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## The Ordovician from France and neighbouring areas of Belgium and Germany

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1 The Ordovician from France and neighbouring areas of Belgium and  
2 Germany

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25

26 Abbreviated title: Ordovician of France and neighbouring areas

27

28 **Abstract:** The Ordovician successions of France and neighbouring areas of Belgium and  
29 Germany are reviewed and correlated based on international chronostratigraphic and regional  
30 biostratigraphic charts. The same three megasequences related to the rift, drift, and docking of  
31 Avalonia with Baltica can be tracked in Belgium and neighbouring areas (Brabant Massif and  
32 Ardenne inliers), western (Rhenish Massif) and northeastern Germany (Rügen). The  
33 remaining investigated areas were part of Gondwana in the Ordovician. The Armorican  
34 Massif shares with the Iberian Peninsula a Furongian–Early Ordovician gap (Toledanian or  
35 Norman gap), and a continuous Mid–Late Ordovician shelf sedimentation. The Occitan  
36 Domain (Montagne Noire and Mouthoumet massifs), eastern Pyrenees and northwestern  
37 Corsica share with southwestern Sardinia continuous shelf sedimentation in the Early  
38 Ordovician, and a Mid Ordovician "Sardic gap". In the Ordovician, the Maures Massif  
39 probably belonged to the same Sardo-Occitan domain. The Vosges and Schwarzwald massifs  
40 display comparable, poorly preserved Ordovician successions, suggesting affinities with the  
41 Teplá-Barrandian and/or Moldanubian zones of Central Europe.

42

43

44 In western Europe, Ordovician sedimentary rocks generally occur as scattered, disconnected  
45 massifs or inliers surrounded and/or hidden by Mesozoic and/or Cenozoic cover.  
46 Reconstructing their original layout, stratigraphy and palaeogeographic relationships with  
47 other regions has been a major challenge during the last decades. All Ordovician strata  
48 occurring in present-day Belgium, France and Germany have undergone a long and complex  
49 journey. They were all deposited on a Cadomian or Cambrian basement with clear  
50 Gondwanan affinities. In the Furongian–Early Ordovician, the onset and diachronic

51 propagation, from present-day southwest to northeast, of a major rift along the western  
52 margin of Gondwana gave rise to the Rheic Ocean (e.g. Linnemann *et al.* 2008; von Raumer  
53 and Stampfli 2008; Cocks and Fortey 2009; Stampfli *et al.* 2011). The opening of the Rheic  
54 Ocean resulted in the rifting of Avalonia, as an independent microcontinent, away from  
55 Gondwana (e.g. Prigmore *et al.* 1997; Verniers *et al.* 2002; Cocks and Fortey 2009; Servais  
56 and Sintubin 2009; Domeier 2016). In the Ordovician, several areas included in this chapter  
57 were thus belonging to the eastern part of Avalonia: the Brabant Massif and Ardenne inliers  
58 in Belgium, the Rhenish Massif in western Germany, and Rügen Island, in northeastern  
59 Germany (Fig. 1). In contrast, the Armorican Massif, the Occitan Domain (Montagne Noire  
60 and Mouthoumet massifs), eastern Pyrenees, the Maures Massif, Corsica, and the Vosges (in  
61 France) and Schwarzwald (in western Germany) were located on the opposite (southern) side  
62 of the Rheic ocean, still forming part of Gondwana (Fig. 1). Avalonia drifted northwards,  
63 away from Gondwana, before docking with Baltica in the Late Ordovician (closure of the  
64 Tornquist Ocean; TS in Fig. 1), and slightly later with Laurentia, in the Silurian (closure of  
65 the Iapetus Ocean; IS in Fig. 1), resulting in the Caledonian orogeny (e.g. Verniers *et al.*  
66 2002; Cocks and Fortey 2009).

67 In the late Palaeozoic, the closure of the Rheic Ocean led to the formation of an extensive  
68 orogenic belt extending from the Appalachian and Ouachita mountains (to the West) to the  
69 Caucasus (to the East). This major continental collision or "Variscan orogeny" between  
70 Laurentia–Avalonia–Baltica (to the North) and Gondwana (to the South) was particularly  
71 complex, because of the occurrence of numerous terranes, related to back-arc basins and  
72 narrow oceanic domains on both sides of the closing Rheic Ocean (Pharaoh 1999; Matte  
73 2001; Ballèvre *et al.* 2009; Faure *et al.* 2009; Guillot and Ménot 2009; Stampfli *et al.* 2011;  
74 von Raumer *et al.* 2013). A consequence of the Variscan "collage" was to bring together, on  
75 both sides of the Rheic suture (RS on Fig. 1), most pieces of the jigsaw forming present-day



76 Europe, and extending from the Iberian Massif (see Gutiérrez-Marco *et al.* 2002) to the Alps  
77 (Ferretti *et al.* this volume) and the Bohemian Massif (Kraft *et al.* this volume). With the  
78 exception of the Brabant Massif, all Ordovician rocks now exposed in western Europe were  
79 more or less affected by the Variscan orogeny. For example, in some areas (e.g. Maures  
80 Massif, central zones of the Vosges and Schwarzwald), they endured medium to high-grade  
81 metamorphism (Faure *et al.* 2014; Skrzypek *et al.* 2014; Gerbault *et al.* 2018). Even in  
82 regions exposing Ordovician rocks with very low to low-grade metamorphism, the strata were  
83 more or less affected by Variscan tectonics, involving folds and/or displacements along  
84 several tens of kilometers along shear zones or thrusts (e.g. Ballèvre *et al.* 2009; Pouclet *et al.*  
85 2017; Herbosch *et al.* 2020). Moreover, in some areas (e.g. Alps, eastern Pyrenees), Variscan  
86 structures were reactivated and/or overprinted during the Alpine orogeny, thus making their  
87 interpretation even more difficult.

88 During the last 30 years, the major advances in biostratigraphy (relying mostly on  
89 acritarchs, chitinozoans and graptolites, less frequently on brachiopods, conodonts and  
90 trilobites) have greatly contributed to the understanding of the complex pre-Variscan history  
91 of western Europe, leading to a wide (and sometimes contradictory) set of palaeogeographic  
92 scenarios and regional syntheses, as e.g. in the Armorican Massif (Le Corre *et al.* 1991; Paris  
93 and Robardet 1994; Robardet *et al.* 1994a; Ballèvre *et al.* 2009), Belgium (Verniers *et al.*  
94 2002; Linnemann *et al.* 2012; Herbosch *et al.* 2020), Corsica (Faure *et al.* 2014; Avigad *et al.*  
95 2018), the Ebbe inlier (e.g. Maletz 2000; Koch *et al.* 2014), the Maures Massif (Bellot 2005),  
96 the Occitan Domain (Dégardin *et al.* 1995; Laumonier *et al.* 1995; Pouclet *et al.* 2017), or the  
97 Vosges (Skrzypek *et al.* 2014). A few more integrative reviews have been produced, and  
98 include syntheses on the Ordovician of France (Paris *et al.* 1999; Nardin *et al.* 2014), Belgium  
99 and France (Robardet *et al.* 1994b), or Belgium and Germany (Verniers *et al.* 2002; Servais *et*  
100 *al.* 2008). The aim of the chapter is thus to provide an updated overview of the Ordovician of

101 Belgium, France, and both western and northeastern Germany through a small number of  
102 selected regions, and to correlate their sedimentary successions with both global subdivisions  
103 (series and stages) of the International Chronostratigraphic Chart and the biostratigraphic time  
104 scales elaborated by Goldman *et al.* (2020) for the Ordovician.

105

## 106 **The Ordovician of Belgium and neighbouring areas (by Herbosch, Verniers** 107 **and Servais)**

108

### 109 *General framework*

110

111 Ordovician strata exposed in the Caledonian basement of Belgium and neighbouring areas  
112 include two major units: those formed by the Brabant Massif (BM), the Brabant  
113 Parautochthon and the Haine-Sambre-Meuse Overturned Thrust Sheets (HSM-OTS) to the  
114 North, and the Ardenne Allochthon to the South (Fig. 2). The Midi-Eifel Thrust Fault  
115 displaced the Ardenne Allochthon several tens of km to the North (Hance *et al.* 1999; Mansy  
116 and Lacquement 2002) during the Variscan orogeny at the end of the Carboniferous. The  
117 Ordovician crops out also in the Condroz Inlier, a narrow strip along the Midi-Eifel Fault  
118 (Fig. 2). In the Ardenne Allochthon, the basement is exposed in four inliers: Stavelot-Venn,  
119 Rocroi, Givonne and Serpont. This Caledonian basement belongs to the Avalonia  
120 microcontinent (Fig. 1; Cocks and Torsvik 2002; Verniers *et al.* 2002). Overall, the  
121 sedimentation during the Ordovician is terrigenous and deposited in a deep marine anoxic  
122 environment, which explains the scarcity of benthic macrofossils. As a result it is only since  
123 the 1970s that researches with graptolites, acritarchs and chitinozoans have allowed the  
124 establishment of a reliable bio- and chronostratigraphy.

125

## 126 *The Brabant Massif*

127

128 The Brabant Massif (BM) consists of a largely concealed, NW-SE trending, low-grade  
129 metamorphic slate belt developed during the early Palaeozoic, in the sub-surface of northern  
130 Belgium (Verniers *et al.* 2002; Debacker *et al.* 2005). Along its southern edge, river incisions  
131 provide narrow outcrop areas through the cover (Fig. 2). To the South, the Brabant  
132 Parautochthon and/or the HSMS-OTS (Belanger *et al.* 2012) are overthrust by the Ardenne  
133 Allochthon along the Midi-Eifel Fault, which represents the Variscan Thrust Front (Fig. 2).  
134 The BM consists of a very thick siliciclastic sequence, ranging from the lower Cambrian in  
135 the core to the upper Silurian along the rims. The sedimentary record is continuous, except for  
136 a hiatus from the middle Tremadocian to the Dapingian (Fig. 3; Herbosch *et al.* 2008;  
137 Herbosch and Verniers 2014). The total thickness exceeds 13 km, with less than 1.8 km for  
138 the Ordovician (Linnemann *et al.* 2012, fig. 7).

139 The three sedimentary megasequences bounded by basin-wide unconformities recognized  
140 in the Welsh basin by Woodcock (1990) have also been recognized in the BM and the entire  
141 basement (Fig. 3; Woodcock 1991; Vanguetaine 1992; Verniers *et al.* 2002; Herbosch *et al.*  
142 2020). The Ordovician of the BM begins with the lower Tremadocian black slates of the  
143 uppermost part of the Mousty Formation dated by the graptolite *Rhabdinopora praeparabola*  
144 (Wang and Servais 2015). This formation passes gradually to the Chevlipont Formation  
145 formed by wavy bedded grey siltstones. It is dated by the graptolites *Rhabdinopora parabola*  
146 and *R. anglica* (Wang and Servais 2015) from about the lower half of the Tremadocian, i.e.  
147 from 485.5 to 482.7 Ma (Goldmann *et al.* 2020). Then, a stratigraphic hiatus of about 13 myr  
148 precedes the deposition of the Abbaye de Villers Formation (Fig. 3). It is interpreted as linked  
149 to the drifting of the Avalonia microplate from Gondwana and the opening of the Rheic  
150 Ocean (Cocks and Torsvik 2002; Verniers *et al.* 2002; Linnemann *et al.* 2012).

151 Megasequence 2 begins with the Abbaye de Villers and Tribotte formations consisting of  
152 silty slates (Herbosch and Verniers 2014). They are correlated with the upper Dapingian to  
153 the lower Darriwilian by acritarchs, which belong to the *F. hamata*-*S. rarrigulata* Biozone  
154 (Vanguetaine and Wauthoz 2011). This acritarch biozone can be correlated with the *I.*  
155 *gibberulus* and *A. cucullus* graptolite biozones (Cooper and Molyneux 1990). The overlying  
156 Rigenée Formation, formed by dark silty slate, records a regional transgression (Paris *et al.*  
157 2007) and is dated by graptolites (*D. artus* and *D. murchinsoni*; Maletz and Servais 1998) and  
158 acritarchs (*F. sartbernardensis*, *A. virgata*, *F. hamulata*; Servais 1991, 1993) to most of the  
159 Darriwilian (Herbosch and Verniers 2014, fig. 3). The Ittre Formation is formed by an  
160 alternation of fine sandstones, siltstones and slates interpreted as distal turbidites (Servais  
161 1991). It was dated by graptolites (*N. gracilis* and *M. foliaceus*; Maletz and Servais 1998) and  
162 chitinozoans (*S. cervicornis* Biozone; Vanmeirhaeghe 2006) as Sandbian. The Bornival  
163 Formation essentially formed by mudstones was correlated by chitinozoans (no index species)  
164 with the lowest Katian (Vanmeirhaeghe 2006). The Cimetière de Grand-Manil Formation was  
165 not biostratigraphically dated. The Huet Formation is formed by greenish slates and siltstones  
166 with shelly facies level (e.g. brachiopods, bryozoans, corals and crinoids; Mailleux 1926)  
167 interpreted as distal tempestites. It was dated by chitinozoans: *T. bergstroemi* (Van Grootel *et*  
168 *al.* 1997) and *F. spinifera* biozones (Vanmeirhaeghe 2006) and correlated with the middle  
169 Katian (Herbosch and Verniers 2014, fig. 3). The Fauquez Formation includes graptolitic  
170 black slates interpreted as very distal turbidites deposited in a deep marine environment. It  
171 was correlated by graptolites (*P. linearis*; Maletz and Servais 1998) and chitinozoans (*F.*  
172 *spinifera*; Vanmeirhaeghe, 2006) with the middle Katian.

