e.g. Holmes, 1929; Du
343 Toit, 1937), but the curved structure remained largely ignored for multiple decades (e.g.
344 Martínez Catalán et al., 2015). The hypothesis of a large-scale curvature in Central Iberia made
345 a comeback at the beginning of the 21st century with a study of Variscan porphyroblast
346 kinematics across Iberia by Aerden ~~in~~ 2004. Since then, several attempts to unveil its geometry
347 and kinematics have been made with contrasting results.

348 The elusive nature of the Central Iberian curve resides in the poor exposure of its
349 putative hinge (Fig. 2). The hinge of the Cantabrian orocline crops out extensively and the
350 changes in thrust and fold axes trend are observable at high-resolution from aerial photographs
351 and are readily mapped using outcrop-scale observations. In contrast, the alleged hinge of the
352 Central Iberian curve is largely covered by Mesozoic and Cenozoic basins (Fig. 2). The
353 curvature is most recognizable at the boundary between the Galicia-Tras os Montes and Central
354 Iberian zones (Fig. 2A; Aerden, 2004; Martínez-Catalán, 2012). The thrust fault that bounds
355 those areas traces close to a 180° of curvature and marks the emplacement of the most distal
356 units. Before the revival of Staub's curved geometry along the entire Central Iberian Zone, there
357 were several attempts to explain the curved shape of the Galicia Tras-os-Montes Zone. Some
358 consider the Galicia Tras-os-Montes Zone a block that escaped during an early Variscan (C1)
359 non-cylindrical collision, forming a extrusion wedge towards the areas undergoing lesser
360 amount of shortening (Martínez-Catalán, 1990, Dias da Silva, 2015; in press); or alternatively a
361 klippe of a larger allochthonous thrust sheet, product of an interference pattern between C2, E1
362 and C3 structures (e.g. Ries and Shackleton, 1971; Martínez Catalán et al., 2002; Rubio
363 Pascual et al., 2013; Díez-Fernández et al., 2015).

364 In addition to the Galicia Tras-os-Montes Zone, other areas showing a certain degree of
365 curvature are to the E and SE of the Central Iberian Zone. There, an approximately 20° change
366 in strike of the Iberian ranges (NE Iberia, Fig. 2A) is observed, which represents the only known
367 outcrop of the hinge of the Central Iberian curve's outer arc. The rest of the curvature has been

368 deduced with indirect observations leading to three competing geometric proposals for the
369 Central Iberia curve (Fig. 2B). The main arguments used to constrain the geometry of the
370 Central Iberian curve are: (1) the geometry of Galicia Tras-os-Montes folds and the orientation
371 of observed garnet inclusion trails (Aerden, 2004; Fig. 2B-1); (2) aligned aeromagnetic
372 anomalies and fold trends in the Iberian ranges and the E-SE Central Iberian Zone (Martínez-
373 Catalán, 2012; Fig. 2A and 2B-2) and; (3) the regional distribution of paleocurrents recorded in
374 Ordovician quartzites (Shaw et al., 2012; Fig. 2B-3 and 3). All proposed geometries share two
375 features: (1) The curvature runs parallel to the Central Iberian Zone, and is located in the
376 center-west of Iberia, and (2) all place the Galicia Tras-os-Montes Zone in the core of the curve
377 with the curves axial trace cross-cutting the Morais Complex, a set of mafic and ultramafic rocks
378 that is roughly circular in shape (Fig. 2B; Dias da Silva et al., in press).

379 Aerden (2004) compared the orientation of inclusions in metamorphic porphyroblasts
380 across the Variscan allochthonous terranes of the NW Iberian Massif, and found that inclusion
381 trails maintain a constant north–south orientation. Comparing such results with the trend of the
382 Variscan fold axes in the central Iberian Zone (Fig. 2A) and a daring interpretation of the
383 aeromagnetic anomalies of the Iberian Peninsula (Fig. 5A), Aerden suggested a geometry in
384 which the Central Iberian curve was more prominent in the outer arc than in the inner arc (Fig.
385 2B-1). In Aerden's view the geometry of the Galicia Tras-os Montes Zone does not represent a
386 large-scale curvature, but rather the original shape of the nappe, perhaps re-tightened during
387 C3 deformation. In contrast, the Iberian Ranges and the SE Central Iberian Zone represent the
388 more curved sector (Fig. 2B-1). In the model of Aerden (2004), the Ossa Morena and South
389 Portuguese Zones are not part of the Central Iberian curvature.

390 Martínez-Catalán (2012) reinterpreted Aerden's analysis of aeromagnetic map data (Fig.
391 5A) and the interpretive structural trends of C1-C2 fold axes from Central Iberian Zone
392 structures (Fig. 2A). In Martínez-Catalán's model, the Central Iberian curvature is a symmetric
393 arcuate shape in which orogen trend changes equally in the inner and outer arc, and is
394 comparable in size to the Cantabrian Orocline, but with opposite curvature and less shortening.
395 This geometric model also excludes the Ossa Morena and the South Portuguese Zones as
396 elements involved in the formation of the curvature (Fig. 2B-2).

397 Finally, Shaw et al. (2012) studied the orientation of paleocurrents in Ordovician
398 Armorian Quartzite (e.g. Aramburu, 2002), which is one of the most prominent rocks exposed
399 in Iberia (Fig. 3). The authors found that paleocurrents fanned outward with respect to the
400 Cantabrian Orocline curve and are approximately perpendicular to the structural trend
401 throughout the peninsula (Fig. 3). Shaw et al. (2012) assumed that the direction and sense of

402 paleocurrents were parallel throughout all zones, and concluded that the Central Iberia curve is
403 a 'S' shape isoclinal structure similar in magnitude to the Cantabrian Orocline (Fig. 2B-3). It is
404 unclear from the Shaw et al. (2012) model the involvement of the Ossa Morena and South
405 Portuguese Zones in the overall curve (if any), nor the prospective location of the external zones
406 of the orogen (Cantabrian Zone) with respect to the overall curvature. 

407 5 Move over once, move over twice: Kinematic constraints

408 Late Variscan kinematic data (315-290 Ma; C3, E2, C4 phases) in the Central Iberian
409 curve were scarce prior to revival of Staub's Central Iberian curve (e.g. Vergés, 1983; Julivert et
410 al., 1983; Parés and van der Voo, 1992). More recently, a wealth of studies have been
411 published on the kinematics of forming the Central Iberian curve (Fig. 2B), which are reviewed
412 below.

413 5.1 Structural Geology and Geochronology

414 Curved orogens that result from differential vertical-axis rotations develop remarkable
415 structures within their hinges where compressive and extensive radial structures often develop
416 in combination with tangential shear structures (e.g. Li et al., 2012; Eichelberger and McQuarrie,
417 2015). With the re-emergence of the Central Iberian curve debate, several studies have re-
418 evaluated the well-documented structures from the Central Iberian Zone to constrain the origin
419 and kinematics of curvature. The majority of studies focused on the hinge zone of the curve in
420 the area surrounding Galicia Tras-os-Montes (e.g. Dias da Silva et al., 2014; Jacques et al.,
421 2018a), but some explored more outer-arc areas (e.g. Palero-Fernández et al., 2015; Gutiérrez-
422 Alonso et al., 2015). The following paragraphs synthesize the findings of new field, structural,
423 and geochronological analyses from around the hinge of the Central Iberian curve and its
424 surrounding regions. The reviewed studies identify several deformation events that are linked to
425 regional Variscan deformation phases (Fig. 2A).

426 1. An early generation of upright to overturned cylindrical folds with an associated axial
427 planar cleavage (C1). The C1 fold axes plunge variably from horizontal to nearly vertical
428 (e.g. Jacques et al., 2018a, 2018b). The original trend of the fold axes was parallel to the
429 orogen (e.g. Pastor-Galán et al., 2019b), however interference with younger deformation
430 events has created complicated geometries (e.g. Díez Fernández et al., 2013; Palero-
431 Fernández et al., 2015). The emplacement of the allochthonous units of Galicia Tras-os-
432 Montes zone (commonly referred as C2) is closely associated with development of C1
433 folds, but is restricted to shear zones located along the boundary between the latter and

the Central Iberian Zone. This phase includes orogen-parallel emplacement of the allochthonous Galicia Tras-Os Montes units and its associated thrusts (Fig. 2A). The non-coaxial nature of the emplacement of this allochthonous nappe produced folding interference and local vertical-axis rotations (Dias da Silva et al., *in press*). Prograde Barrovian metamorphism (known as M1) reached its pressure peak at the end of C2 (Rubio Pascual et al., 2013).

2. After C1 and C2, the resulting thickened crust gravitationally collapsed (Macaya et al., 1991; Escuder Viruete et al., 1994; Díaz-Balda et al., 1995; Díez-Montes, 2010). This gravitational collapse (phase E1) formed gneiss-dome core complexes between 330 and 317 Ma (e.g. Díez Fernández and Pereira, 2016) especially at the core of the Central Iberian curve (Fig. 2C; e.g. Martínez-Catalán, 2012). This phase formed large subhorizontal extensional detachments that exhumed to depths of the middle crust (e.g. Rubio-Pascual et al., 2013; Dias da Silva et al., *in press*). General decompression produced a Buchan-type metamorphic event (M2; e.g. Rubio-Pascual et al., 2016, Solís-Alulima et al., 2019) and widespread anatexitic melting (e.g. López-Moro et al., 2018; Pereira et al., 2018). E1 phase developed a fold system with sub-horizontal axes and a penetrative subhorizontal cleavage (e.g. Dias da Silva et al., *in press*). Mapped folding geometries indicate the deflection of C1 folds into overturned positions within the E1 deformation zones (e.g. Díez Fernández et al., 2013; Díez Fernández and Pereira, 2016; Pastor-Galán et al., 2019b). In addition to large-scale extensional deformation and Buchan metamorphism, E1 developed a regional dome-and-basin pattern, resulting in portions of the allochthonous terranes tectonically transported into basins (e.g. Días da Silva et al., *in press*).
3. The structures developed during C1-C2 compression and E1 extension, are re-folded by a younger shortening phase (C3; syn-Cantabrian Orocline). C3 formed upright open folds and conjugate sub-vertical shear zones (e.g. Gutiérrez-Alonso et al., 2015; Díez Fernández and Pereira, 2017; Dias da Silva et al., *in press*). C3 was coeval with regional retrograde metamorphism (M3) and with intrusion of mantle derived granitoids (Fig. 2C; e.g. Gutiérrez-Alonso et al., 2011a), surrounded by contact metamorphic aureoles (e.g. Yenes et al., 1999). The age of the C3 event ranges from 315 and 290 (e.g. Jacques et al., 2018a), concomitant with the formation of the Cantabrian Orocline (e.g. Pastor-Galán et al., 2015a). Ductile deformation, including folding with axial planar cleavage (e.g. Dias da Silva et al., 2014; Pastor-Galán et al., 2019b) and shear zones, occurred at the early stages of C3 (315-305 Ma; Gutiérrez-Alonso et al., 2015; Díez-Fernández and Pereira,