173 Megasequence 3 begins with the Madot Formation showing many volcanic and volcano-  
174 sedimentary rocks interstratified with siltstones rich in shelly debris (Verniers *et al.* 2005)  
175 deposited in a shallow shelf. The transition from the Fauquez to the Madot Formation marks

176 an abrupt and important change in bathymetry. The shallowing could be linked to the docking  
177 of Avalonia with Baltica (Cocks and Torsvik 2005; Torsvik and Cocks 2011; Linnemann *et*  
178 *al.* 2012) and was correlated by the chitinozoans *T. bergstroemi* (Vanmeirhaeghe *et al.* 2005)  
179 and *C. rugata* (Mortier 2014) with the upper half of the Katian (Fig. 3). The lower Gouteux  
180 Member of the Brutia Formation includes grey bioturbated slates and was assigned by the  
181 chitinozoans *G. gamachiana* (Samuelsson and Verniers 2000) and *S. oulebsiri* (Mortier 2014)  
182 to the uppermost Katian – Hirnantian and extends most probably even into the Silurian.

183

#### 184 *The Condroz Inlier*

185

186 The Condroz Inlier (CI) is a narrow strip about 65 km long and 0.5 to 4 km wide composed of  
187 Ordovician and Silurian siliciclastic rocks, emplaced mostly as a series of tectonic blocks  
188 along the Midi-Eifel Fault Zone in the HSM-OTS (Fig. 2). The inlier is flanked tectonically to  
189 the North by Middle Devonian rocks and to the South by Lower Devonian rocks  
190 (Vanmeirhaeghe 2006). Given its position within the strongly faulted Variscan front zone and  
191 with a very poor degree of exposure, the stratigraphy of the CI is extremely difficult to  
192 establish and not fully resolved. In the main Central CI part, the sedimentary record extends  
193 from the base of the Ordovician to the upper part of the Silurian with an approximative  
194 thickness of over 1.1 km for the Ordovician (Vanmeirhaeghe 2006).

195 Megasequence 1 is reduced to the Chevlipont Formation observed in the Wépion borehole  
196 (Graulich 1961). Megasequence 2 begins with a 10 cm thick basal conglomerate, followed by  
197 the upper slope dark mudstones of the Huy Formation correlated by chitinozoans (*S. formosa*  
198 Biozone) with the middle Darriwilian (Vanmeirhaeghe 2006). The unconformity is well  
199 constrained, with a hiatus from the middle Tremadocian to the lower Darriwilian  
200 (Vanmeirhaeghe 2006; Wang and Servais 2015), i.e. a time interval close to that observed in

201 the BM (Fig. 3). After an observational gap, the sedimentation continued with the Chevreuil  
202 and the Vitruval-Bruyère formations deposited on the outer shelf to upper slope. The  
203 Chevreuil Formation was dated by chitinozoans (typical association, no index species) as late  
204 Darriwilian to early Sandbian (Vanmeirhaeghe 2006). The Vitruval-Bruyère Formation is  
205 subdivided into four members: the Giroux Member is dated by graptolites (*N. gracilis*  
206 Biozone) and chitinozoans (*L. stentor* Biozone) as lower Sandbian, the poorly dated La  
207 Bruyère Member, the Sart-Bernard and the upper part of the Rue de Courrière members, both  
208 correlated by chitinozoans (*S. cervicornis* Biozone) with the lower Katian (Vanmeirhaeghe  
209 2006, fig. 47). The uppermost Rue de Courrière Member is composed of siltstones with shell  
210 beds, sandstones and microconglomerates, which mark the base of a short stratigraphic hiatus  
211 spanning the Onnian (middle Katian) and points to the occurrence of a para-conformity. The  
212 repercussion of the drastic change of bathymetry observed in the BM at the end of  
213 Megasequence 2 (see above) lead in the Central CI to the emergence, as this region is globally  
214 shallower than the BM (Vanmeirhaeghe 2006, p. 202; Herbosch and Verniers 2014, fig. 11).

215 Megasequence 3 starts with the Fosses Formation, which is composed of shelf deposits and  
216 is subdivided into three members (Fig. 3). The two lower units were dated by chitinozoans:  
217 the Bois de Presles Member (*T. bergstroemi* and *C. rugata* biozones) is middle to late Katian  
218 in age, and the Faux-les-Tombes Member (*B. umbilicata* Biozone), latest Katian  
219 (Vanmeirhaeghe 2006). The upper Tihange Member is correlated with the Hirnantian by  
220 inference, as the overlying Bonne-Espérance Formation is early Rhuddanian in age  
221 (Vanmeirhaeghe 2006). However, the study of several fossil levels (Pereira *et al.* 2021, fig. 2)  
222 demonstrates that the upper half of the Tihange Member belongs to the *A. ascensus* and *P.*  
223 *acuminatus* graptolite biozones indicating an early Rhuddanian age (Silurian).

224

225 *The Stavelot-Venn Inlier*

226

227 The Stavelot-Venn Inlier (SVI) is located in the NE of the Ardenne Allochthon straddling the  
228 German border (Fig. 2). It is the inlier closest to the Variscan front and it was transported  
229 about 10 km northwards during the thrusting of the Ardenne Allochthon (Hance *et al.* 1999).  
230 It has undergone Variscan orogenesis, but the presence of an earlier Caledonian tectonic event  
231 is still a matter of debate (Piessens and Sintubin 1997; Sintubin *et al.* 2009; Herbosch *et al.*  
232 2020). The SVI shows a continuous, mainly terrigenous, sedimentation ranging from the early  
233 Cambrian to the Middle Ordovician.

234 The Ordovician comprises three formations: the Jalhay, Otré and Bihain formations,  
235 together about 1.2 km thick. A detailed sedimentological study of the Jalhay Formation (Fig.  
236 2; Lamens 1985) has shown a regressive sequence documenting successively the basin plain  
237 (Solwaster Member), the slope (Spa Member) and finally the upper part of the slope  
238 (Lierneux Member). The two lower members have been dated by graptolites of the *R.*  
239 *praeparabola* to *R. flabelliformis anglica* biozones to the lower Tremadocian (Wang and  
240 Servais 2015). The Lieerneux Member contains acritarchs of the informal Zone 8  
241 (Vanguetaine 1992) corresponding to the upper part of the Tremadocian (Herbosch 2021).  
242 Hence, the Jalhay Formation spans most of the Tremadocian (Fig. 3).

243 The Otré Formation contains three members that are completely devoid of microfossils  
244 due to their highly oxidizing deep-water depositional environment (Herbosch *et al.* 2016).  
245 However, the discovery of conodonts of the *P. proteus* Biozone at the boundary between the  
246 Meuville and Les Plattes members (Vanguetaine *et al.* 2004) allows assignment of this level  
247 to the uppermost Tremadocian or the lower Floian. Thus, the Otré Formation extends from  
248 the top of the Tremadocian to approximately the upper part of the Floian (Fig. 3).

249 The Bihain Formation comprises two members. The upper Salmchâteau Member contains  
250 acritarchs of the informal Zone 9 (Vanguetaine 1986, 1992), which can be correlated with

251 the *I. gibberulus* and *A. cucullus* graptolite biozones (Cooper and Molyneux 1990; Servais *et*  
252 *al.* 2017; Herbosch, 2021). The Bihain Formation extends approximately from the uppermost  
253 part of the Floian to the lowermost Darriwilian (Fig. 3). The sedimentary record of the SVI  
254 ends with the Salmchâteau Member of the Bihain Formation.

255

### 256 *The Rocroi Inlier*

257

258 The Rocroi Inlier (RI) is a E-W, about 20 by 70 km elongated area, located at the SW edge of  
259 the Ardenne Allochthon (Fig. 2). It is bounded by angular unconformity with above the  
260 uppermost Silurian and/or Lower Devonian except in the SW, where it disappears under the  
261 Mesozoic and Cenozoic cover. The RI shows a continuous siliciclastic sedimentation from the  
262 lower Cambrian to the Middle Ordovician (Beugnies 1963). The Ordovician is represented by  
263 a single formation and crops out in a very small area on the French-Belgian border.

264 The sedimentary record begins with the Vieux-Moulin de Thilay Formation (Fig. 3), which  
265 was defined by Beugnies (1963) as the top of the Revin Group (Cambrian). It was  
266 lithostratigraphically redefined as belonging to the Ordovician by Geukens (1981), a  
267 hypothesis confirmed by Roche *et al.* (1986), who found acritarchs from the informal Zone 9  
268 (Vanguetaine 1986, 1992). This zone was correlated with the *I. gibberulus* and *A. cucullus*  
269 graptolite biozones (see above) ranging from the uppermost Dapingian to the lowermost  
270 Darriwilian. Therefore, this formation has the same stratigraphic range as the Salmchâteau  
271 Member of the Bihain (SVI) and the Abbaye de Villers (BM) formations (Fig. 3).

272

### 273 *Givonne and Serpont inliers*

274



275 Ordovician strata are considered to be absent in these two less well known inliers (Beugnies  
276 1960; Geukens and Richter 1962; Herbosch 2021).

277

## 278 **The Ordovician of the Rhenish Massif and Rügen (by Lefebvre and** 279 **Servais)**

280

### 281 *General framework*

282

283 In western Germany, Ordovician strata are exposed in the eastern part of the Stavelot-Venn  
284 inlier (SVI; see above) and also in the two Ebbe and Solingen–Remscheid–Altena inliers, in  
285 the Rhenish Massif (Fig. 1; Servais *et al.* 2008, and references therein). Thick, tectonically  
286 complex Ordovician successions are also known from boreholes in Rügen (Fig. 1), in  
287 northeastern Germany, and Pomerania, in northwestern Poland (Trela, this volume). All these  
288 regions display relatively comparable successions of Ordovician rocks, yielding similar  
289 palynomorph assemblages and low-diversity deep shelf faunas dominated by graptolites and  
290 phyllocarids. Moreover, strong lithological, faunistic and geochemical similarities between  
291 Ordovician successions in Rügen, the Rhenish Massif, the SVI, the Brabant Massif, the  
292 English Lake District and the Welsh Basin strongly support the view that all these areas were  
293 part of the same Avalonia microcontinent (Maletz and Servais 1993; Servais *et al.* 1998,  
294 2008; Pharaoh 1999; Samuelsson *et al.* 2002; Verniers *et al.* 2002). In the last thirty years,  
295 there has been significant progress on the bio- and chronostratigraphy of the Ordovician of the  
296 Rhenish Massif and Rügen through a series of detailed studies based on acritarchs,  
297 chitinozoans and graptolites.

298

### 299 *Rhenish Massif*

300

301 In western Germany, the most extensive and fossiliferous succession of Ordovician  
302 sedimentary rocks is exposed in the Ebbe and Solingen–Remscheid–Altena inliers, both  
303 situated near Cologne (Sauerland), in the northeastern part of the Rhenish Massif  
304 (*Rheinisches Schiefergebirge*; Fig. 1). In this region, the c. 800 m thick Herscheid Beds  
305 (*Herscheider Schichten*) consist of a monotonous, siliciclastic succession (mudstones and  
306 siltstones), classically subdivided into four units (Maletz 2000; Eiserhardt *et al.* 2001; Servais  
307 *et al.* 2008; Koch *et al.* 2014).

308 The stratigraphically lowermost unit (Plettenberg Bänderschiefer Formation) is represented  
309 by 65 m thick dark (grey-blue) compact mudstones with intercalated pyrite-rich siltstones.  
310 The base of this formation and putatively underlying strata are not exposed in the Sauerland.  
311 The Bänderschiefer Plettenberg Formation yielded a graptolite assemblage typical of the *H.*  
312 *lentus* Biozone, which corresponds to the lower part of the *D. artus* Biozone (Fig. 4; Maletz  
313 and Servais 1993; Maletz *in* Eiserhardt *et al.* 2001; Servais *et al.* 2008; Koch *et al.* 2014). The  
314 mid Darriwilian age of the Plettenberg Bänderschiefer Formation was also confirmed by  
315 chitinozoans (Samuelsson *et al.* 2002). Other faunal elements include foraminiferans,  
316 phyllocarids, trilobites, and trace fossils (Riegraf and Niemeyer 1996; Koch and Brauckmann  
317 1998; Koch 1999; Eiserhardt *et al.* 2001). The age and faunal composition (graptolites,  
318 palynomorphs, trilobites) of the Plettenberger Bänderschifer are very similar to those of the  
319 Huy Formation in Belgium (Condroz inlier; CI, see above; Servais *et al.* 2008; Koch *et al.*  
320 2014). The trilobite assemblage indicates relatively deep, outer shelf to slope environmental  
321 conditions (Owens and Servais 1997; Servais *et al.* 2008).

322 The overlying Kiesbert Tonschiefer Formation (150–200 m thick) is composed of dark  
323 (grey-blue to black) mudstones with rare sandstone interbeds. This unit was correlated with  
324 the middle Darriwilian, based on its graptolite content, which is typical of the *N. fasciculatus*

325 Zone (upper part of the *D. artus* Biozone), and also on its chitinozoan assemblage (Fig. 4;  
326 Maletz and Servais 1993; Maletz in Eiserhardt *et al.* 2001; Samuelsson *et al.* 2002; Servais *et*  
327 *al.* 2008; Koch *et al.* 2014). The Kiesbert Tonschiefer Formation also yielded trace fossils  
328 (*Tomaculum problematicum*), brachiopods, conulariids, ostracods, phyllocarids, and trilobites  
329 (Beyer, 1941, 1943; Koch and Brauckmann 1998; Koch 1999; Schallreuter and Koch 1999;  
330 Servais *et al.* 2008).

331 The Rahlenberg Grauwackenschiefer Formation is a 300 m thick succession of thin bedded  
332 dark (blue to grey-black) silty mudstones with sandy layers that yielded siliceous concretions  
333 in its upper part. The fauna is scarce and comprises conulariids, graptolites, phyllocarids, and  
334 trilobites (Eisenack 1939; Servais *et al.* 2008). Palynomorphs indicate an early Sandbian age  
335 (Fig. 4; Eisenack 1939; Samuelsson *et al.* 2002; Servais *et al.* 2008; Koch *et al.* 2014), in a  
336 good accordance with the presence of the graptolite *Pseudoclimacograptus* sp. (Maletz 2000).