468 2017; Jacques et al., 2018b) followed by brittle deformation that formed cross-joint sets
469 and vein swarms with Sn-W mineralizations (Jacques et al., 2018a; 2018b). The
470 conjugated shear zones, some of them with hundreds of kilometers of displacement, had
471 activity during the period 315-305 based on direct Ar-Ar dating of the shear zones
472 (Gutiérrez-Alonso et al., 2015) and cross-cutting relationships with precisely dated
473 igneous rocks (Díez-Fernández and Pereira, 2017). Note that these shear zones show a
474 younger age with respect to the sinistral shear zones that bound the Ossa Morena and
475 South Portuguese zones (340-330 Ma; e.g. Dallmeyer et al., 1993). New studies in the
476 Central Iberian Zone have determined that several folds, previously interpreted as C1
477 (e.g. the Tamames-Marofa-Sátão synform) are C3 structures, possibly nucleated within
478 existing C1-C2 structures (e.g. Dias da Silva et al., 2017; Jacques et al., 2018b). The
479 remarkable continuity along the Central Iberian Zone of these folds (Fig. 2A), previously
480 interpreted as C1 (e.g. Díez-Balda et al., 1990; Abalos et al., 2002; Dias and Ribeiro,
481 1994; Dias et al., 2016), suggest the ubiquity and importance of this deformation phase.

- 482 4. Subsequent to C3 deformation, a brittle shortening event (C4) together with some late
483 extensional faults occurred across the region (E2; Fig. 2A; Dias and Ribeiro 1991; Dias
484 et al. 2003; Rubio Pascual et al., 2013; Arango et al., 2013; Fernández-Lozano et al.
485 2019; Dias da Silva et al., in press). E2 developed core complex-like structures that
486 further telescoped the M2 metamorphic isograds between the anatetic cores of gneiss
487 domes and the hanging wall units. This event also favoured sub-horizontal folding and
488 kink-band generation in the upper structural levels. Post-Variscan shortening structures
489 in Northern Iberia are characterized by a N-S compressive regime (C4) allowing the
490 formation of brittle NNE-SSW and NNW-SSE faults and associated sub-vertical and sub-
491 horizontal widespread kink-bands (e.g. Aller et al., 2020).

492 5.2 Paleomagnetism

493 Paleomagnetism investigates the record of the Earth's ancient magnetic field as it is
494 recorded in the rock record. Among other features, rocks can record the orientation of the
495 magnetic field at the time of magnetization (e.g. Tauxe, 2010). The recorded magnetic vector
496 can be geometrically defined by two components: inclination, which is a function of the
497 paleolatitude (being 90° at the poles and 0° at the equator) at the time of magnetization
498 acquisition; and declination, which is a measure of the horizontal angular difference between the
499 recorded magnetic direction and true north, thereby allowing for the quantification of any
500 vertical-axis rotations if a north reference direction is known for the region of interest at the time

501 of magnetization acquisition. Paleomagnetism is the best tool to quantify vertical-axis rotations
502 in orogens due to the independence of the magnetic field from ~~the~~ orogen deformation and
503 evolution (e.g. Butler, 1998).

504 Despite its uniqueness to study paleolatitudes and vertical-axis rotations,
505 paleomagnetism is not flawless. Paleomagnetic data can yield spurious rotations when the local
506 and regional structures are not properly studied and their geometries and kinematic histories not
507 adequately corrected for (e.g. Pueyo et al., 2016). In addition, the age of magnetization
508 acquisition is not necessarily equivalent to the age of the sampled rock. Remagnetizations are
509 ubiquitous, especially in orogens (Weil and van der Voo, 2002; Pueyo et al., 2007; Huang et al.,
510 2017). In remagnetized rocks, the primary magnetization is replaced or overprinted due to a set
511 of geologic processes acting alone or in concert - usually represented by a combination of
512 thermal or chemical reactions (Jackson, 1990). Nevertheless, remagnetizations can be useful
513 for interpreting deformation history if the relative timing of the overprint can be established and a
514 well-constrained reference direction for that age is known (e.g. Weil et al., 2001; Izquierdo-
515 Llavall et al. 2015; Calvín et al., 2017).

516 In addition to knowing the structural geology and the timing of magnetization of the
517 studied rocks, understanding and quantifying local and regional vertical-axis rotations require a
518 paleomagnetic reference pole for comparison. Permian and Mesozoic paleomagnetic studies in
519 Iberia indicate that Iberia was a relatively stable plate from at least Guadalupian times (ca. 270
520 Ma) to the opening of the Bay of Biscay in the Cretaceous (e.g. Gong et al., 2008; Vissers et al.,
521 2016). Weil et al. (2010) calculated the most modern Early Permian pole for stable Iberia, which
522 will be used herein as a reference for any vertical-axis rotation analysis (hereafter, eP pole or
523 eP component). Weil et al.'s Virtual Geomagnetic Pole (VGP) values are Plat = 43.9; Plong =
524 203.3 and $\alpha_{95} = 5.4$ and when transform into paleomagnetic components has a $\sim 0^\circ$ inclination
525 (equatorial) and declinations that range from 150° to 160° (from NW to SW respectively)
526 depending on where in Iberia you are referencing. In Fig. 6 (red arrows), a compilation of
527 declinations that form part of this composite pole and other eP components found in recent
528 studies are presented.

529 For the Central Iberian curve, the voluminous paleomagnetic database from the
530 Cantabrian Orocline can be used to partially constrain its kinematics (e.g. Weil et al., 2013). The
531 orocline test for the Cantabrian Orocline (fig. 4) quantifies the degree of differential vertical-axis
532 rotation of variously striking Variscan segments in northern Iberia. If the Central Iberian curve is
533 a product of vertical-axis rotation, paleomagnetic declinations would bend around the Central
534 Iberian curve opposite to that of the Cantabrian Orocline. With a well constrained orocline test,

535 as in the Cantabrian Orocline (Fig. 4), one can use the paleomagnetic strike-test correlation
536 slope to establish expected declinations for any along-strike portion of the orogen (Pastor-Galán
537 et al., 2017b).

538 Before the resurgence of the Central Iberian curve, the only available pre-Permian
539 paleomagnetic studies to the South and west of the Cantabrian Zone in the Iberian Massif were
540 focused on the Beja Gabbroic Massif, Portugal (Perroud et al., 1985) and the Almadén syncline
541 volcanics (Perroud et al., 1991; Pares & Van der Voo, 1992). The study in the Beja area showed
542 varied inclinations and declinations in the gabbros, and complex overprints elsewhere. Perroud
543 et al (1985) did not consider any structural correction for the results, assuming the gabbro was
544 undeformed. Recently, Dias da Silva et al. (2018) showed that the area underwent intense
545 deformation during the Carboniferous. Therefore interpretation of this dataset is complicated
546 without knowing the proper structural correction needed to restore the magnetization to its
547 palinspastic orientation.

548 Several articles with new paleomagnetic studies around the Central Iberian curve have
549 been published since 2015 (Fig. 5). In general, these studies have reported a pervasive late
550 Carboniferous (320 to 300 Ma) (re-)magnetization in sedimentary and igneous rocks (e.g.
551 Pastor-Galán et al., 2015a; 2017b; Fernández-Lozano et al. 2016), which is largely
552 penecontemporaneous to the intrusion of E1 extensional granites (López-Moro et al., 2018) and
553 C3 syn-orocline mantle derived granitoids (Fig. 2C; e.g. Gutiérrez-Alonso et al., 2011a). The
554 following section describes the magnetizations from oldest to youngest.

555 Pastor-Galán et al. (2016) sampled for paleomagnetic analyses both E1 extensional
556 granites (Fig. 2C; ~20 Ma; e.g. López-Moro et al., 2018) from the Tormes and Martinamor
557 domes, and C3 mantle derived granitoids in the Central System (Fig. 2C; 305-295 Ma; e.g.
558 Gutiérrez-Alonso et al., 2011a). Both sets of plutons are located around the Galicias-Tras-os-
559 Montes hinge of the Central Iberian curve (Fig. 6-5). The authors found an original component in
560 E1 granites supported by a positive reversal test in both domes (Fig. 7). The magnetization has
561 an inclination (Inc.) = 15° (paleolatitude (λ) = -7.6°) and declination (Dec.) = 81° (Fig. 7), which
562 imply a northward movement of 700 km and a ~70° CCW rotation with respect to the C3
563 granites that showed an eP component (Dec. ~ 150, Inc. ~ 0). Considering the positive reversal
564 test in E1 granites and the significant difference in inclinations with respect to C3 granitoids (eP
565 component), a magnetization age of older than 318 Ma was proposed (pre Kiaman superchron,
566 317 Ma - 267 Ma, e.g. Langereis et al., 2010), which was interpreted as a primary
567 magnetization. The 70° CCW Pennsylvanian rotation recorded in rocks from the Central Iberian
568 curve hinge zone is in agreement with the expected rotation of the southern limb of the

569 Cantabrian Orocline (Fig. 4; Weil et al., 2013).
570 At the putative outer arc of the Central Iberian curve, the Iberian Ranges (Fig. 2),
571 paleomagnetic and structural studies of Devonian and Permian rocks (Pastor-Galán et al.,
572 2018) revealed that the eP component from Permian rocks had rotated ~22° CW during the
573 Cenozoic (Fig. 8; cf. Pastor-Galán et al., 2018). The Permian and Mesozoic rocks from the
574 Iberian Ranges show a consistent ~22° CW rotation with respect to the Apparent Polar Wander
575 Path for Iberia (e.g. Pastor-Galán et al. 2018). This rotation likely happened during the Alpine
576 orogeny, in which the northern area of the Iberian Range underwent more shortening than the
577 southern part, resulting in a regional CW vertical-axis rotation (Izquierdo-Llavall et al., 2019).
578 After restoring the Cenozoic rotation, the Devonian rocks show a positive reversal and fold-test
579 with inclinations that are steeper than expected from the eP component (Dec. = 85.3°, Inc. =
580 12.7°, λ = -6.4). This component is statistically indistinguishable from that of the E1 granites and
581 the southern branch of the Cantabrian Zone, showing the same 70° CCW rotation from the time
582 they were remagnetized (estimated in 318 Ma) to the timing of the eP component (Fig. 8;
583 Pastor-Galán et al., 2018). Once Cenozoic rotation is corrected for, the structural and
584 paleomagnetic trends of the Iberian ranges become parallel to those in the southern limb of the
585 Cantabrian Orocline, ruling out a Variscan or older origin for the outer Central Iberian curve (Fig.
586 8).

587 The remaining paleomagnetic works published on Central and SW Iberia rocks all reveal
588 a ubiquitous late Carboniferous to Early Permian remagnetizations during the Kiaman
589 superchron (Fernández-Lozano et al., 2016; Pastor-Galán et al., 2015a; 2016; 2017b; Leite
590 Mendes, in press). The authors of these papers calculated the expected declination for each
591 site as if they were part of the Cantabrian Orocline (Fig. 9A). All localities where magnetizations
592 pre-date the formation of the Cantabrian Orocline show the same expected rotations as the
593 southern limb of the Cantabrian Orocline, regardless of their position within the Central Iberian
594 curve (to the hinge: Tormes and Martinamor domes, Iberian ranges; to the southern limb:
595 Almadén syncline and South Portuguese Zone). Other locations, especially limestones from the
596 Central Iberian Zone, have declinations and inclinations in between the primary 318 Ma
597 component of the E1 granites and the post-orocline eP component (Fig. 9B). Pastor-Galán et al.
598 (2015a; 2016) interpreted these results as being caused by a remagnetization that was acquired
599 during Cantabrian Orocline formation and therefore recorded intermediate steps between the
600 component of the E1 granites and eP. Those authors suggest that the large amount of syn-
601 orocline mantle derived granitoids that intruded the Central Iberian Zone (C3 granitoids)
602 triggered the hinterland remagnetization.

603 Finally, two previous studies identified an earlier magnetization in the Almadén syncline
604 region of the SE Central Iberian Zone (Perroud et al., 1991; Pares & Van der Voo, 1992).
605 However, Leite Mendes et al. (in press) argue that these studies are likely misinterpreted.
606 Perroud et al. (1991), applied a complicated structural correction restoring a putative plunge of
607 the regional structural axis to all sites, including those where the syncline axis does not plunge.
608 Leite Mendes et al. (in press) re-sampled the syncline where its axis is sub-horizontal and
609 obtained a negative fold test, implying that the magnetization is not primary as previously
610 interpreted. Their results, however, are similar in orientation to those components published
611 from previous studies prior to any structural correction (Perroud et al., 1991 and Parés and van
612 der Voo, 1992).

613 Two additional studies sampled Laurussian margin sequences that are today adjacent to
614 the Cantabrian Orocline region (Fig. 10). To the north, the SW area of Ireland preserves a Late
615 Paleozoic basin filled with Devonian red sandstone and Carboniferous limestone and siltstone,
616 which was sampled by Pastor-Galán et al. (2015a). To the south is the aforementioned results
617 from the South Portuguese Zone (Leite Mendes et al., in press). Both areas are interpreted to
618 have previously been part of the Laurussian continent, on the opposite side of the Rheic Ocean
619 suture at the time of Variscan collision (Fig. 10; e.g. Pastor-Galán et al., 2015b). In contrast, the
620 rest of Iberia was part of, or proximal to, Gondwana (e.g. Franke et al., 2017). These
621 Paleomagnetic results from the Laurussian margin suggest that the rotations involved in the
622 formation of the Cantabrian Orocline occurred along both sides of the Rheic suture proximal to
623 both its northern and southern limb (Fig. 10A and B). Pastor-Galán et al. (2015b) hypothesized
624 a so-called Greater Cantabrian Orocline that would have bent the entire Appalachian/Variscan
625 orogen around a vertical-axis, affecting at least the continental margins of both Gondwana and
626 Laurussia.

627 5.3 The implications of not being a secondary orocline

628 The most relevant new data regarding the kinematics of the Central Iberian curve is the
629 paleomagnetic study from the Iberian Ranges (Calvín et al., 2014; Pastor-Galán et al., 2018).
630 These results confirm that the present-day variation in trend of the tectonostratigraphic units,
631 generally attributed to Variscan tectonics (e.g. Weil et al., 2013; Shaw et al., 2012; 2014), is
632 likely a product of Cenozoic Alpine orogeny. Izquierdo-Llavall et al. (2019) confirmed that the
633 interpreted Alpine rotations correspond well with the amount of shortening reconstructed in
634 Meso-Cenozoic basins. The best preserved and most continuous outcrop in the Central
635 Iberian's outer arc is not a Variscan structure, casting doubt that Central Iberian curve's is

related to Variscan kinematics. The results are also a reminder that the regional effects of Alpine deformation are often underestimated, especially close to the major Iberian Alpine fronts: the Pyrenees, Iberian Ranges, and the Betics.

Overall, new paleomagnetic data from the Central Iberian curve and nearby areas reveal pervasive late Carboniferous remagnetizations and regional vertical-axis rotations of the same sense and magnitude to those expected for the southern arm of the Cantabrian Orocline. The new paleomagnetic data indicate that a post ~320 Ma formation for the Central Iberia curve due to vertical-axis rotations is not supported (Pastor-Galán et al., 2016). The distribution in space and time of paleomagnetic results discards the formation of the Central Iberian curve as a product of Variscan gravitational collapse (E1, ~330-317 Ma) or concomitant to the Cantabrian Orocline (C3). So far, no pre-E2 paleomagnetic component has been found, and consequently, paleomagnetic data cannot reject an early orogenic origin for the inner arc of the Central Iberian curve (C1-C2, older than 330 Ma).

From a structural geology point of view, the Central Iberian curve does not display the classic geometries and structural interference patterns as found in other established oroclines (i.e., those systems that involve differential vertical-axis rotations, e.g. Li et al., 2012; van der Boon et al., 2018; Meijers et al., 2017; Rezaeian et al., in press). The geometry and structural behaviour of oroclines should resemble, at the crustal-scale, a regional vertical-axis fold preserved in plan-view, either formed by buckling (e.g. Johnston et al., 2001) or bending (e.g. Cifelli et al., 2008) mechanisms. In oroclines, pre-existing structures tend to follow fold trends around the curvature (e.g. Rosenbaum, 2014; Li et al., 2018). In addition, orocline cores tend to preserve radial structures and shortening patterns in the inner arc and orocline parallel shear zones and extension structures in their outer arc (e.g. Ries and Shackleton, 1976; Eichelberger and McQuarrie, 2015), similar to what is observed in multilayer folds (e.g. Fossen, 2016).

The structural geometry of the Central Iberian curve lacks such patterns.

Paleomagnetism from the Iberian Ranges indicate that the Cantabrian and West Asturian Leonese zones do not follow the Central Iberian curve, instead they continue their NWW-SEE trend into the Mediterranean in what it is now the Betic chain (Rodríguez-Cañero et al., 2018; Jabaloy-Sánchez et al., 2018; van Hinsbergen et al., 2020). Structural trends in the Ossa Morena and the South Portuguese Zone do not show any change in along-strike structural trend that supports large-scale CW rotations (e.g. Pérez-Cáceres et al., 2015; Quesada et al., 2019), whereas existing paleomagnetic data from those zones (Leite Mendes et al., in press) support a model of CCW rotation associated with the broader southern arm of the Cantabrian Orocline. In the Central Iberian and Galicia Tras-os-Montes zones, the trend of curvature is irregular (see C1

670 fold patterns in Fig. 2A) and nowhere are the expected inner and outer arc-related structures
671 preserved (e.g. Dias da Silva et al. in press).

672 The curved shape of C1 fold axes in the Central Iberian zone is better explained by fold
673 interference patterns than vertical-axis rotations (e.g. Pastor-Galán et al., 2019b). Moreover, the
674 curved shape of the Galicia Tras-os-Montes allochthonous nappe, which was emplaced orogen-
675 parallel, shows no evidence of vertical-axis rotation related structures (Fig. 2A; Dias da Silva et
676 al., in press). Other authors describe the changes in trend around the Central Iberian curve
677 expressed by C1 folds (Fig. 2A) as the product of fold interference patterns (e.g. Gutiérrez-
678 Alonso, 2009; Palero-Fernández et al., 2015; Jacques et al., 2018b; Dias da Silva et al., in
679 press). Pastor-Galán et al. (2019b) showed that curved C1 folds in the Central Iberian Zone
680 around the Galicia Tras-os-Montes boundary (Fig. 2A) are coaxial with C3 folds after restoring
681 the effects of C2 and E1 deformation phases (Fig. 11A). Both C1 and C3 formed under similar
682 shortening directions. In the same area, Jacques et al. (2018b) found similar fold interference
683 patterns, in addition they described kinematic incompatibility with the expected CW rotations
684 that would have occurred if the Central Iberian curve was an orocline. In other areas of the
685 Central Iberian Zone, the curved shape of C1 folds has been described as an interference
686 between C1 structures and their reorientation caused by C3 shear zones (Fig. 2A; e.g. Palero-
687 Fernández et al., 2015; Dias et al., 2016), or alternatively the interference between C1, C3 and
688 the E2 structures (Fig. 2A; Gutiérrez-Alonso, 2009; Arango et al., 2013; Rubio Pascual et al.
689 2013).