337 The Solingen Tonschiefer Formation is the stratigraphically highest Ordovician unit in the  
338 Sauerland area. It consists of about 200 m thick dark (grey to black) mudstones with silty to  
339 sandy levels, yielding very few fossil remains (graptolites, phyllocarids and trilobites) and  
340 trace fossils (Servais *et al.* 2008; Koch *et al.* 2014). An early Katian age was proposed for this  
341 formation, based on chitinozoans (Fig. 4; Samuelsson *et al.* 2002; Servais *et al.* 2008; Koch *et*  
342 *al.* 2014). The Solingen Tonschiefer Formation is overlain by upper Silurian (Pridoli) rocks of  
343 the Köbbinghauser Schichten Formation. This implies a stratigraphic gap extending from the  
344 middle Katian to most of the Silurian (Timm *et al.* 1981; Maletz and Servais 1993; Verniers  
345 *et al.* 2002).

346 In summary, the Ebbe and Solingen–Remscheid–Altena inliers expose a 800 m thick,  
347 discontinuous Ordovician succession comprising middle Darriwilian (Bänderschiefer  
348 Plettenberg and Kiesbert Tonschiefer formations), lower Sandbian (Rahlenberg  
349 Grauwackenschiefer Formation) and lower Katian (Solingen Tonschiefer Formation) rocks

350 (Fig. 4). Palaeontological data suggest the persistence (at least from the mid Darriwilian to the  
351 early Katian) of low diversity assemblages associated to relatively deep (outer shelf, slope)  
352 palaeoenvironmental conditions (Owen and Servais 1997; Samuelsson *et al.* 2002; Verniers *et*  
353 *al.* 2002; Servais *et al.* 2008; Koch *et al.* 2014).

354

### 355 *Rügen*

356

357 In northeastern Germany, a particularly thick Ordovician sedimentary succession (over 3,000  
358 m) was described from numerous boreholes drilled in Rügen (Fig. 1) during the 1960s (Jaeger  
359 1967; Servais 1994; Maletz 1998, 2001; Beier *et al.* 2001; Verniers *et al.* 2002; Servais *et al.*  
360 2008). This mainly pelitic and clastic succession is synsedimentary deformed, locally  
361 overturned (Franke and Illers 1994; Beier and Katzung 1999; Samuelsson *et al.* 2000),  
362 suggesting unstable environmental conditions (slumps) along an active continental margin  
363 and/or tectonic stacking (Servais and Katzung 1993; Giese *et al.* 1994; McCann 1998). With  
364 the exception of graptolites and palynomorphs, only trace fossils (e.g. *Nereites*) and  
365 phyllocarid crustaceans were reported from the boreholes (Zagora 1997; Maletz 2001). A  
366 subdivision of the Ordovician succession into three tectonically limited lithostratigraphic units  
367 was established, based on the thickest and most complete drilled successions (e.g. the Rügen  
368 5 borehole) (Giese *et al.* 1994; Beier *et al.* 2001; Servais *et al.* 2008).

369 The stratigraphically lowermost unit, the Varnkevitz Sandstein Formation, was only  
370 observed in the deepest and most complete borehole (Rügen 5). It consists of about 400 m of  
371 fine sandstones and intercalated shales (Giese *et al.* 1994; Beier *et al.* 2001). About 60 m  
372 above the base of this formation, a 200 m thick interval of black shales yielded both acritarchs  
373 typical of the *messaoudensis-trifidum* assemblage and a *Lagenochitina destombesi*-dominated  
374 chitinozoan association, both supporting a late Tremadocian age (Fig. 4; Servais and Katzung

375 1993; Servais and Molyneux 1997; Samuelsson 1999; Beier *et al.* 2001; Samuelsson *et al.*  
376 2000; Servais *et al.* 2008).

377 The overlying Arkona Schwarzschiefer Formation is a very thick succession of black  
378 shales (over 1,000 m) observed in most boreholes (Samuelsson *et al.* 2000; Maletz 2001).  
379 Both its lower and upper parts are tectonically disturbed, and no borehole yielded a complete,  
380 continuous succession (many intervals are missing, probably due to tectonics). These levels  
381 have yielded graptolite assemblages characteristic of the middle and upper Darriwilian  
382 (equivalents of the *H. lentus*, *P. elegans*, *P. distichus*, and ?*H. teretiusculus* biozones) and the  
383 lower Sandbian (*N. gracilis* Biozone) (Fig. 4; Jaeger 1967; Maletz 1998, 2001).  
384 Palynomorphs provided a comparable range of ages for the Arkona Schwarzschiefer  
385 Formation (Burmam 1968, 1970; Samuelsson *et al.* 2000; Samuelsson and Servais 2001;  
386 Vecoli and Samuelsson 2001; Servais *et al.* 2008).

387 The Nobbin Grauwacken Formation is a very thick (~1,800 m) succession dominated by  
388 greywackes and shales, which was only observed in some boreholes. Graptolite and  
389 palynomorph assemblages yielded congruent ages (middle Darriwilian–early Sandbian) for  
390 this unit, similar to those obtained for the Arkona Schwarzschiefer Formation (Fig. 4; Maletz  
391 1998, 2001; Samuelsson *et al.* 2000; Servais *et al.* 2001, 2008). This implies that both  
392 formations were probably concurrently deposited in distinct environmental conditions; the  
393 black shales of the Arkona Schwarzschiefer Formation were probably related to more distal  
394 (bathyal?) settings than the greywackes of the Nobbin Grauwacken Formation (Zagora 1997;  
395 Samuelsson *et al.* 2000).

396 In summary, the subsurface of the island of Rügen yielded a very thick, though tectonically  
397 complex, Ordovician succession comprising Tremadocian (Varnkevitze Sandstein Formation)  
398 and middle Darriwilian to lower Sandbian (Arkona Schwarzschiefer and Nobbin Grauwacken  
399 formations) sedimentary rocks (Fig. 4). The Ordovician successions of Rügen, the Rhenish

400 Massif (see above) and Pomerania (NW Poland; see Trela, this volume) share many  
401 similarities in terms of lithologies, faunas, stratigraphy (e.g. latest Tremadocian–early  
402 Darriwilian gap), and associated environmental conditions (both were originally deposited in  
403 deep-water settings), thus supporting the view that they were parts of a same  
404 palaeogeographic area (Jaeger 1967), in eastern Avalonia (Maletz and Servais 1993; Servais  
405 *et al.* 1998; Pharaoh 1999; Maletz 2001; Samuelsson *et al.* 2002; Verniers *et al.* 2002).

406

## 407 **The Ordovician of the Armorican Massif (by Vidal, Loi, Lefebvre and** 408 **Ghienne)**

409

### 410 *General framework*

411

412 The Armorican Massif is usually subdivided into four tectonic units (Fig. 5): the Leon  
413 Domain (LD), the North Armorican Domain (NAD), the Median Armorican Domain (MAD,  
414 and the South Armorican Domain (SAD) (Matte 2001; Ballèvre *et al.* 2009). As the LD did  
415 not contain any Palaeozoic sedimentary successions, only the three other domains will be  
416 described below. A subdivision into similar tectonic units, with comparable stratigraphic and  
417 faunal features, occurs also in the Iberian Peninsula, thus strongly supporting the view that  
418 these two areas, united by a similar evolution during the Variscan orogeny, originally  
419 belonged to a same Ibero-Armorican palaeogeographic Domain (e.g. Paris and Robardet  
420 1977, 1994; Robardet and Gutiérrez-Marco 1990; Robardet *et al.* 1994b; Gutiérrez-Marco *et*  
421 *al.* this volume).

422 The NAD (i.e. Cotentin, zone bocaine and Mortain-Domfront area) and the MAD (i.e.  
423 Châteaulin, Menez-Belair and Laval synclines) are separated by the North Armorican Shear  
424 Zone (NASZ). Their Palaeozoic successions are quite similar from the Middle Ordovician

425 onwards (Fig. 6). Based on their tectonic history and faunistic affinities, the NAD and MAD  
426 are often united within a single Median-North Armorican Domain (MNAD; Robardet *et al.*  
427 1994a, b; Paris 2016). However, recent U-Pb analyses on detrital zircons coupled with Sm-Nd  
428 and Lu-Hf isotope analyses identified different zircon populations in the Lower Palaeozoic  
429 successions of these two areas (Dabard *et al.* 2021). In the Ordovician sediments of the MAD,  
430 the occurrence of Stenian detrital zircons suggests affinities with the eastern African "Sahara  
431 metacraton" and the Arabian Nubian Shield. In contrast, the absence of Stenian and late  
432 Tonian ages in zircon populations in Ordovician sediments from the NAD implies distinct  
433 source areas, probably located more westward. These features suggest that the NAD and the  
434 MAD were geographically separated during the Early Ordovician, and moving closer only  
435 during the Variscan orogeny (Fig. 1).

436 The MAD is separated from the SAD (i.e. south of Saint-Julien-de-Vouvantes unit, Saint-  
437 Georges-sur-Loire and Ancenis synclines in southern Brittany, and Chantonay syncline in  
438 Vendée) by the northern branch of the South Armorican Shear Zone (SASZ, Fig. 5). The  
439 Palaeozoic succession of the SAD is markedly distinct from those of the NAD and MAD. The  
440 three domains differ in the lower part of the Ordovician succession, which rests conformably  
441 on the lower Cambrian in Normandy-NAD (Séries de Carteret; Doré 1969; Pillola 1993), on  
442 the Miaolingian in the SAD (Schistes à *Paradoxides* of Cléré-sur-Layon, South of Angers;  
443 Cavet *et al.* 1966) and unconformably on the Brioverian basement (see below) in central  
444 Brittany (MAD; Robardet *et al.* 1994a, b).

445 The main features of the Ordovician succession in the MAD, which is the most complete  
446 and best exposed within the Armorican Massif, will be presented in detail below, with brief  
447 references about differences with the NAD and the SAD.

448

449 *Lower Ordovician*

450

451 In central Brittany, the Ordovician succession (Fig. 6) begins with basal red-bed sequences  
452 (*Séries Rouges Initiales* or Cap de la Chèvre Formation), followed by the geographically  
453 widespread Grès Armorica Formation. These lowest Ordovician deposits rest  
454 unconformably on the Brioverian succession (Gougeon *et al.* 2022 and references therein).  
455 Probably diachronous within the Armorican Massif, the Brioverian succession is dated as late  
456 Neoproterozoic in both the NAD and the SAD, based on their stratigraphic situation below  
457 the Cambrian. In the MNAD, recent investigations on historical and new outcrops have  
458 yielded low diversity trace-fossils with not penetrative but simple horizontal patterns,  
459 microbially induced sedimentary structures and very rare soft-body fossils (*Nimbia* sp.),  
460 coupled with U-Pb detrital zircon datings, all suggesting an Ediacaran–Fortunian age for at  
461 least the upper part of the Brioverian succession (Gougeon *et al.* 2018, 2022; Néraudeau *et al.*  
462 2018).

463 Within the Armorican Massif, red-bed sequences are discontinuous (named Cap de la  
464 Chèvre Formation in the southern part of the Crozon Peninsula, Pont-Réan Formation in the  
465 synclines south of Rennes, and Margat Formation close to Angers in the SAD) and start with  
466 a polymict conglomerate overlain by sandstones interbedded with silty-clayey sediments  
467 (Bonjour 1988; Bonjour and Chauvel 1988; Suire *et al.* 1991; Egal *et al.* 1996). The source  
468 area of sediments is the Brioverian basement (Dabard *et al.* 2021). The main sedimentary  
469 structures are either plane- or cross-bedding laminations in the sandstones and desiccation  
470 cracks in some siltstones or argillites levels. The geometry of sedimentary bodies and the  
471 associated structures suggest depositional conditions ranging from alluvial fans and braided  
472 river systems to alluvial plains (Bonjour 1988; Durand 1989; Suire *et al.* 1991; Egal *et al.*  
473 1996). Then, the environment progressively opened up to the marine domain, probably as a  
474 storm-influenced delta. Despite the lack of body fossils, the occurrence of living organisms is



475 evidenced in the upper part of these red-bed sequences by the record of bioturbation traces  
476 such as *Cruziana*-type bilobed tracks and *Skolithos*-type vertical burrows.

477 This lowermost part of the Ordovician succession was deposited on a rather unstable  
478 continental basement, as evidenced by the contemporary effusive volcanism and variations in  
479 lateral thickness, which are indicative of a significant tectonically induced subsidence (Noblet  
480 1983; Bonjour 1988). This instability was controlled by extensional tectonics (Ballard *et al.*  
481 1986; Dauteuil *et al.* 1987; Brun *et al.* 1991) expressed in the Cambrian and likely associated  
482 with the opening of the Rheic Ocean between Avalonia and Gondwana (Paris and Robardet  
483 1990; Prigmore *et al.* 1997; Linnemann *et al.* 2008). The dating applied to interbedded  
484 volcanoclastic levels have assigned numeric ages ranging from  $472 \pm 5$  Ma (Rb-Sr, red-bed  
485 sequences of northern Brittany; Auvray *et al.* 1980) to  $465 \pm 1$  Ma (U-Pb, Cap de la Chèvre  
486 Formation; Bonjour *et al.* 1988). The latter age is, however, inconsistent with the  
487 biostratigraphic correlations of overlying formations (e.g. the chitinozoan *E. brevis* Zone,  
488 indicative of the upper Floian; see Paris 1990, 2016).

489 The thickness of the Grès Armoricaïn Formation varies from tens to several hundreds of  
490 meters. Depending on the regions, it consists of either an undivided mass of sandstones with  
491 silty-clayey intercalations (e.g. Domfront in Normandy, northern flank of Menez-Bélaïr,  
492 Vendée), or is heterogeneous and subdivided into three members: (i) a lower member  
493 including thick sandstone beds only separated by few silty to clayey intercalations. This  
494 member can comprise iron layers, which used to be exploited in the synclines south of Rennes  
495 (Chauvel 1968); (ii) a middle member (variously named "Congrier Member", "Gador  
496 Member" or *Schistes intermédiaires* depending on the region), consisting of siltstones,  
497 argillites with micaceous sandstone intercalations; and (iii) an upper member, which is mainly  
498 composed of sandstones with silty-clayey intercalations (e.g. in the Châteaulin synclinorium,

499 synclines south of Rennes), and containing sometimes highly radioactive placers (rutile and  
500 zircon grains; Faure 1978; Pistis *et al.* 2008, 2016).

501 One of the main features of the Grès Armorica Formation is the occurrence of numerous  
502 sedimentary structures (e.g. hummocky cross stratification, planar laminations, erosive basal  
503 surfaces, wave and current ripples and clay drapes). They reflect shallow environments  
504 subject to tidal influence and to fair-weather and storm waves (Durand 1985; Guillocheau *et*  
505 *al.* 2009). Moreover, this formation has yielded highly diverse trace fossils, which are either  
506 vertical or horizontal (Durand 1985). Surprisingly, the fauna is sparse with few *Lingula*-rich  
507 levels and, in some areas, brachiopods such as "*Dinobolus*". In the uppermost part of the Grès  
508 Armorica Formation, rare bivalves (Babin 1966) and trilobites (Henry 1980) occur.

509 Sedimentological and palaeobiological data suggest that the Grès Armorica Formation  
510 was deposited in environments ranging from a protected marine domain (bay, lagoon) to the  
511 top of the continental shelf (shoreface to upper offshore; Durand 1985; Dabard *et al.* 2007;  
512 Guillocheau *et al.* 2009; Pistis *et al.* 2016). Some clayey levels of the middle and upper  
513 members yielded chitinozoans typical of the *E. brevis* Biozone (Paris 1990), thus supporting a  
514 Floian age for these levels. The huge variations in thickness observed for the Grès Armorica  
515 Formation (i.e. from a few meters up to 600 meters) suggest very high local rates of  
516 sedimentation (Noblet 1983; Durand 1985; Dabard *et al.* 2009), probably driven by an active  
517 tectonic influence.

518

### 519 *Middle Ordovician – Sandbian*

520

521 During the Dapingian–Sandbian interval, the sedimentation was dominated by the deposition  
522 of dark, fine-grained shales, previously called "Schistes à *Calymene*", and currently named  
523 the Postolonnec, Andouillé, Le Pissot and Traveusot formations in the MNAD, Schistes

524 d'Angers and Fresne formations in the SAD (Fig. 6). These 150 to 400 m thick silty-clayey  
525 units bear levels with siliceous, phosphatic and/or calcareous concretions, shell-beds, and  
526 sandstone intercalations that can be several dozen of meters thick (e.g. Kerarvail Member,  
527 about 30 meters thick in the Crozon Peninsula). The main sedimentary structures are planar  
528 bedding, hummocky cross stratifications and numerous A- or B-type shell beds (type A: thin,  
529 lenticular single coquinas; type B: thick, continuous and composite shell concentrations,  
530 *sensu* Botquelen *et al.* 2004). All these structures are indicative of a storm wave-influenced  
531 sedimentation. These deposits are related to various environments ranging from beach  
532 (shoreface) to lower offshore (Guillocheau 1983; Loi *et al.* 1999; Loi and Dabard 2002;  
533 Dabard *et al.* 2007; Guillocheau *et al.* 2009). Phosphatic and siliceous concretions, and crusts  
534 are common in Middle Ordovician sedimentary rocks. They are of early diagenetic origin and  
535 are promoted by the stability of the water/sediment interface. These concretions, which are  
536 indicative of sedimentary condensation, were mainly produced during sea-level rises (Loi *et*  
537 *al.* 1999; Loi and Dabard 2002; Dabard *et al.* 2007; Dabard and Loi 2012). Depending on the  
538 depositional environment, they can be linked to composite shell beds (B-type shell beds *sensu*  
539 Botquelen *et al.* 2004).

540 Bioturbation is less diverse than in the underlying Grès Armorica Formation, with  
541 *Planolites* and *Teichichnus* as the most common trace fossils (Guillocheau 1983; Guillocheau  
542 *et al.* 2009). In contrast, invertebrate faunas are particularly diverse and consist mainly of  
543 benthic taxa (Vidal *et al.* 2011a): bivalves (Babin 1966), brachiopods (Mélou 1973, 1975,  
544 1976; Botquelen and Mélou 2007), echinoderms with aristocystitid diploporites, crinoids,  
545 ophiuroids and stylophorans (Chauvel 1941, 1980, 1981; Chauvel and Nion 1969, 1977;  
546 Lefebvre and Vizcaïno 1999; Lefebvre 2000, 2007; Hunter *et al.* 2007; Lefebvre *et al.* 2015,  
547 2022; Blake *et al.* 2016), ostracods (Vannier 1986a, b), and trilobites (Henry 1980; Vidal *et*  
548 *al.* 2011b; Gendry *et al.* 2013; Courville and Gendry 2016). Additional faunal elements

549 include conodonts (Lindström *et al.* 1974), graptolites (Philippot 1950; Henry *et al.* 1976;  
550 Paris and Skevington 1979) and organic-walled microfossils, such as acritarchs (Deunff 1951,  
551 1954, 1958; Henry 1969; Paris and Le Hérisse 1992) and chitinozoans (Paris 1981, 2016).  
552 The ages inferred from graptolites and chitinozoans range from the Dapingian (*D. ornensis*  
553 Biozone) to the Sandbian (*L. dalbyensis* Biozone; Paris 1981, 1990, 2016; Dabard *et al.*  
554 2015). The benthic faunas display clear Gondwanan affinities, particularly with the  
555 Mediterranean Province and are particularly similar to those described in the Iberian  
556 Peninsula, highlighting the long-established similarities between the faunas of these two  
557 regions (e.g. Henry *et al.* 1976; Paris and Robardet 1977, 1994). Some levels record the  
558 immigration of some taxa following transgressive events, as for example the *Marrolithus*  
559 *bureaui* level (Henry *et al.* 1993), which coincides with the appearance of Balto-Scandinavian  
560 chitinozoans in the lower Sandbian (e.g. *L. stentor*; Paris 2016).

561 Differences in the composition of Darriwilian–Sandbian benthic assemblages in the North-  
562 Median- and South Armorican Domains mainly result from distinct environmental settings,  
563 with a southward (in present-day geography) deepening trend (Henry 1989; Henry *et al.* 1997;  
564 Lefebvre 2007; Courville and Gendry 2016). In the NAD, benthic diversity is low and  
565 trilobite taxa (Homalonotinae) are typically related to shallow-water environments (Henry  
566 1989 and references therein). In the MAD, the faunal diversity is significantly higher and  
567 characteristic of median shelf settings with both typical endobenthic and epibenthic trilobite  
568 taxa (Henry 1989; Courville and Gendry 2016). In the SAD, fossils are rare and outcrops  
569 discontinuous. However, assemblages are typical of distal shelf environments, with strong  
570 affinities with coeval Bohemian faunas from the Prague Basin (Henry *et al.* 1997; Lefebvre *et*  
571 *al.* 2010). Assemblages correspond to the "atheloptic biofacies" with benthic trilobites either  
572 blind or with atrophied eyes, associated with pelagic forms with hypertrophied eyes  
573 (Cyclopygidae), along with lagynocystid mitrate stylophorans, typical of deep-water

574 environments of the Prague Basin (Henry *et al.* 1997; Lefebvre 2007; Lardeux *et al.* 2008;  
575 Jauhier and Gendry 2017).

576 Several transgressive and regressive cycles were identified in the Middle Ordovician–  
577 Sandbian succession of the MAD, based on the detailed analysis of sedimentary facies  
578 (Dabard *et al.* 2007, 2009; Paris *et al.* 2007). Transgressive episodes identified within the *D.*  
579 *bullata*, *S. formosa* and *L. pissotensis* chitinozoan biozones were recognized in other peri-  
580 Gondwanan areas (Dabard *et al.* 2007; Paris *et al.* 2007; Videt *et al.* 2010). In the Crozon  
581 Peninsula, two key sections within the Postolonnec Formation provided a continuous, well-  
582 exposed framework for the detailed analysis of sedimentary facies and gamma-ray record  
583 (Dabard *et al.* 2015). There, a back stripping procedure was applied and calibrated with  
584 graptolite biozones to convert the observed variations in depositional environments into  
585 apparent sea-level changes, and ultimately into eustatic variations. The resulting pattern (Fig.  
586 7) supports the occurrence of short-term and high amplitude cyclic sea-level fluctuations,  
587 suggesting glacio-eustasy. This pattern suggests the occurrence of ice-house conditions and  
588 oscillations of an ice-sheet on Gondwana at least at the lower limit of the *L. pissotensis*  
589 Biozone (at the transition between the Kerarvail and Morgat members), but probably as early  
590 as the *S. formosa* chitinozoan Biozone during the Darriwilian (i.e. at the transition between  
591 the Kerloc'h and Corrégou members, Fig. 6; Dabard *et al.* 2015). The expression of these ice  
592 sheet-induced oscillations differs in other parts of the Armorican Massif, depending on their  
593 initial location along a proximo-distal transect and/or erosional surfaces. For example, no  
594 sandy member (equivalent to Kerarvail in Crozon) is observed in the central and eastern parts  
595 of the MAD, characterized by more distal palaeoenvironmental conditions. In Normandy, the  
596 coarser, shallow-water facies of the Middle Ordovician successions and the quality of the  
597 outcrops do not make it possible to achieve any coherent, detailed analysis. Nevertheless, all

598 these different sections may be correlated based on their chitinozoan assemblages (Fig. 6;  
599 Paris 2016).

600

601 *Katian*

602

603 In many regions of the Mediterranean Province, the Sandbian–Katian transition coincides  
604 with a major regressive episode (Robardet 1981; Villas 1992; Bourahrouh *et al.* 2004). In  
605 both Brittany and Portugal, the recovery of the sedimentation process is expressed by a  
606 conglomeratic layer bearing phosphatic pebbles, ferruginous and phosphatic oolites (Paris  
607 1979, 1981; Young 1988). Palynological data suggest a stratigraphic gap, at least locally:  
608 chitinozoans extracted from the conglomerate matrix belong to the *E. tanvillensis* Biozone  
609 (Katian), whilst those from the pebbles belong to the *L. deunffi* or *L. dalbyensis* biozones  
610 (Sandbian; Paris 1981; Bourahrouh 2002; Vidal *et al.* 2011*b*; Dabard *et al.* 2015). This level  
611 represents the base of a new transgressive sequence corresponding to the Kermeur and Saint-  
612 Germain-sur-Ille formations in the median synclinorium, and to the Grès du Châtellier  
613 followed by the Schistes de Riadan south of Rennes (Dabard *et al.* 2009; Paris 2016). These  
614 sandstone-dominated deposits contain silty and clayey intercalations, as well as phosphatic,  
615 siliceous and/or calcareous concretions. The sedimentary structures (hummocky cross-  
616 stratifications, symmetric ripples [=oscillation ripples], clay drapes, inversion in palaeocurrent  
617 directions, evidence of emersion) reflect shallow environments subjected to tidal influence  
618 and to fair-weather and storm waves. The depositional environments inferred from the  
619 sedimentary facies range from barrier and back-barrier nearshore settings (e.g. base of the  
620 Kermeur Formation; Vidal *et al.* 2011*b*) to the upper offshore (Gorini *et al.* 2008) throughout  
621 several transgression-regression cycles (Gorini *et al.* 2008; Dabard *et al.* 2009). Some levels  
622 are extensively bioturbated by well-diversified trace fossils (Mélou and Plusquellec 1975),

623 including *Equilibrichnia* (Vidal *et al.* 2011b) and spiral-shaped traces. Benthic assemblages  
624 are scattered in the arenaceous base of the Kermeur Formation, and in the siltstones of the  
625 upper part of this formation, equivalent to the Châtellier and Riadan formations. The main  
626 taxa are trilobites (see Henry 1980; Vidal *et al.* 2011b), diploporite echinoderms, brachiopods,  
627 and rare bivalves (Babin and Mélou 1972; Mélou 1985; Botquelen and Mélou 2007) and  
628 gastropods. These faunas are less diversified than those occurring in underlying formations.  
629 Accurate stratigraphic assignments of the Kermeur, Saint-Germain-sur-Ille, Châtellier and  
630 Riadan formations are provided by chitinozoans, indicative of the Katian (from the *E.*  
631 *tanvillensis* to the *A. barbata* biozones; Paris 1981, 2016; Bourahrouh 2002; Gorini *et al.*  
632 2008). No coeval section is exposed in the SAD, but concretions yielded a typical Katian  
633 trilobite fauna with *Deanaspis*, *Dreyfussina*, *Eudolatites* and *Prionocheilus* (Beaulieu *et al.*  
634 2014).

635 The widespread upper Katian calcareous facies occurring in most regions of the  
636 Mediterranean Province have been interpreted as reflecting a period of global warming, the  
637 so-called "Boda Event" (Fortey and Cocks 2005). In the Armorican Massif, they correspond  
638 to spatially restricted units, such as the Vaux Limestone in Normandy (Sées syncline) and the  
639 Rosan Formation in the Crozon Peninsula (Fig. 6). In the latter region, calcareous facies are  
640 associated with an anorogenic volcanic complex suggestive of shallow submarine to partially  
641 aerial conditions (Juteau *et al.* 2007; Caroff *et al.* 2009). The limestones yielded brachiopods  
642 (Mélou 1990), bryozoans, conodonts (Lindström and Pelhate 1971; Paris *et al.* 1981), and  
643 locally also echinoderms (crinoids and cystoids; Chauvel and Le Menn 1972) and trilobites  
644 together with the oldest occurrence of rugose corals in the Armorican Massif. Brachiopods  
645 and conodonts correspond to a mid-late Katian age (*A. ordovicicus* Zone; Paris *et al.* 1981;  
646 Mélou 1990; Ferretti *et al.* 2014).

647

648 *Hirnantian*

649

650 In the Mediterranean Province, the Hirnantian stage is characterized by a well-recorded  
651 glacial episode in most regions (e.g. Algeria, Morocco, Portugal, Spain; Ghienne *et al.* 2007;  
652 Loi *et al.* 2010; Videt *et al.* 2010). In the Armorican Massif, this interval corresponds to silty-  
653 clayey deposits containing scattered clasts of various sizes and lithologies, which are  
654 interpreted as dropstones (Dangeard and Doré 1971; Robardet and Doré 1988). Hirnantian  
655 deposits constitute the *Pélites à fragments* (diamictites) in Normandy, the uppermost part of  
656 the Saint-Germain-sur-Ille Formation in Menez-Bélaire and Laval synclines, the Chesnaie  
657 Formation in the South of Rennes or the Cosquer Formation in western Brittany. These facies  
658 are interpreted as glaciomarine deposits composed of fragments originating from melting ice  
659 (Hamoumi 1981; Hamoumi *et al.* 1981; Robardet and Doré 1988).