690 Overall, new geometric and kinematic data favor the interpretation that the Central
691 Iberian curve is not a structure formed by differential vertical-axis rotation as was the Cantabrian
692 Orocline, but one formed as a consequence of several competing processes. It is clear from the
693 current data that a combination of several deformation events caused the orientation of
694 structures that today delineate the shape of the Central Iberia curve. These include: (1) the
695 northern part of the outer-arc is the product of an Alpine rigid block rotation instead of Variscan
696 differential vertical-axis rotation (Pastor-Galán et al., 2018); (2) the curvature of the Galicia Tras-
697 os-Montes allochthonous nappe reflects its original shape and could be defined as a primary
698 curve (see Weil and Sussman, 2004), since it was emplaced orogen-parallel and shows no sign
699 of vertical-axis rotations at any time (fig. 2A; Dias da Silva et al., in press); (3) Structural
700 analysis shows that fold interference patterns explain the geometry of the curved trends of
701 Central Iberian Zone's C1 folds (Pastor-Galán et al., 2019b), whose kinematics are incompatible
702 with the required CW rotations expected if the curve is an orocline (Jacques et al., 2018b).

703 **6 Get Back: Ideas flowing out and endless questions**

704 The pioneering works in the last decade that resurrected the idea of a Central Iberian
705 curve, speculated that both the Cantabrian and Central Iberian zones buckled together as
706 secondary oroclines (Fig. 12; Martínez-Catalán 2011; Shaw et al., 2012, 2014; Shaw and
707 Johnston, 2016; Carreras and Druguet, 2014). Later, Martínez Catalán et al. (2014) and Díez
708 Fernández and Pereira (2017) reformulated Martínez-Catalán's 2011 hypothesis and proposed
709 that the Central Iberian curve formed as an orocline between 315 and 305 Ma, and assigning
710 the Cantabrian Orocline a time frame between 305 and 295 Ma (Fig. 12). The proposed tectonic
711 mechanisms to support these early kinematic models are varied: (1) buckling of a ribbon
712 'Armorican' continent (Fig. 12A; Shaw et al., 2014; 2016); (2) buckling of a completely formed
713 Variscan orogen during a putative 'Pangea B' to 'Pangea A' transition in the late Carboniferous
714 (Fig. 12B; Carreras and Druguet, 2014; Martínez-Catalán et al., 2011); (3) indentation of
715 Laurussia into Gondwana during the early stages of collision (at present day SW Iberia, South
716 Portuguese Zone), producing first the Central Iberian curve as a mega drag-fold during
717 Carboniferous times and then slightly later the Cantabrian Orocline as a consequence of an
718 indentation process (fig. 12C; Simancas et al., 2013).

719 The reviewed data in sections 4 and 5 contradict the aforementioned hypotheses.
720 Paleomagnetism and structural patterns (section 5; Fig. 6-11) disagree with the necessary CW
721 rotations required to support a late Carboniferous orocline origin for the Central Iberian curve
722 (Models in Fig. 12A and B). In addition, the sense and magnitude of the vertical-axis rotations
723 observed in SW Iberia (Fig. 10) imply that the South Portuguese (Avalonian segment) and Ossa
724 Morena zones moved together with the southern limb of the Cantabrian Orocline during the
725 Pennsylvanian and Early Permian. This means that the South Portuguese Zone was already
726 parallel to the general trend of the Variscan orogen prior to Cantabrian Orocline formation,
727 implying the lack of a Laurussian rigid indenter into Gondwana (e.g. Simancas et al., 2013). This
728 discrepancy leaves orogen-parallel terrane transport as a possible explanation to the kinematics
729 observed in Ossa Morena and South-Portuguese Zones (e.g. Pérez-Cáceres et al., 2016). At
730 the same time, paleomagnetism from SW Iberia backs the hypothesis of a Greater Cantabrian
731 Orocline extended into both Gondwana and Laurussia in its northern and southern limbs (Fig.
732 10; Pastor-Galán et al., 2015b).

733 In spite of the kinematic constraints and structural patterns, which do not support a
734 vertical-axis origin for the Central Iberian curve in Late Carboniferous time, other geometric
735 constraints remain challenging. The curved shape of the aeromagnetic and gravity anomalies of
736 Iberia are real (Fig. 5). These striking patterns could be due to Variscan-Alpine structural

737 interference, for example the previous example from the Iberian Ranges, but currently there is
738 not enough data to rigorously test this hypothesis. Shaw et al. (2012) supported their hypothesis
739 of a secondary orocline by assuming that paleocurrents were parallel through Iberia during
740 Ordovician times. However, some of the observed deflections in the paleocurrents studied by
741 Shaw et al. (2012; see Fig. 3) are also explained by Alpine vertical-axis rotations (the case of
742 the Iberian ranges) and fold interference patterns (SE of the Central Iberian Zone). Others
743 (Central and SW of the Central Iberian Zone) may be explained by a local response to basin
744 architecture (Fig. 3), where paleo-flow directions would trend toward the deepest basin
745 throughs. The Ordovician basin architecture of Iberia allows for opposite directed paleocurrents
746 from both sides of such throughs (Fig. 3). However, the Early Paleozoic basin architecture in
747 Iberia and their local deformation events require further research (Sánchez-García et al., 2019).

748 Although kinematic evidence is still scarce for the earliest Variscan movements, we
749 argue that pre-orogenic physiographic features, such as the opening of a marginal restricted
750 ocean between Gondwana and its distal platform at 395 Ma (Fig. 13A; Pin et al., 2002;
751 Gutiérrez-Alonso et al., 2008b; Arenas et al., 2016) explains the rounded shape of the Galicia
752 ~~tras-os-Montes~~ curve as a primary ~~arc~~. During the collision, the latter irregularity would cause
753 the orogen-parallel emplacement of allochthonous nappe (Fig. 13B; Dias da Silva et al., in
754 press) and the left-lateral movements of the Ossa Morena and South Portuguese Zones in SW
755 Iberia (Fig 13A, B, C; Quesada, 2019). During the late Carboniferous, possibly due to a plate
756 reorganization during the final amalgamation of Pangea (Fig. 13D; e.g. Gutiérrez-Alonso et al.,
757 2008a; Pastor-Galán et al., 2015a), the far-field stress-field likely changed and buckled the
758 entire orogen around a vertical axis (Gutiérrez-Alonso et al., 2004), including both the
759 Gondwana and Laurussia margins (Fig. 13E; Pastor-Galán et al., 2015b).