660 In these levels, the macrofauna is scarce. In the Crozon Peninsula, a single locality  
661 (Trégarvan section) yielded a fossiliferous sandstone level with *Hirnantia sagittifera* (Mélou  
662 1990), elsewhere named “Grès de Roudou Hir” or “Grès de Lamm Saoz” at the top of the  
663 Cosquer Formation (Plusquellec *et al.* 1999). This ubiquitous brachiopod is emblematic of the  
664 Hirnantian Stage. This age is also supported by palynological data (chitinozoans), which are  
665 typical of the *T. elongata* Biozone (Bourahrouh 2002; Paris 2016).

666

667 **The Ordovician of the Montagne Noire, Mouthoumet and eastern Pyrenees**  
668 **massifs (by Álvaro, Casas, Lefebvre, Monceret and Vizcaino)**

669

670 *General framework*

671



672 After the late Cambrian rifting along the so-called Atlas – Ossa-Morena – North Armorican  
673 rift (western Gondwana; Álvaro *et al.* 2021), drifting conditions prevailed during the  
674 Ordovician. However, the evolution of the high-latitude western Gondwanan margin was not  
675 a monotonous journey, but an eventful drift episodically punctuated by major geodynamics  
676 events. One of them was initially identified as a Middle Ordovician gap in southwestern  
677 Sardinia: the Sardinic Phase (see Loi *et al.* this volume) has during the last few decades become  
678 a standard for correlation in western Europe, in spite of sometimes different conceptions both  
679 regionally and academically. Similar gaps involving the absence of Middle Ordovician strata  
680 have been invoked to recognize Sardinic-style geodynamic events in the eastern Pyrenees,  
681 southern France and the Alps (Stampfli *et al.* 2002). There, the Sardinic phase is characterized  
682 by generalized cortical uplift, denudation of exposed uplifted areas under subaerial  
683 conditions, Middle Ordovician stratigraphic gaps of about 25–30 million years, broad felsic  
684 plutonism and volcanism (with felsic rocks now orthogneisses after Variscan deformation and  
685 metamorphism) with calc-alkaline affinity, and the record of alluvial to fluvial deposits  
686 onlapping an inherited palaeorelief (Álvaro *et al.* 2018; Casas and Álvaro 2019). However,  
687 the record of Ordovician volcanism associated with local listric faults in neighbouring areas  
688 cannot be used as a supporting argument for contemporaneous Sardinic events (Álvaro *et al.*  
689 2020).

690

### 691 *Tectonostratigraphic framework*

692

693 The southern Massif Central or Occitan Domain (Pouclet *et al.* 2017) comprises the Variscan  
694 tectonostratigraphic units of the Thiviers-Payzac and Génis, Rouergue-Albigeois, Montagne  
695 Noire, southern Cévennes and Mouthoumet massifs (Álvaro *et al.* 2021) (Fig. 8). Two massifs  
696 are selected for comparison here: the Montagne Noire and the Mouthoumet massifs. The

697 former has been traditionally subdivided into three tectonostratigraphic units, which are: (i)  
698 the northern Montagne Noire with a southward tectonic vergence, comprising Lower  
699 Palaeozoic low-grade metasedimentary units; (ii) the Axial Zone, an elongated dome of  
700 migmatized orthogneisses displaying an Ordovician age for their protoliths (460–450 Ma;  
701 Roger *et al.* 2004), which also includes migmatites and micaschists derived from  
702 Neoproterozoic–Ordovician rocks; and (iii) the southern Montagne Noire made up of south-  
703 facing nappes including a complete and fossiliferous Cambrian–Ordovician succession  
704 (Vizcaïno *et al.* 2001) (Fig. 8). The emplacement of recumbent folds during the Visean–early  
705 Namurian altered the pre–Variscan (palaeogeographic) proximal-to-distal trend, placing the  
706 (proximal) Axial Zone in an intermediate position between the southern and (distal) northern  
707 Montagne Noire flanks. The Mouthoumet Massif or inlier, located between the Montagne  
708 Noire and the North Pyrenean frontal thrust, represents the southernmost prolongation of the  
709 southern Massif Central (Berger *et al.* 1997).

710 The pre-Variscan rocks exposed in the eastern Pyrenees form an elongated strip that crops  
711 out along the backbone of the Alpine cordillera. The main tectonostratigraphic units  
712 recognized in the eastern Pyrenees are the Canigó/Canigou unit, comprising the Canigó and  
713 the Roc de Frausa/Roc de France massifs, the Albera/Albères-Cap de Creus unit and the  
714 Aspres-Conflent unit (Fig. 8). The Canigó unit exhibits the most complete pre–Variscan  
715 sequence and separates palaeogeographically a proximal sector to the South (Puigmal sector)  
716 from a distal one to the North (Conflent unit), which mimics the same bathymetric trend  
717 displayed across the southern to northern sides of the Montagne Noire (Álvaro *et al.* 2021).

718

719 *Lower Ordovician (pre–Sardic) stratigraphic and volcano-sedimentary features*

720

721 A comprehensive summary of the Lower Ordovician stratigraphic nomenclatures from the  
722 southern Montagne Noire was presented by Vizcaïno *et al.* (2001), who subdivided the  
723 succession into several formations, alternating between shale-dominant (La Gardie, Saint-  
724 Chinian, Setso and Landeyran units) and sandstone-dominant (La Dentelle, La Maurerie,  
725 Cluse de l'Orb and Foulon) intervals, locally punctuated by carbonate interbeds (Val d'Homs  
726 and Mounio formations). A regional biozonation (Fig. 9) of the Tremadocian–Floian interval  
727 was proposed by Vizcaïno and Álvaro (2003), Tortello *et al.* (2006), and Álvaro *et al.* (2007,  
728 2008). This biostratigraphic scheme relies on correlations with Avalonia (England), Baltica  
729 (Scandinavia) and high-latitude (peri-)Gondwana areas (e.g. Bohemian Massif and Sardinia).  
730 It comprises, from bottom to top the *P. geinitzi* (mid Tremadocian), *S. (C.) pusilla* (mid–late  
731 Tremadocian), *E. filacovi* (late Tremadocian), *T. miqueli* (latest Tremadocian–early Floian),  
732 *T. shui landayranensis* (mid Floian), *C. maynardensis* (mid Floian), *N. (N.) arenosus* (mid  
733 Floian), *A. incisus* (late Floian) and *H. primitivus* biozones (late Floian) (Fig. 9). The base of  
734 each biozone corresponds to the FAD of its index taxon (Vizcaïno and Álvaro 2003). This  
735 biozonation was completed by the presence of conodonts in the limestone interbeds belonging  
736 to the Tremadocian *P. deltifer* Biozone (Álvaro *et al.* 2007; Serpagli *et al.* 2007).

737 In the southern Montagne Noire, the Lower Ordovician succession records a shallowing-  
738 upward trend from the moderately deep shelf settings, below storm wave base, of the Saint-  
739 Chinian Formation (upper Tremadocian) to the more proximal, storm-dominated deposits of  
740 the Cluse de l'Orb and Foulon formations (mid Floian) (Courtessole *et al.* 1985; Dabard and  
741 Chauvel 1991; Vidal 1996; Vizcaïno and Lefebvre 1999; Vizcaïno *et al.* 2001). In the upper  
742 part of the Lower Ordovician succession, the black shales of the Landeyran Formation  
743 correspond to the onset of more distal, deeper offshore settings (Vizcaïno and Lefebvre 1999;  
744 Vizcaïno *et al.* 2001; Van Iten and Lefebvre 2020).

745 These changes in environmental conditions are also reflected by turnovers in the  
746 composition and diversity of faunal assemblages. The Saint-Chinian and La Maurerie  
747 formations yielded very diverse and abundant faunas comprising brachiopods, conulariids,  
748 echinoderms, graptolites, hyolithids, machaeridians, molluscs, and trilobites (e.g. Thoral  
749 1935; Courtessole *et al.* 1983; Vizcaïno *et al.* 2001). In marked contrast, the overlying Cluse  
750 de l'Orb and Foulon formations are characterized by low-diversity assemblages dominated by  
751 non articulate brachiopods, eocrinoids and trilobites (e.g. Thoral 1935; Courtessole *et al.*  
752 1985; Dabard and Chauvel 1991; Vidal 1996; Vizcaïno and Lefebvre 1999; Vizcaïno *et al.*  
753 2001). The Landeyran Formation documents a major faunal transition, with the re-occurrence  
754 of several groups absent from the underlying Cluse de l'Orb and Foulon formations, but  
755 present in the Saint-Chinian and La Maurerie formations (e.g. agnostids, raphiophorid  
756 trilobites, stylophoran echinoderms). The Landeyran Formation yielded very diverse faunal  
757 assemblages including brachiopods, conulariids, echinoderms, graptolites, hyolithids,  
758 machaeridians, molluscs, ostracods, and trilobites (e.g. Dean 1966; Capéra *et al.* 1978;  
759 Courtessole *et al.* 1983; Vizcaïno *et al.* 2001; Tortello *et al.* 2006; Van Iten and Lefebvre  
760 2020).

761 Two Early Ordovician magmatic events took place in the Occitan Domain that are absent  
762 in the Pyrenees. First, a Tremadocian felsic magmatism is recorded in the parautochthonous  
763 and allochthonous units of the Albigeois and northern Montagne Noire (Larroque Volcanic  
764 Formation), and the Mouthoumet Massif (Davejean acidic volcanics). This Tremadocian  
765 felsic magmatism was interpreted as the result of an episode of massive crustal melting  
766 related to asthenospheric upwelling leading to lithospheric doming and continental extension  
767 and decompressionally driven mantle melting, which triggered abundant subaerial explosive  
768 and effusive rhyolitic eruptions similar in facies and composition to the “Ollo de Sapo” event  
769 reported in the Iberian Massif. The second Early Ordovician magmatic event is illustrated by

770 the Peyrebrune and Davejean Volcanic Complex basalts, which display a typical initial rift  
771 tholeiite (IRT) magmatic signature indicating contributions of both asthenospheric and  
772 lithospheric mantle sources (Pouclet *et al.* 2017).

773 In contrast, Furongian–Lower Ordovician strata in the eastern Pyrenees represent a  
774 quiescent time span with no remarkable tectonic activity. An unconformity-free monotonous  
775 succession of shales with subsidiary sandstone interbeds is recognized there as the Serdinya  
776 Formation (Jujols Group), with a characteristic sandstone-dominated upper part distinguished  
777 as the Font Frède Member (Padel *et al.* 2018). Acritarchs recovered from the uppermost part  
778 of the formation in the southern Canigó Massif have yielded a broad Furongian–Early  
779 Ordovician age (Casas and Palacios 2012).

780

### 781 *The Sardinic Phase*

782

783 In the Montagne Noire, Mouthoumet and eastern Pyrenees massifs, a late Early–Mid  
784 Ordovician phase of uplift, tilting and denudation of a broad palaeorelief subaerially exposed  
785 led to the Sardinic unconformity. Uplift was associated with magmatic activity, which  
786 continued until the Late Ordovician. Consequently, Middle Ordovician strata are absent in the  
787 Occitan Domain.

788 In the eastern Pyrenees, the Sardinic Phase is linked to a time gap of about 16–23 million  
789 years. Coeval with the late Early–Mid Ordovician phase of generalized uplift and denudation,  
790 key magmatic activity led to the intrusion of voluminous granitoids, about 500 to 3000 m  
791 thick and encased in Ediacaran–Cambrian Series 2 strata. These granitoids constitute the  
792 protoliths of the large orthogneissic laccoliths that punctuate the backbone of the central and  
793 eastern Pyrenees. They form, from west to east, the Aston (467–470 Ma; Denèle *et al.* 2009;  
794 Mezger and Gerdes 2016), Hospitalet (about 472 Ma; Denèle *et al.* 2009), Canigó (472–462

795 Ma; Cocherie *et al.* 2005; Navidad *et al.* 2018), Roc de Frausa (477–476 Ma; Cocherie *et al.*  
796 2005; Castiñeiras *et al.* 2008) and Albera (about 470 Ma; Liesa *et al.* 2011) massifs (Fig. 8),  
797 which imply a dominant Floian–Dapingian age. Felsic volcanic equivalents have been  
798 documented in the Albera Massif, where subvolcanic rhyolitic porphyroid rocks have yielded  
799 similar ages to those of the main gneissic bodies (about 474–465 Ma; Liesa *et al.* 2011) (Fig.  
800 2).

801 Magmatism persisted until the Late Ordovician, yielding another set of magmatic rocks  
802 that constitute the protoliths of the Cadí (ca. 456 Ma; Casas *et al.* 2010), Casemí (446 to 452  
803 Ma; Casas *et al.* 2010), Núria (ca. 457 Ma; Martínez *et al.* 2011) and Canigó G1-type (ca. 457  
804 Ma; Navidad *et al.* 2018) gneisses. The lowermost part of the Ediacaran Canaveilles Group  
805 hosts metre-scale thick bodies of metadiorite sills related to an Upper Ordovician protolith,  
806 (ca. 453 Ma; Casas *et al.* 2010). Coeval calc-alkaline ignimbrites, andesites and volcanoclastic  
807 rocks are interbedded in the Upper Ordovician succession of the Bruguera and Ribes de  
808 Freser areas (Martí *et al.* 2019). In the latter area, a granitic body with granophyric texture,  
809 dated at ca. 458 Ma by Martínez *et al.* (2011), intruded at the base of the Upper Ordovician  
810 succession. In the La Pallaresa dome, some metre-scale rhyodacitic to dacitic subvolcanic  
811 sills, Late Ordovician in age (ca. 453 Ma; Clariana *et al.* 2018), occur interbedded within the  
812 pre-unconformity strata and close to the base of the Upper Ordovician.