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1464 **Captions**

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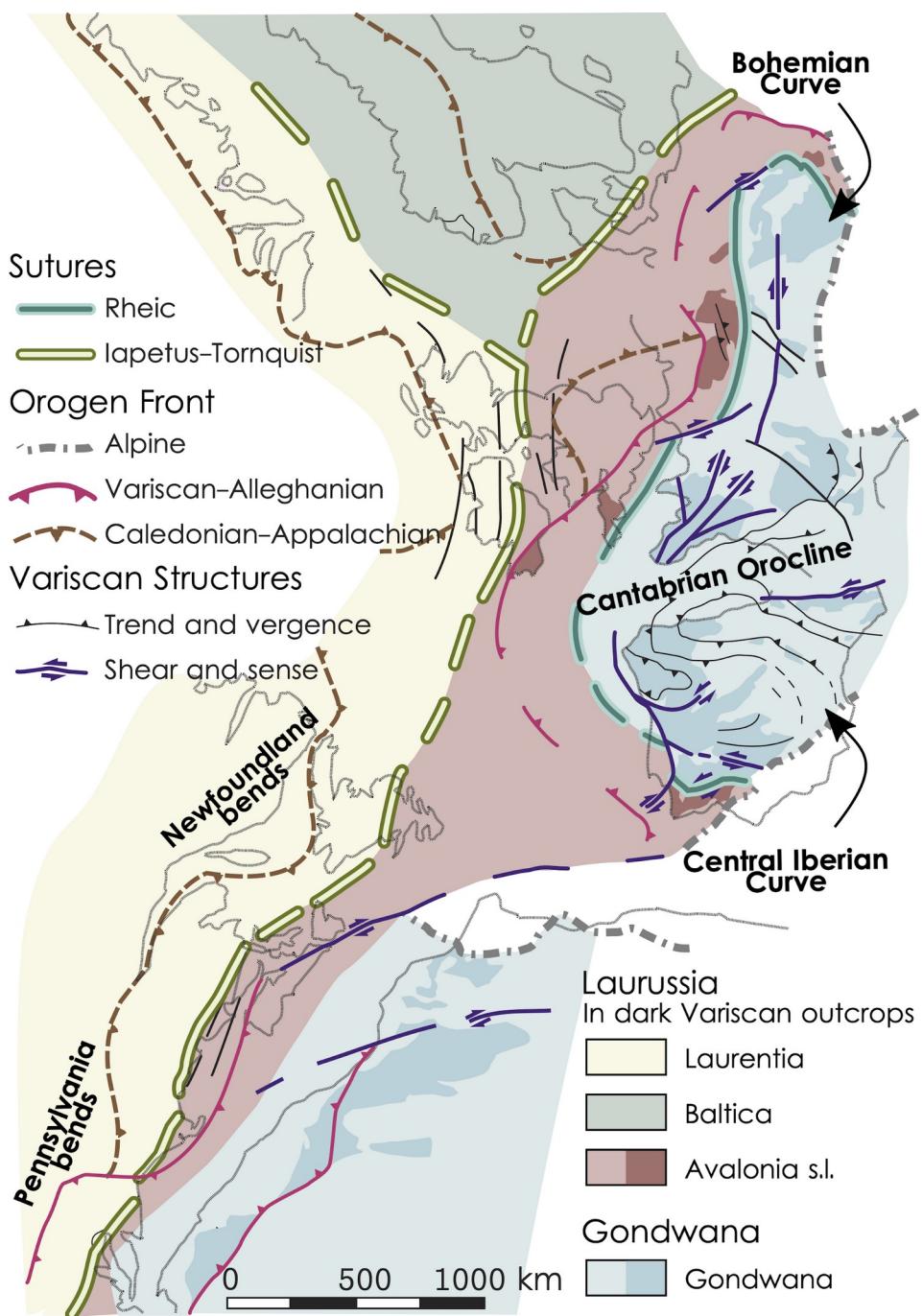
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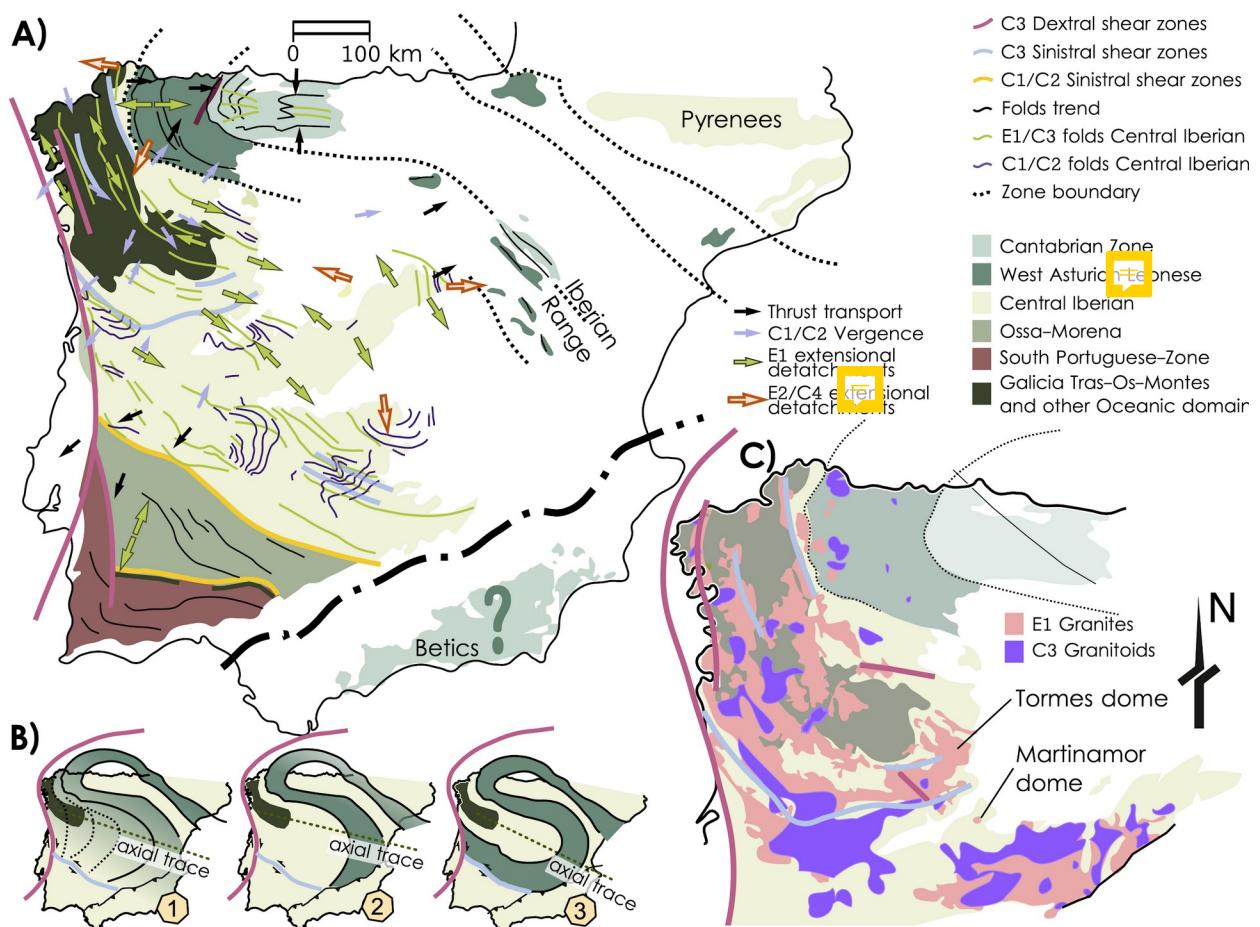
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1551 Fig. 13 Preliminary kinematic proposal for the Iberian Variscides. A) Pre-colisional stage
1552 after the opening of the Galicia Tras-os-Montes restricted seaway (e.g. Pin et al., 2002;
1553 Gutiérrez-Alonso et al., 2008a; Arenas et al., 2016). The irregular shape of the margin and the
1554 younging westwards deformation front (e.g. Daleyer et al., 1997) resulted in tectonic escape
1555 towards the still open Rheic Ocean (e.g. Braid et al., 2011; Murphy et al., 2016). B) After closure
1556 of the Rheic ocean, C1 and C2 structures formed. The Galicia Tras-os-Montes was emplaced
1557 orogen-parallel (e.g. Martínez-Catalán et al., 1990; Dias da Silva et al., in press), preserving the
1558 shape of the seaway, i.e. a primary arc. C) The gravitational collapse of the orogen produced
1559 widespread anatexis and folding interference in the hinterland and the emplacement of the
1560 foreland fold-and-thrust belt. D) At Pennsylvanian times a change in the far-field stress buckled
1561 the Variscan belt around a vertical axis (see Gutiérrez-Alonso et al., 2008; Weil et al., 2013;
1562 Pastor-Galán et al., 2015a for details), creating new interference patterns and a lithospheric
1563 scale response (see Gutiérrez-Alonso et al., 2004, 2011a; Pastor-Galán et al., 2012a). E) When
1564 the orocline became too tight to keep rotating, new cross-cutting brittle structures (C4) formed
1565 and minor extensional collapse (E2) occurred (e.g. Fernández-Lozano et al., 2019; Dias da
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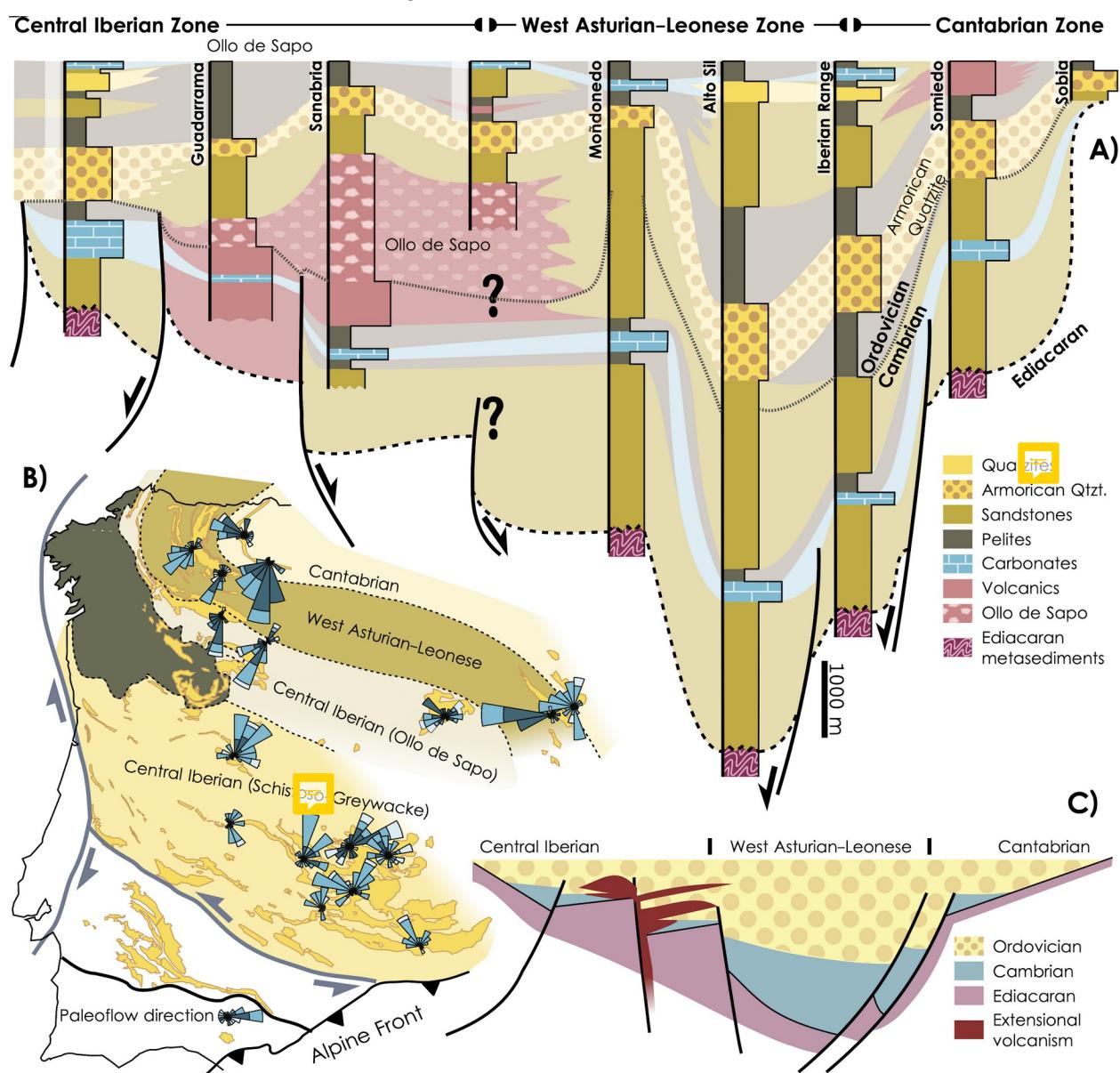
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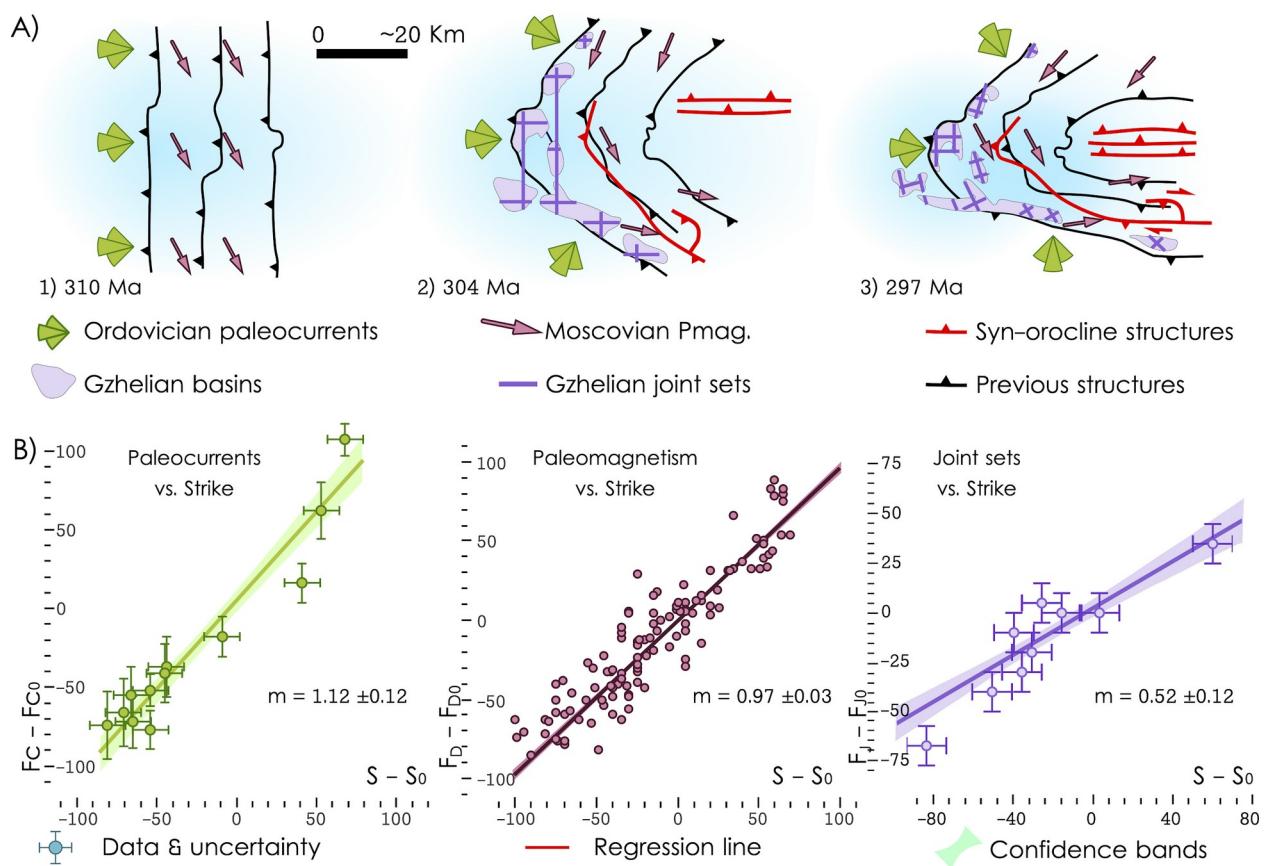
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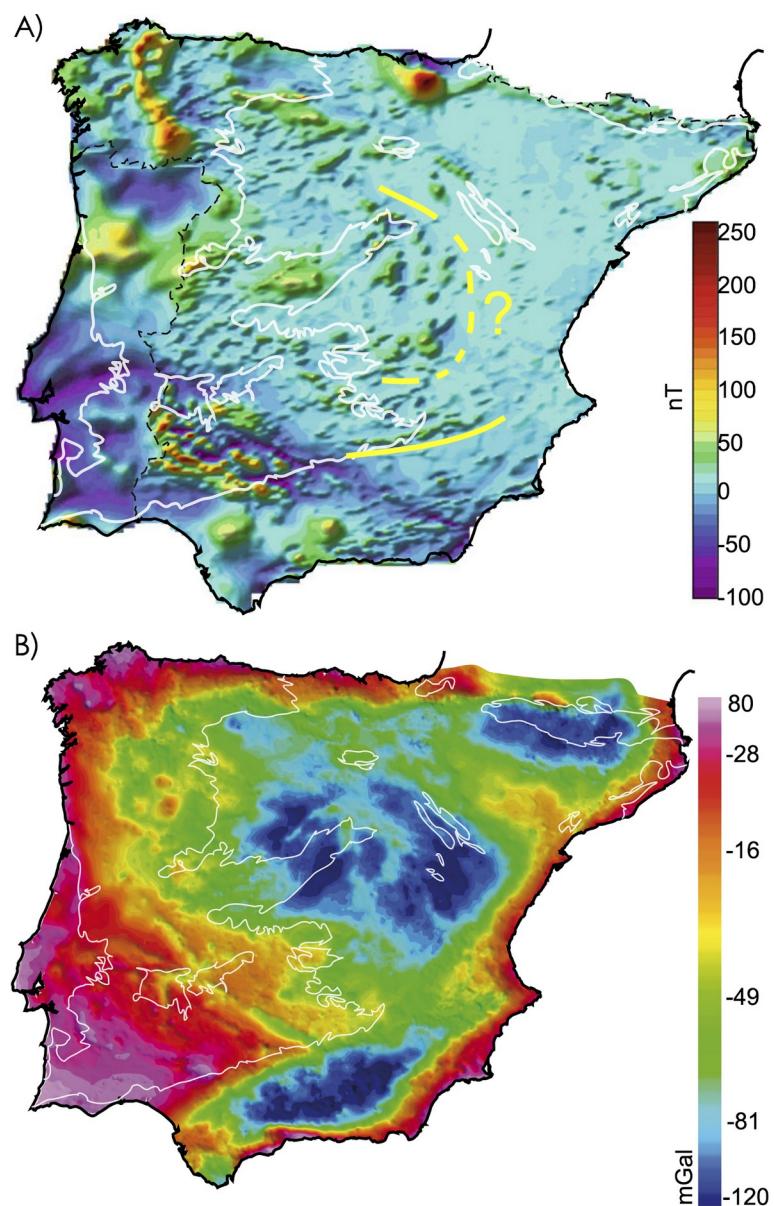
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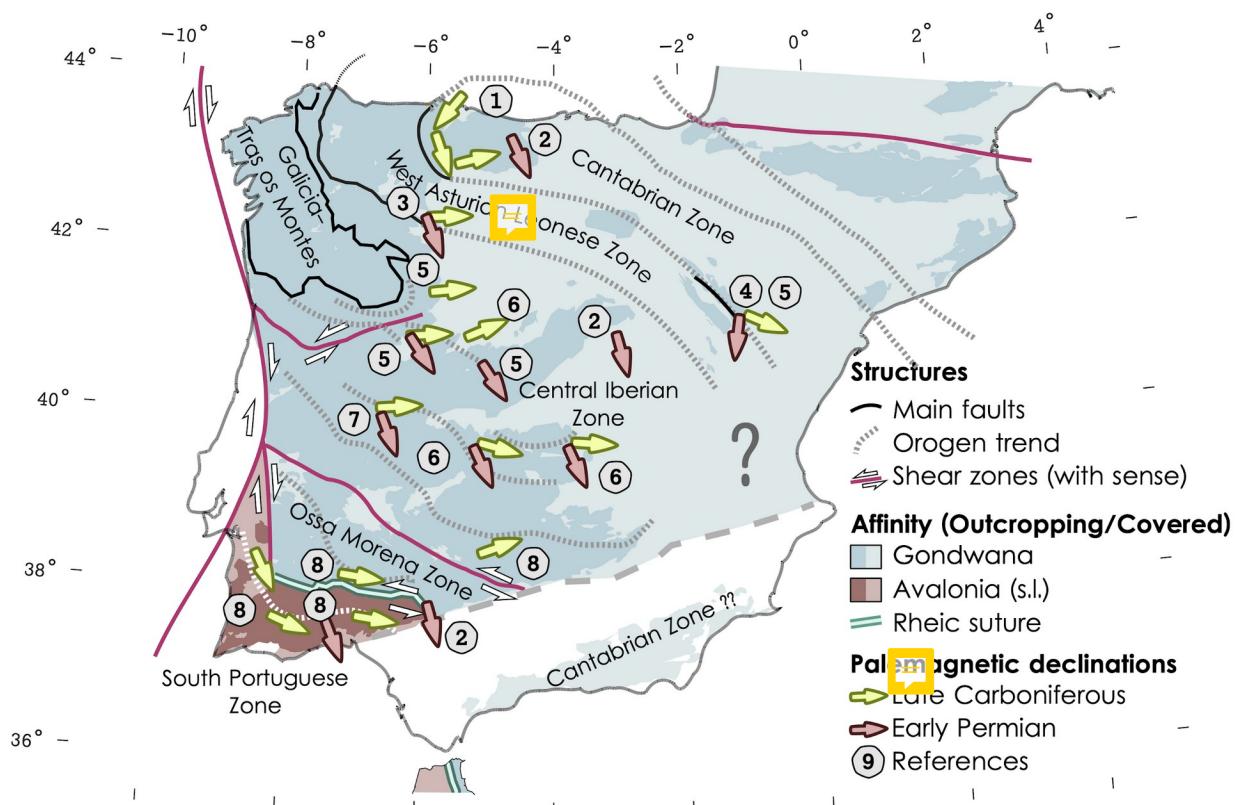