813 In the Axial Zone of the Montagne Noire, some migmatitic orthogneisses, derived from  
814 Ordovician metagranites bearing large K-feldspar phenocrysts, were emplaced at about 471  
815 Ma (Somail orthogneiss; Cocherie *et al.* 2005), 456 to 450 Ma (Pont de-Larn and Gorges  
816 d’Héric gneisses, Roger *et al.* 2004) and ca. 455 Ma (Saint Eutrope Gneiss; Pitra *et al.* 2012)  
817 (Fig. 9). They intruded the “Schistes X” or St-Pons-Cabardès Group, a poorly constrained  
818 Ediacaran stratigraphic succession capped by the Sériès tuff, dated at about 545 Ma (Lescuyer  
819 and Cocherie 1992). The age of migmatization has been inferred from U–Pb dates on

820 monazite from migmatites and anatectic granites at 333 to 327 Ma (Charles *et al.* 2009),  
821 which would represent a Variscan crustal melting event.

822

### 823 *Late Ordovician (post-Sardic) stratigraphic and volcano-sedimentary features*

824

825 Upper Ordovician–Lower Devonian rocks of the southern Montagne Noire and the  
826 Mouthoumet Massif rest paraconformably or with angular discordance on an inherited (pre–  
827 Sardic) Cambrian–Lower Ordovician palaeorelief. Upper Ordovician sedimentation started  
828 along the Cabrières and Mouthoumet rift branches with deposition of basaltic lava flows and  
829 lahar deposits (Roque de Bandies and Villerouge formations) of continental tholeiitic  
830 signature, indicative of continental fracturing (Álvaro *et al.* 2016). Overlying this tholeiitic  
831 volcanic episode, an Upper Ordovician stratigraphic succession can be recognized infilling  
832 these grabens and half-grabens, composed, from bottom to top, of sandstone complexes  
833 (lower Katian Glauzy and Gascagne formations), shales interbedded by carbonate (upper  
834 Katian Gabian and Montjoi formations), and an incisive glaciogenic unconformity,  
835 represented by the Hirnantian Marmaraine diamictites in the Mouthoumet Massif (Fig. 9).  
836 The chronostratigraphic control of this succession is mainly based on brachiopods belonging  
837 to the *Nicolella* and *Hirnantia* communities (Colmenar *et al.* 2013; Álvaro *et al.* 2016).

838 A similar stratigraphic arrangement can be recognized in the eastern Pyrenees, where the  
839 Sardic Phase was succeeded by a Late Ordovician extensional interval responsible for the  
840 opening of (half-)grabens infilled with the basal Upper Ordovician alluvial-to-fluvial  
841 conglomerates (La Rabassa Conglomerate Formation; Hartevelt 1970). Major variations in  
842 the thickness of the Upper Ordovician strata, along with drastic variations in grain size and  
843 thickness, are probably related to the development of palaeotopographies controlled by faults  
844 and the subsequent erosion of uplifted palaeoreliefs. Katian strata are represented by the

845 shale-dominant Cava Formation and the shale/carbonate alternations of the Estana and El  
846 Baell formations (Puđu *et al.* 2018 and references therein). The Hirnantian glaciation is  
847 identified as an erosive unconformity, capped by the Ansovell black shales and the Quartzite  
848 Barr formations, finally sealed by Silurian sediments (Puđu *et al.* 2019).

849

## 850 **The Ordovician of the Maures Massif and Corsica (by Lefebvre)**

851

### 852 *General framework*

853

854 Before the opening of Mediterranean back-arc basins in the Miocene, both Corsica and  
855 Sardinia were adjacent to the Maures Massif in Provence, southern France (Corsini and  
856 Rolland 2009; Schneider *et al.* 2014; Franke *et al.* 2017; Gerbault *et al.* 2018; Álvaro *et al.*  
857 2021; Loi *et al.* this volume). These three regions belong to the southern Variscan belt of  
858 Europe and share a particularly complex geology (Bellot 2005; Gerbault *et al.* 2018).

859

### 860 *Maures Massif*

861

862 The low-grade metamorphic Palaeozoic rocks exposed in the western part of the Maures  
863 Massif (Cap Sicié, Fenouillet, Maurette and Loli units) are generally considered as equivalent  
864 to those occurring in other areas of the external zone of the southern Variscan belt (e.g.  
865 Montagne Noire, Mouthoumet, Pyrenees, southwestern Sardinia) and belonging to the  
866 Gondwanan margin (Bellot 2005; Franke *et al.* 2017; Gerbault *et al.* 2018).

867 In the western part of the Maures Massif, the Palaeozoic succession is poorly preserved,  
868 and the precise age of most units, poorly constrained (Bellot 2005). Most units were assigned



869 to the Silurian and, so far, no evidence of Ordovician rocks could be demonstrated (Gueirard  
870 *et al.* 1970; Paris *et al.* 1999; Bellot 2005; Schneider *et al.* 2014).

871 The central part of the Maures Massif (Collobrières, Bormes, Cap Nègre and Cavalaire  
872 units) consists of highly metamorphosed late Precambrian–lowermost Ordovician volcano-  
873 sedimentary and magmatic rocks, possibly equivalent to the succession occurring in the  
874 Nappe Zone of central Sardinia (Bellot 2005; Gerbault *et al.* 2018).

875 The eastern part of the Maures Massif (Cavalières and Petites Maures units) is  
876 characterized by high-grade (high pressure) Cambro-Ordovician metagranites and Ordovician  
877 metagabbros, suggesting possible affinities to northeastern Sardinia and southwestern Corsica  
878 (Bellot 2005; Faure *et al.* 2014; Schneider *et al.* 2014; Franke *et al.* 2017; Gerbault *et al.*  
879 2018; Álvaro *et al.* 2021; Loi *et al.* this volume).

880

## 881 *Corsica*

882

883 In northwestern Corsica, a non-metamorphic Palaeozoic succession is preserved in the  
884 Argentella area, near Galeria (Baudelot *et al.* 1976; Rossi *et al.* 1995, 2009; Barca *et al.* 1996;  
885 Paris *et al.* 1999; Faure *et al.* 2014; Avigad *et al.* 2018). This succession lies unconformably  
886 on a more than 1,000 m thick basement of Neoproterozoic rocks (amphibolites,  
887 metagreywackes, micaschists and quartzites), polydeformed and metamorphosed during the  
888 Cadomian orogeny (Barca *et al.* 1996; Faure *et al.* 2014; Avigad *et al.* 2018). The overlying  
889 Palaeozoic succession comprises several informal stratigraphic units.

890 The lowermost unit (Ciuttone Sandstone) corresponds to about 50 m thick unfossiliferous,  
891 folded, grey sandstones containing clasts and pebbles reworked from underlying Precambrian  
892 micaschists (Baudelot *et al.* 1976; Barca *et al.* 1996; Avigad *et al.* 2018). A Cambrian to  
893 Early Ordovician age is generally assigned to the Ciuttone Sandstone based on a comparison

894 with the eastern Pyrenees and Sardinia, both characterized by similar sandstone-dominated,  
895 pre-Sardic deposits (Barca *et al.* 1996; Paris *et al.* 1999; Rossi *et al.* 2009). This age is also  
896 congruent with dating obtained from detrital zircons (~555 Ma) suggesting that the Ciuttone  
897 Sandstone was derived from weathered upper Ediacaran rocks (Avigad *et al.* 2018).

898 The upper part of this unit is channeled by the eroding base of the overlying Monte  
899 Martinu Conglomerate, which corresponds to a ~100 m thick succession of conglomerates (at  
900 the bottom) and quartzites (at the top) (Baudelot *et al.* 1976; Barca *et al.* 1996; Avigad *et al.*  
901 2018). Zircons extracted in the upper part of the Monte Martinu Conglomerate yielded an age  
902 of 476 ±26 Ma (Early Ordovician), supporting the interpretation of this unit as consisting of  
903 Upper Ordovician (post-Sardic) deposits (Rossi *et al.* 1995; Avigad *et al.* 2018).

904 The Monte Martinu Conglomerate is tectonically overlain by the ~150 m thick Campu  
905 Orbu Shale. The dark shales of this unit were originally tentatively assigned to the Silurian,  
906 based on poorly preserved acritarchs and chitinozoans (Beaudelot *et al.* 1976). However, the  
907 occurrence of typical Hirnantian diamictites (20–30 m thick) in the overlying Capu Russellu  
908 unit suggests that the age of the Campu Orbu Shale has to be reconsidered and assigned to the  
909 Upper Ordovician (Rossi *et al.* 1995; Paris *et al.* 1999; Avigad *et al.* 2018).

910 The upper part of the Capu Russellu unit consists of 5–6 m of dark shales and lydites  
911 yielding typical assemblages of Rhuddanian and Aeronian (Silurian) graptolites (Štorch 1994;  
912 Barca *et al.* 1996; Paris *et al.* 1999; Avigad *et al.* 2018).

913

## 914 **The Ordovician of Schwartzwald and the Vosges (by Lefebvre and Servais)**

915

### 916 *General framework*

917

918 The Vosges Massif in eastern France and Schwarzwald (Black Forest) in southwestern  
919 Germany are parts of the same geologically complex structure inherited from the Variscan  
920 orogeny and now occurring on the western and eastern shoulders of the Cenozoic Rhine  
921 Graben, respectively (Wickert and Eisbacher 1988; Servais *et al.* 2008). These two isolated  
922 massifs occupy a central position in the Variscan orogen belt between the French Massif  
923 Central and the Bohemian Massif (Vaida *et al.* 2004; Servais *et al.* 2008; Skrzypek *et al.*  
924 2014; Franke *et al.* 2017). The Vosges and Schwarzwald both consist of three similar litho-  
925 tectonic complexes, all intruded by Variscan granitic plutons: (1) a northern complex with  
926 successive belts of (meta-) sedimentary Palaeozoic rocks; (2) a central complex characterized  
927 by HP/HT polymetamorphic gneisses; and (3) a southern complex comprising a mélange of  
928 autochthonous and allochthonous (volcano-) sedimentary Palaeozoic rocks.

929 Since the pioneer work of Kossmat (1927), the northern complex has been traditionally  
930 assigned to the Saxothuringian Zone, whereas Moldanubian affinities were suggested for both  
931 the central and southern complexes (Wickert and Eisbacher 1988; Eisbacher *et al.* 1989;  
932 Fluck *et al.* 1991; Paris *et al.* 1999; Montenari *et al.* 2000; Vaida *et al.* 2004; Servais *et al.*  
933 2008). In this scheme, the Lalaye-Lubine fault zone (LLFZ) in the Vosges and its  
934 prolongation in Schwarzwald, the Baden-Baden fault zone, were interpreted as corresponding  
935 to the former boundary between the subducted Saxothuringian passive margin and the  
936 overlying Moldanubian plate (Wickert and Eisbacher 1988; Eisbacher *et al.* 1989). Recently,  
937 strong lithological affinities with the Teplá-Barrandian Zone (Bohemian Massif) were  
938 established for the northern complexes of the Vosges and Schwarzwald, suggesting that they  
939 were originally located south of the Saxothuringian-Moldanubian suture, and thus not part of  
940 the Saxothuringian Zone (Fig. 1; Skrzypek *et al.* 2014; Álvaro *et al.* 2021).

941

942 *Northern complex*

943

944 The northern complex is best exposed in the Vosges, where it consists of three successive,  
945 fault-bounded units: from North to South, the Bruche, Steige and Villé units (Wickert and  
946 Eisbacher 1988; Skrzypek *et al.* 2014). These three units show an increasing metamorphic  
947 grade towards the LLFZ, which forms the southern boundary of the Northern Vosges  
948 (Wickert and Eisbacher 1988). The Bruche unit is composed of Middle Devonian to lower  
949 Carboniferous unmetamorphosed sedimentary rocks and volcanics (Skrzypek *et al.* 2014).  
950 The Steige unit mainly consists of very low-grade (greenschist facies), red-coloured  
951 metasediments (mainly phyllites and metasandstones) preserving primary bedding structures  
952 and trace fossils (Clauer 1970; Reitz and Wickert 1989). In Andlau and Biarville, the Steige  
953 unit yielded two distinct, relatively diverse, poorly preserved chitinozoan assemblages  
954 (Doubinger 1963; Doubinger and von Eller 1963). The Andlau assemblage includes  
955 *Conochitina brevis*, thus supporting a late Early to early Middle Ordovician age, whereas the  
956 Biarville fauna comprises younger (Silurian) chitinozoans (Doubinger 1963; Doubinger and  
957 von Eller 1963). The structurally deeper Villé unit corresponds to a thick series of low-grade  
958 (greenschist facies) monotonous clastic series of meta-sediments (phyllites, meta-sandstones,  
959 and meta-conglomerates), locally interrupted by a several-meter thick rhyolitic tuff (Reitz and  
960 Wickert 1989). Although primary sedimentary structures are not preserved due to incipient  
961 schistosity, trace fossils (*Tomaculum*) and putative sponge remains were reported in the Villé  
962 unit (Doubinger and von Eller 1963, 1967; Ross 1964). This unit also yielded a poorly  
963 preserved, though diverse acritarch assemblage suggesting a late Cambrian to Early  
964 Ordovician age (Reitz and Wickert 1989).

965 In the Schwarzwald, the northern complex is less extensive and restricted to the Baden-  
966 Baden–Gaggenau area (Eisbacher *et al.* 1989; Servais *et al.* 2008). It comprises a thick  
967 succession of low-grade (greenschist facies) rocks generally considered as equivalent to both

968 the Villé and Steige units (Maas 1981; Montenari *et al.* 2000). This succession is usually  
969 subdivided into four 'series' (Sittig 1965; Montenari *et al.* 2000; Servais *et al.* 2008). The  
970 lowermost one (*Untere Schindelklamm Serie*) mainly consists of quartzitic siltstones and  
971 quartzites. It is overlain by the diabases (meta-volcanics) of the *Basischer Zug*. The third part  
972 (*Obere Schindelklamm Serie*) comprises greywackes, and it is overlain by the uppermost  
973 *Traischbach Serie* formed by a pelitic succession with intercalated carbonate olistoliths. An  
974 extensive palynological sampling through all four series yielded a single, poorly preserved  
975 acritarch assemblage in the uppermost part of the *Traischbach Serie* (Montenari and Servais  
976 2000; Montenari *et al.* 2000). This assemblage comprises the genera *Acanthodiacrodium*,  
977 *Caldariola*, *Cymatiogalea*, *Dasydiacrodium* and *Stelliferidium*, thus suggesting a late  
978 Cambrian to Early Ordovician age, in good accordance with that obtained for the Villé unit in  
979 the Vosges (Montenari and Servais 2000; Montenari *et al.* 2000; Servais *et al.* 2008).