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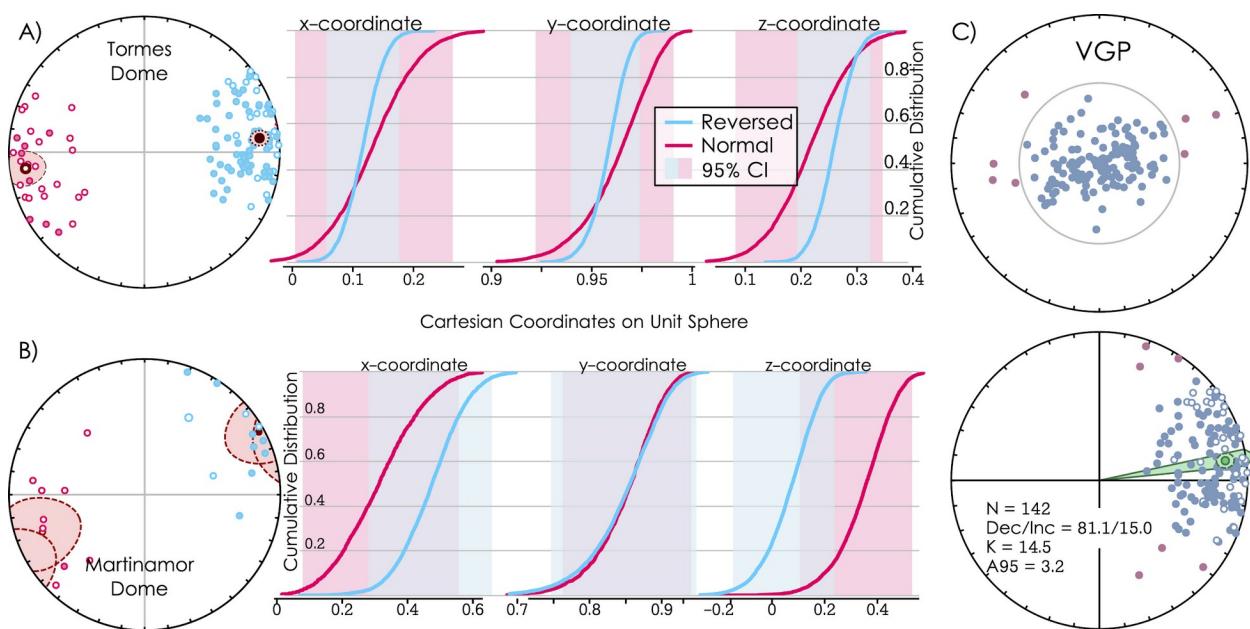
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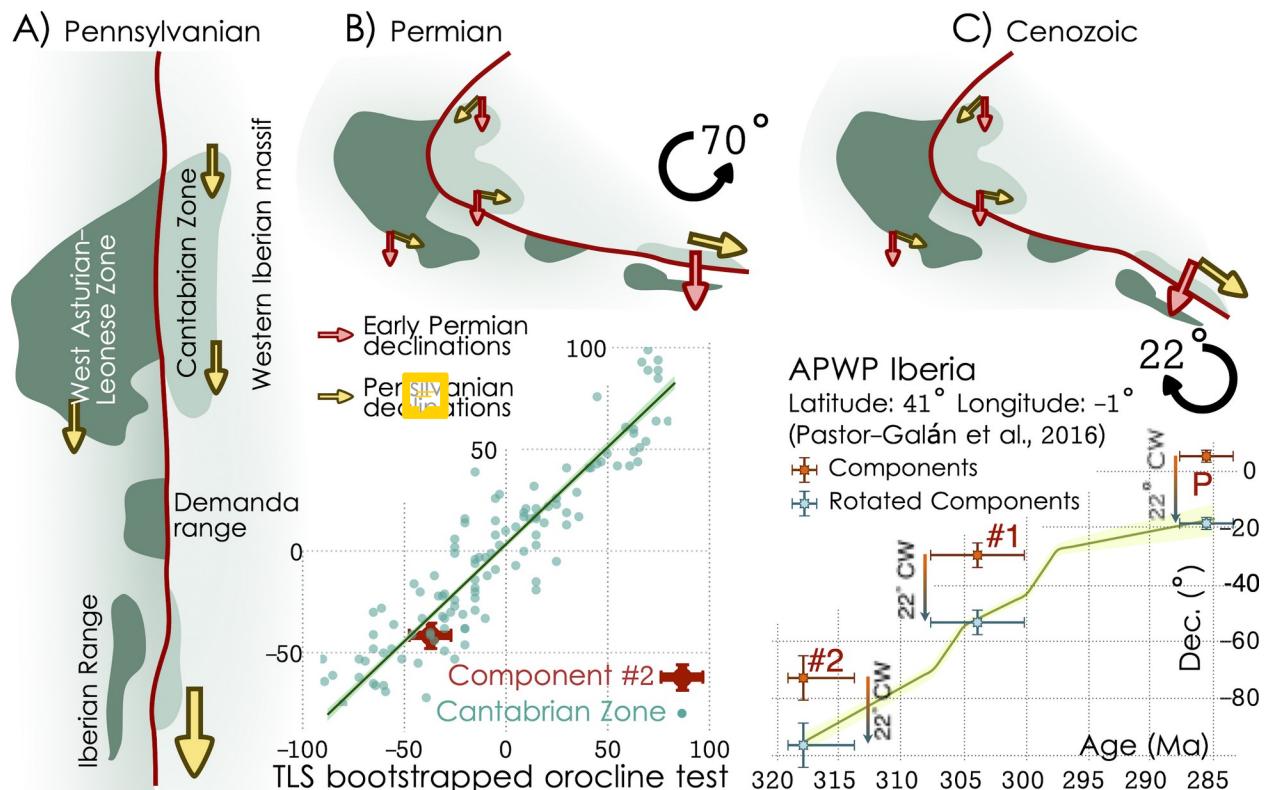
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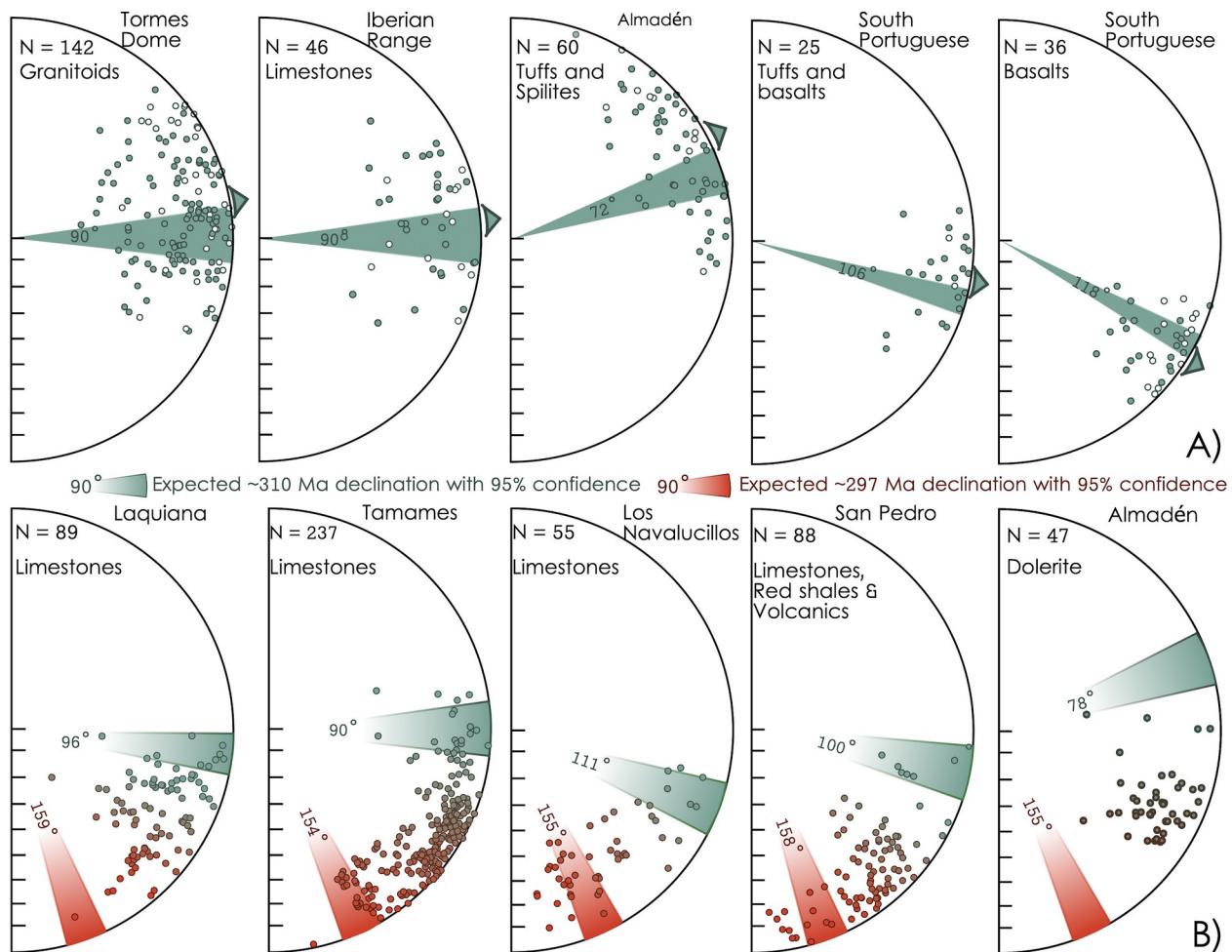
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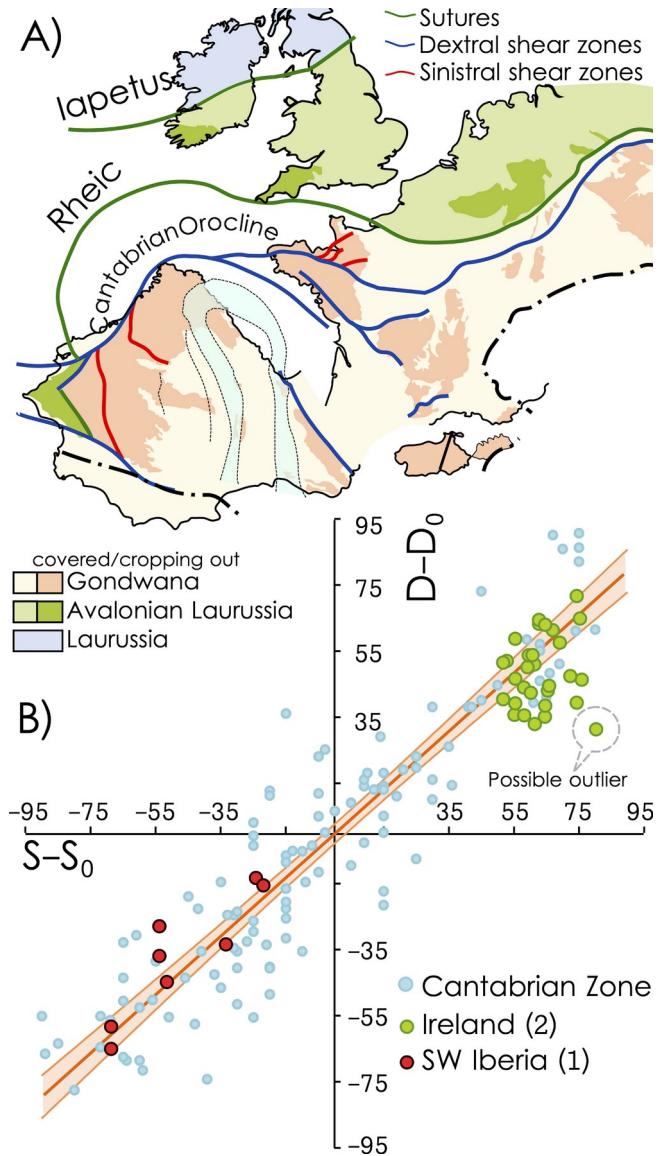
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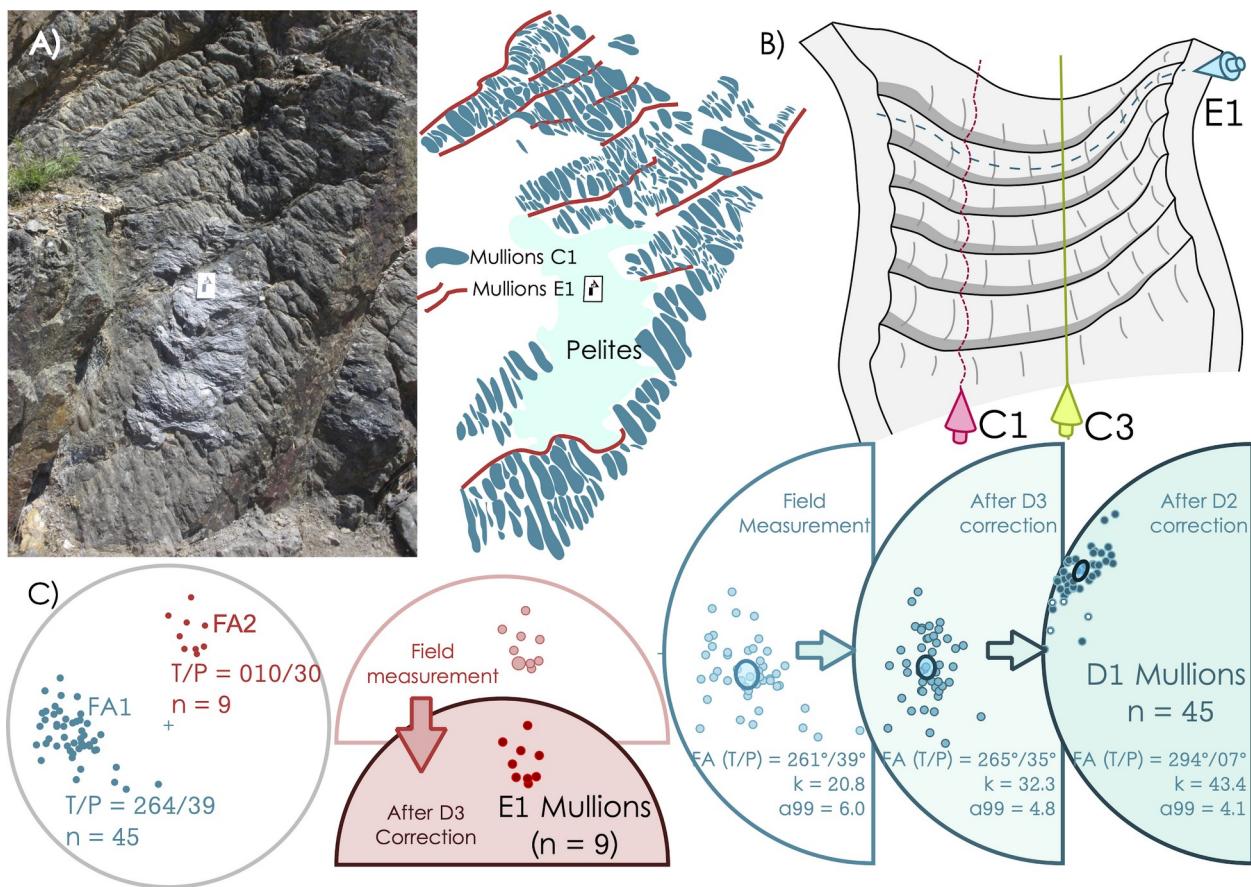
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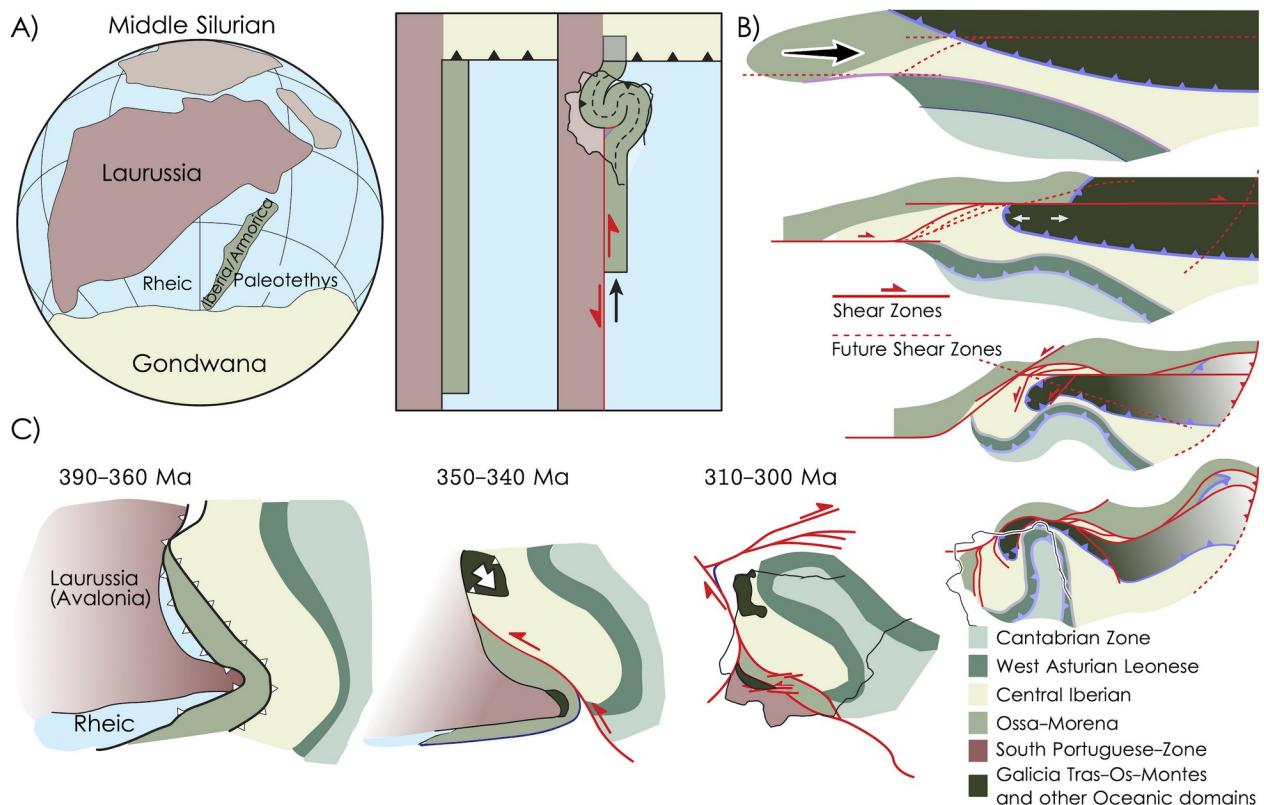
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1648 Fig. 12 Pioneering hypothesis for the Central Iberian curve. Note that none of them fulfill
 1649 the most recent geometric and kinematic criteria. A) Simplified ribbon continent model after
 1650 Johnston et al. (2013) and Shaw and Johnston (2016). B) Dextral mega-shear model from
 1651 Martínez-Catalán (2011). C) Kinematic model with indentation and left-lateral shearing after
 1652 Simancas et al. (2013)