980

### 981 *Central and southern complexes*

982

983 Although the high-grade gneisses of the central complex in the Vosges and Schwarzwald  
984 have not yielded any fossils, an Early Palaeozoic (late Cambrian to Silurian) age for their  
985 sedimentary protoliths could be deduced from the included detrital zircons dated at 490 and  
986 450 Ma (Skrzypek *et al.* 2014; Franke *et al.* 2017).

987 In the Vosges, the southern complex comprises a thick series of autochthonous and  
988 allochthonous Upper Devonian–lower Carboniferous sedimentary rocks intruded by granitic  
989 plutons (Eisbacher *et al.* 1989; Skrzypek *et al.* 2014).

990 In the Badenweiler-Lenzkirch area (Schwarzwald), the southern complex also includes  
991 low-grade meta-sediments (greenschist facies; phyllites) structurally below the  
992 unmetamorphosed Upper Devonian–lower Carboniferous succession. The age of the phyllites

993 from the Badenweiler-Lenzkirch area long remained poorly known, and they were tentatively  
994 assigned to the Devonian (Schäfer 1957; Metz and Rein 1958) or to the lower Carboniferous  
995 (Maas 1961; Krohe and Eisbacher 1988). The occurrence of putative remains of Ordovician  
996 palynomorphs in this unit was first suggested based on poorly preserved material (Hann *et al.*  
997 1995; Sawatzki *et al.* 1997). However, the description of better-preserved palynomorphs  
998 finally confirmed the Early Palaeozoic age of the phyllites (Montenari and Maas 1996;  
999 Montenari *et al.* 2000; Vaida *et al.* 2004; Servais *et al.* 2008). At least two distinct  
1000 fossiliferous lithostratigraphic units were documented in the Badenweiler-Lenzkirch area.  
1001 One consists of meta-greywackes and meta-pelites yielding a Lower–Middle Ordovician  
1002 acritarch assemblage, including the eponymous taxa of the *E. brevis* (upper Floian) and ?*C.*  
1003 *protocalix* (Darriwilian) biozones (Vaida *et al.* 2004). The second level corresponds to meta-  
1004 siltstones containing a relatively diverse palynomorph assemblage (acritarchs, chitinozoans  
1005 and cryptospores) suggesting a Silurian (Wenlock–Ludlow) age (Montenari and Maas 1996;  
1006 Montenari *et al.* 2000; Vaida *et al.* 2004).

1007

## 1008 **Concluding remarks and perspectives**

1009

1010 In spite of a complex Variscan history and overprinting, the original depositional contexts  
1011 and palaeogeographic affinities of the Ordovician strata, now exposed in scattered massifs  
1012 throughout western Europe can be reconstructed, based on various and complementary  
1013 approaches relying e.g. on geodynamics (magmatic affinities, metamorphism, palaeomagnetic  
1014 data and tectonics), palaeontology (biostratigraphy and faunal affinities), and sedimentology  
1015 (stratigraphic gaps and proximal-distal gradients).

1016 In the Ordovician, the regions situated in present-day Belgium and neighbouring areas of  
1017 France and Germany (Brabant Massif, Condroz and Ardenne inliers), western Germany

1018 (Rhenish Massif) and in the subsurface of northeastern Germany (Rügen) were all part of the  
1019 Avalonia microcontinent. In most of these areas, the Ordovician succession is characterized  
1020 by a ~13 myr hiatus extending from the mid Tremadocian and embracing the whole Floian  
1021 (Fig. 3), which is generally interpreted as resulting from the drift of Avalonia away from  
1022 Gondwana (Cocks and Torsvik 2002; Verniers *et al.* 2002; Linnemann *et al.* 2012). This  
1023 interpretation is confirmed by the major and relatively sudden shift in depositional  
1024 environments recorded in the Stavelot-Venn inlier, at the transition between megasequences 1  
1025 and 2 (Fig. 3; Vanguetaine 1992; Verniers *et al.* 2002; Linnemann *et al.* 2012; Herbosch *et*  
1026 *al.* 2016, 2020). In this area, instead of a hiatus, the Lower Ordovician succession records a  
1027 sharp transition from shallow shelf environmental conditions (upper Lierneux Member of the  
1028 Jalhay Formation, upper Tremadocian) to bathyal settings (Otré Formation, Floian). In the  
1029 Brabant Massif deep and often anoxic environmental conditions persisted throughout the Mid  
1030 Ordovician and part of the Late Ordovician, whereas in the Condroz, Rhenish Massif and  
1031 Rügen the depositional environment was shallower on the shelf (megasequence 2, Fig. 3;  
1032 Verniers *et al.* 2002; Servais *et al.* 2008; Linnemann *et al.* 2012). In most areas, the transition  
1033 between megasequences 2 and 3 around mid-Katian times (Fig. 3) marks a shift from deep  
1034 and/or outer shelf settings to shallower environments and even emersion in the Condroz, very  
1035 likely related to the initial stages of the soft docking of eastern Avalonia with Baltica (e.g.  
1036 Verniers *et al.* 2002; Linnemann *et al.* 2012). The three megasequences documented in  
1037 Belgium, western and northeastern Germany are comparable to those originally described  
1038 from the Welsh Basin and the English Lake District (Woodcock 1990; Molyneux *et al.* this  
1039 volume), thus supporting close palaeogeographic Avalonian affinities for all these regions  
1040 (Katzung *et al.* 1995; Maletz, 2000; Verniers *et al.* 2002; Servais *et al.* 2008; Linnemann *et*  
1041 *al.* 2012).

1042 With only exception of the southern part of the Ardenne inliers, all Ordovician rocks  
1043 exposed in present-day France and western Germany (Schwarzwald) were originally  
1044 deposited on the southern margin of the Rheic Ocean, on the Gondwanan passive margin or  
1045 nearby (Fig. 1; Robardet *et al.* 1994b; Paris *et al.* 1999; Servais and Sintubin 2009; Pouclet *et*  
1046 *al.* 2017; Álvaro *et al.* 2021; Caroff, in press). However, reconstructing their original  
1047 depositional setting, relative position and relationship with other (peri-)Gondwanan regions  
1048 remains difficult. A close palaeogeographic link between the Armorican Massif and the  
1049 Iberian Peninsula has been long established based on strong lithological and faunistic  
1050 similarities (e.g. Paris and Robardet 1977, 1994; Robardet and Gutiérrez-Marco 1990;  
1051 Robardet *et al.* 1994b). Probable affinities between these two regions are further supported by  
1052 the occurrence of a similar Furongian–mid Floian hiatus (Fig. 6). Designated the "lacune  
1053 normande" in the Armorican Massif (Le Corre *et al.* 1991) and "Toledonian gap" in the  
1054 Iberian Massif (Alvaro *et al.* 2021), this ~22 myr long hiatus possibly resulted, in both areas,  
1055 from the uplift of Ediacaran–Cambrian rocks (Sánchez-García *et al.* 2019). This "Norman  
1056 hiatus" has no equivalent in Corsica, the Occitan Domain (Montagne Noire and Mouthoumet  
1057 massifs), eastern Pyrenees and in the Vosges-Schwarzwald massifs.

1058 In marked contrast with the close proximity of the Armorican Massif and the Iberian  
1059 Massif, continuous shallow shelf sedimentation is recorded in the Lower Ordovician  
1060 successions of northwestern Corsica (Avigad *et al.* 2018), the northern and southern  
1061 Montagne Noire (Fig. 9; Vizcaïno *et al.* 2001), the Mouthoumet Massif (Colmenar *et al.*  
1062 2013), as well as in other areas (e.g. Maures Massif and Western Alps), yielding less well-  
1063 preserved, metamorphosed rocks (Bellot 2005; Guillot and Ménot 2009). Moreover, the  
1064 Ordovician successions of Corsica and all regions of the Occitan Domain are characterized by  
1065 a major hiatus extending from the Dapingian to the early Katian boundary (Fig. 9). Several  
1066 authors (Pouclet *et al.* 2017; Alvaro *et al.* 2021) proposed to link this stratigraphic gap to the



1067 Sardinic phase, originally described in the Sulcis-Iglesiente area of southwestern Sardinia (see  
1068 Loi *et al.* this volume). The occurrence of a similar stratigraphic unconformity (mirroring the  
1069 Sardinic unconformity) in northwestern Corsica, the Occitan Domain, eastern Pyrenees and  
1070 southwestern Sardinia thus strongly suggests that all these areas were geodynamically linked  
1071 and palaeogeographically close in the Ordovician. In marked contrast with the situation in  
1072 these areas, a continuous shelf sedimentation is recorded in the Middle and Upper Ordovician  
1073 successions of the Armorican Massif (Fig. 6).

1074 It is thus possible to identify two main "clusters" of Ordovician rocks in France. The first  
1075 one, which comprises the Armorican Massif, is characterized by strong affinities with the  
1076 Iberian Massif, a Furongian–mid Floian "Norman gap", and a continuous shelf sedimentation  
1077 in the Mid and Late Ordovician (Fig. 6). The second cluster, which includes northwestern  
1078 Corsica, the Montagne Noire, the Mouthoumet Massif, the eastern Pyrenees and possibly the  
1079 Maures Massif, can be identified based on strong affinities with Sardinia (Sulcis-Iglesiente  
1080 unit, Sarrabus and Gerrei units), a continuous shelf sedimentation during the Early  
1081 Ordovician, and an extensive Dapingian–early Katian Sardinic(?) gap (Fig. 9). However, in the  
1082 late Katian, regions belonging to both clusters are characterized by a similar shift from  
1083 siliciclastic to carbonate sedimentation, and a major faunal transition from cool-water adapted  
1084 assemblages to more temperate ones (see e.g. Dégardin *et al.* 1995; Colmenar *et al.* 2013;  
1085 Lefebvre *et al.* 2022). This major environmental change, which is also recorded in other high-  
1086 latitude (peri-)Gondwanan regions (e.g. Algeria, Carnic Alps, Iberian Peninsula, eastern  
1087 Tafilalt area in Morocco and Sardinia), probably results from a short episode of global  
1088 warming: the "Boda Event" (Villas *et al.* 2002; Fortey and Cocks 2005; Ferretti *et al.* this  
1089 volume).

1090 In the Vosges and Schwarzwald massifs, Ordovician rocks are too poorly preserved to  
1091 establish any clear relationship with better-known successions from the Armorican-Iberian or

1092 the Occitan-Sardinian clusters. In the northern complex of the Vosges-Schwarzwald massifs,  
1093 the identification of Furongian–Early Ordovician acritarch assemblages in both the Villé  
1094 Formation and the Traischbach Serie indicates, at least, the absence of a "Norman gap" in  
1095 these levels. Similarly, the occurrence of typical late Floian and Darriwilian chitinozoans in  
1096 the Badenweiler-Lenzkirch area suggests that no "Sardic gap" occurs in the Ordovician  
1097 succession of southern Schwarzwald. Correlation of these two isolated massifs with other  
1098 Variscan areas thus remains problematic, and future work will probably help clarify if the  
1099 northern complex of the Vosges-Schwarzwald massifs belongs to the Saxothuringian Zone  
1100 (Wickert and Eisbacher 1988; Eisbacher *et al.* 1989) or to the Teplá-Barrandian Zone  
1101 (Skrzypek *et al.* 2014; Álvaro *et al.* 2021). Moldanubian affinities for the central and southern  
1102 Vosges-Schwarzwald massifs seem to be better established (Eisbacher *et al.* 1989; Skrzypek  
1103 *et al.* 2014).

1104 Other West-European areas, not treated here, are in a serious need of reevaluation of their  
1105 Ordovician rocks. One of the most extensive Ordovician successions is probably the  
1106 particularly thick one occurring in the subsurface of the Aquitaine Basin (Paris and Le Pochat,  
1107 1994; Robardet *et al.*, 1994b). Preliminary investigations of boreholes suggests the occurrence  
1108 of over 1,200 m of siliciclastic rocks dated from the Lower and Middle Ordovician, based on  
1109 acritarchs, chitinozoans, graptolites, and trilobites. The presence of Upper Ordovician rocks is  
1110 less poorly constrained, but typical Hirnantian diamictites are present (Robardet *et al.* 1994b).  
1111 Other poorly known regions in serious need of a reevaluation of their Ordovician rocks  
1112 include e.g. the central Pyrenees (see e.g. Robardet *et al.* 1994b; Dégardin *et al.* 1995), the  
1113 southeastern-most extremity of the SAD (Chantonay and Vendée areas; see Pouclet *et al.*  
1114 2017) or the meta-sediments of the (French) western Alps (Guillot and Ménot 2009).  
1115 However, the Montagne Noire is an iconic example illustrating that, even in better-known and  
1116 "classical" areas, much work still remains to be done. In this region, the trilobite-based

1117 biostratigraphic framework (Fig. 9) still needs to be carefully correlated with more "standard"  
1118 regional or international biostratigraphic charts based on acritarchs, chitinozoans, conodonts  
1119 or graptolites.

1120

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1122

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1135

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2045

2046

## 2047 **Figure captions**

2048

2049 **Fig. 1.** Tectonostratigraphic subdivision of the Variscan Orogen in southwestern Europe with  
2050 pre-Variscan setting of the Pyrenees between the Montagne Noire and Sardinia, and: (1)  
2051 Variscan parautochthon; (2) Moldanubian Zone and its prolongation; (3) Teplá-Barrandian  
2052 Zone and its prolongations; (4) Saxothuringian Zone; (5) Rheno-Hercynian Zone; (6)  
2053 Avalonian parautochthon; (7) oceanic sutures; Abbreviations: AM, Armorican Massif; Ard,  
2054 Ardenne; BM, Brabant Massif; Co, Corsica; EP, Eastern Pyrenees; IS; Iapetus suture; Mau,  
2055 Maures Massif; MN, Montagne Noire; Mou: Mouthoumet Massif; RM: Rhenish Massif; RS:  
2056 Rheic suture; Rü: Rügen; SW: Schwarzwald; TS: Thor suture; Vo: Vosges. Adapted and  
2057 modified from Ballèvre *et al.* (2009), Torsvik and Cocks (2013), Skrzypek *et al.* (2014), and  
2058 Álvaro *et al.* (2021).