1653 Fig. 13 Preliminary kinematic proposal for the Iberian Variscides. A) Pre-colisional stage
1654 after the opening of the Galicia Tras-os-Montes restricted seaway (e.g. Pin et al., 2002;
1655 Gutiérrez-Alonso et al., 2008a; Arenas et al., 2016). The irregular shape of the margin and the
1656 younging westwards deformation front (e.g. Daleyer et al., 1997) resulted in tectonic escape
1657 towards the still open Rheic Ocean (e.g. Braid et al., 2011; Murphy et al., 2016). B) After closure
1658 of the Rheic ocean, C1 and C2 structures formed. The Galicia Tras-os-Montes was emplaced
1659 orogen parallel (e.g. Martínez-Catalán et al., 1990; Dias da Silva et al., in press), preserving the
1660 shape of the seaway, i.e. a primary arc. C) The gravitational collapse of the orogen produced
1661 widespread anatexis and folding interference in the hinterland and the emplacement of the
1662 foreland fold-and-thrust belt. D) At Pennsylvanian times a change in the far-field stress buckled
1663 the Variscan belt around a vertical axis (see Gutiérrez-Alonso et al., 2008; Weil et al., 2013;
1664 Pastor-Galán et al., 2015a for details), creating new interference patterns and a lithospheric
1665 scale response (see Gutiérrez-Alonso et al., 2004, 2011a; Pastor-Galán et al., 2012a). E) When
1666 the orocline became too tight to keep rotating, new cross-cutting brittle structures (C4) formed
1667 and minor extensional collapse (E2) occurred (e.g. Fernández-Lozano et al., 2019; Dias da
1668 Silva et al., in press).