2059

2060 **Fig. 2.** Simplified geological map of the southern part of Belgium and neighbouring countries  
2061 showing the main tectonostratigraphic units of the Palaeozoic. To the North, the Brabant  
2062 Parautochthon and/or the Haine-Sambre-Meuse Overturned Thrust sheets are separated from  
2063 the Ardenne Allochthon by the Midi-Aachen Thrust Fault. The four Ardenne inliers, Stavelot-

2064 Venn, Rocroi, Givonne and Serpont, are represented in purple. Modified from Herbosch *et al.*  
2065 (2020).

2066

2067 **Fig. 3.** Ordovician basement stratigraphy of Belgium and neighbouring countries.  
2068 Chronostratigraphy, graptolite and chitinozoan biozones after Goldman *et al.* (2020, fig. 20.2–  
2069 4, column Baltica for chitinozoan and Britain for graptolite zonations). References are  
2070 provided in the text for the bio- and chronostratigraphic position of the formations belonging  
2071 to the Brabant Massif, the Condroz, Rocroi and Stavelot-Venn inliers. Dotted line: boundary  
2072 between formations less precise. Lithostratigraphic units indicated in plain style are members,  
2073 and those in bold, formations. Abbreviations: Ser., Series; Meg., Megasequences.

2074

2075 **Fig. 4.** Ordovician stratigraphy of the Ebbe Inlier (Rhenish Massif, western Germany) and the  
2076 subsurface of Rügen Island (northeastern Germany). Chronostratigraphy, graptolite and  
2077 chitinozoan biozones after Goldman *et al.* (2020, fig. 20.2–4, column Baltica for chitinozoan  
2078 and Britain for graptolite zonations). References are provided in the text for the bio- and  
2079 chronostratigraphic position of the formations. Dotted line: boundary between formations less  
2080 precise. All lithostratigraphic units reported (in bold) are formations. Abbreviation: Ser.,  
2081 Series.

2082

2083 **Fig. 5.** Simplified geological map of the Armorican Massif (western France) illustrating its  
2084 structuration into four distinct domains separated by two main late Carboniferous (Variscan)  
2085 shear zones: the North Armorican Shear Zone (NASZ) and the South Armorican Shear Zone  
2086 (SASZ). All main areas yielding Ordovician rocks mentioned in the text and/or reported on  
2087 Fig. 6 (chronostratigraphic chart) are indicated on the map by numbers, with: (1) Cotentin; (2)  
2088 Zone bocaine; (3) Domfront; (4) Laval; (5) Menez-Belair; (6) Châteaulin; (7) Crozon; (8)

2089 South of Rennes; (9) Angers; (10) Saint-Julien-de-Vouvantes; (11) Saint-Georges-sur-Loire;  
2090 (12) Ancenis. Modified from Henry (1980), Ballèvre *et al.* (2009), Vidal *et al.* (2011b), and  
2091 Paris (2016).

2092

2093 **Fig. 6.** Ordovician stratigraphy of the Armorican Massif (western France).  
2094 Chronostratigraphy and high latitude Gondwanan chitinozoan biozones after Goldman *et al.*  
2095 (2020). References are provided in the text for the bio- and chronostratigraphic position of the  
2096 formations belonging to the NAD (Domfront), the MAD (Laval, Crozon, South of Rennes,  
2097 Angers) and the SAD (Ancenis). Dotted line: boundary between formations less precise.  
2098 Lithostratigraphic units indicated in plain style are members, and those in bold, formations.  
2099 Abbreviation: Ser., Series.

2100

2101 **Fig. 7.** Sequence stratigraphy of the Postolonnec Formation, Crozon, western part of the  
2102 Median Armorican Domain (from Dabard *et al.* 2015). **(A)** Lithostratigraphic column with  
2103 depositional environments deduced from sedimentological features and Gamma-ray data.  
2104 Very high frequency (purple and red curves), high frequency (blue curve) and low frequency  
2105 (pale-blue curve) sequences. **(B)** Time-linear sea-level curve calibrated from the graptolites  
2106 biozones (i.e. *D. artus* at the top of the Kerloc'h Member and *G. linnarsoni* in Morgat  
2107 Member). On this figure the report of the chitinozoan biozones highlights the long-time  
2108 duration of the *L. pissotensis* Biozone, as recorded in the Crozon succession. Abbreviations:  
2109 Fm., Formation; L. Off., Lower Offshore; Mb., Member.

2110

2111 **Fig. 8.** Simplified geological map of the Occitan Domain (Albigeois, Montagne Noire,  
2112 Mouthoumet Massif), and eastern Pyrenees in southern France and neighbouring countries,  
2113 with: (1) Albera (Albères) Massif; (2) Roc de Frausa (Roc de France) Massif; (3) Canigó

2114 (Canigou) Massif; (4) Hospitalet Massif; (5) Aston Massif. Modified from Touzeau *et al.*  
2115 (2012), Pouclet *et al.* (2017), and Padel *et al.* (2018). Other regions of the Occitan Domain  
2116 mentioned in the text are not shown here, because they are lying further North (i.e. Thiviers-  
2117 Payzac and Génis, Rouergue, and southern Cévennes massifs).

2118

2119 **Fig. 9.** Ordovician stratigraphy of the Montagne (southern France). Chronostratigraphy after  
2120 Goldman *et al.* (2020), and trilobite biostratigraphy based on Vizcaïno and Álvaro (2003),  
2121 Tortello *et al.* (2006), Serpagli *et al.* (2007), and Colmenar *et al.* (2013). References are  
2122 provided in the text for the bio- and chronostratigraphic position of the formations. Dotted  
2123 line: boundary between formations less precise. Key to trilobite generic names: *Proteuloma*  
2124 *geinitzi*, *Shumardia (C.) pusilla*, *Euloma filacovi*, *Taihungshania miqueli*, *Thaihungshania*  
2125 *shui landeyranensis*, *Colpocoryphe maynardensis*, *Neseuretus (N.) arenosus*, *Apatokephalus*  
2126 *incisus*, *Hangchungolithus primitivus*, and *Calymenella boisseli*. Lithostratigraphic units  
2127 indicated in plain style are members, and those in bold, formations. Abbreviation: Ser., Series.

1 The Ordovician from France and neighbouring areas of Belgium and  
2 Germany

3  
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25

26 Abbreviated title: Ordovician of France and neighbouring areas ~~of Belgium and Germany~~

27

28 **Abstract:** The Ordovician successions of France and neighbouring areas of Belgium and  
29 Germany are reviewed and correlated based on international chronostratigraphic and regional  
30 biostratigraphic charts. The same three megasequences related to the rift, drift, and docking of  
31 Avalonia with Baltica can be ~~evidenced~~ tracked in Belgium and neighbouring areas (Brabant  
32 Massif, ~~and~~ Ardenne inliers), western (Rhenish Massif) and northeastern Germany (Rügen).  
33 The remaining investigated areas were part of Gondwana in the Ordovician ~~times~~. The  
34 Armorican Massif shares with the Iberian Peninsula a Furongian–Early Ordovician gap  
35 (Toledanian or Norman gap), and a continuous Mid–Late Ordovician shelf sedimentation.  
36 The Occitan Domain (Montagne Noire and Mouthoumet massifs), eastern Pyrenees and  
37 northwestern Corsica ~~are sharing~~ with southwestern Sardinia ~~a~~ continuous shelf  
38 sedimentation in the Early Ordovician ~~times~~, and a Mid Ordovician "Sardic gap". In the  
39 Ordovician ~~times~~, the Maures Massif probably belonged to the same Sardo-Occitan domain.  
40 The Vosges and Schwarzwald massifs display comparable, poorly preserved Ordovician  
41 successions, suggesting affinities with the Teplá-Barrandian and/or Moldanubian zones of  
42 Central Europe.

43

44

45 In western Europe, Ordovician sedimentary rocks generally occur as scattered, disconnected  
46 massifs or inliers surrounded and/or hidden by Mesozoic and/or Cenozoic ~~rock~~ cover.  
47 Reconstructing their original layout, stratigraphy and palaeogeographic relationships with  
48 other regions has been a major challenge during the last decades. All Ordovician strata  
49 occurring in present-day Belgium, France and Germany have undergone a long and complex  
50 journey. They were all deposited on a Cadomian or Cambrian basement with clear

51 Gondwanan affinities. In [the](#) Furongian–Early Ordovician–~~times~~, the onset and diachronic  
52 propagation, from present-day southwest to northeast, of a major rift along the western  
53 margin of Gondwana gave rise to the Rheic Ocean (e.g. Linnemann *et al.* 2008; von Raumer  
54 and Stampfli 2008; Cocks and Fortey 2009; Stampfli *et al.* 2011). The opening of the Rheic  
55 Ocean resulted in the rifting of Avalonia, as an independent microcontinent, away from  
56 Gondwana (e.g. Prigmore *et al.* 1997; Verniers *et al.* 2002; Cocks and Fortey 2009; Servais  
57 and Sintubin 2009; Domeier 2016). In [the](#) Ordovician–~~times~~, several areas included in this  
58 chapter were thus belonging to the eastern part of Avalonia: the Brabant Massif and Ardenne  
59 inliers in Belgium, the Rhenish Massif in western Germany, and Rügen Island, in northeastern  
60 Germany (Fig. 1). In contrast, the Armorican Massif, the Occitan Domain (Montagne Noire  
61 and Mouthoumet massifs), eastern Pyrenees, the Maures Massif, Corsica, and the Vosges (in  
62 France) and Schwarzwald (in western Germany) were located on the opposite (southern) side  
63 of the Rheic ocean, still forming part of Gondwana (Fig. 1). Avalonia drifted northwards,  
64 away from Gondwana, before docking with Baltica in [the](#) Late Ordovician ~~times~~–(closure of  
65 the Tornquist Ocean; TS in Fig. 1), and slightly later with Laurentia, in the Silurian (closure  
66 of the Iapetus Ocean; IS in Fig. 1), resulting in the Caledonian orogeny (e.g. Verniers *et al.*  
67 2002; Cocks and Fortey 2009).

68 In [the](#) late Palaeozoic–~~times~~, the closure of the Rheic Ocean led to the formation of an  
69 extensive orogenic belt extending from the Appalachian and Ouachita mountains (to the  
70 West) to the Caucasus (to the East). This major continental collision or "Variscan orogeny"  
71 between Laurentia–Avalonia–Baltica (to the North) and Gondwana (to the South) was  
72 particularly complex, because of the occurrence of numerous terranes, related to back-arc  
73 basins and narrow oceanic domains on both sides of the closing Rheic Ocean (Pharaoh 1999;  
74 Matte 2001; Ballèvre *et al.* 2009; Faure *et al.* 2009; Guillot and Ménot 2009; Stampfli *et al.*  
75 2011; von Raumer *et al.* 2013). A consequence of the Variscan "collage" was to bring

76 together, on both sides of the Rheic suture (RS on Fig. 1), most pieces of the jigsaw forming  
77 present-day Europe, and extending from the Iberian Massif (see Gutiérrez-Marco *et al.* 2002)  
78 to the Alps (Ferretti *et al.* this volume) and the Bohemian Massif (Kraft *et al.* this volume).  
79 With the exception of the Brabant Massif, all Ordovician rocks now exposed in western  
80 Europe were more or less ~~severely~~ affected by the Variscan orogeny. For example, in some  
81 areas (e.g. Maures Massif, central zones of the Vosges and Schwarzwald), they endured  
82 medium to high-grade metamorphism (Faure *et al.* 2014; Skrzypek *et al.* 2014; Gerbault *et al.*  
83 2018). Even in regions exposing Ordovician rocks with very low to low-grade metamorphism,  
84 the strata were more or less affected by Variscan tectonics, ~~implying-involving~~ folds and/or  
85 displacements along several tens of kilometers along shear zones or thrusts (e.g. Ballèvre *et*  
86 *al.* 2009; Pouclet *et al.* 2017; Herbosch *et al.* 2020). Moreover, in some areas (e.g. Alps,  
87 eastern Pyrenees), Variscan structures were reactivated and/or overprinted during the Alpine  
88 orogeny, thus making their interpretation even more ~~complex~~difficult.

89 During the last 30 years, the major advances in biostratigraphy (relying mostly on  
90 acritarchs, chitinozoans and graptolites, less frequently on brachiopods, conodonts and  
91 trilobites) have greatly contributed to the understanding of the complex pre-Variscan history  
92 of western Europe, leading to a wide (and sometimes contradictory) set of palaeogeographic  
93 scenarios and regional syntheses, as e.g. in the Armorican Massif (Le Corre *et al.* 1991; Paris  
94 and Robardet 1994; Robardet *et al.* 1994a; Ballèvre *et al.* 2009), Belgium (Verniers *et al.*  
95 2002; Linnemann *et al.* 2012; Herbosch *et al.* 2020), Corsica (Faure *et al.* 2014; Avigad *et al.*  
96 2018), the Ebbe inlier (e.g. Maletz 2000; Koch *et al.* 2014), the Maures Massif (Bellot 2005),  
97 the Occitan Domain (Dégardin *et al.* 1995; Laumonier *et al.* 1995; Pouclet *et al.* 2017), or the  
98 Vosges (Skrzypek *et al.* 2014). ~~Few~~A few more integrative reviews have been produced, and  
99 include syntheses on the Ordovician of France (Paris *et al.* 1999; Nardin *et al.* 2014), Belgium  
100 and France (Robardet *et al.* 1994b), or Belgium and Germany (Verniers *et al.* 2002; Servais *et*



101 *al.* 2008). The aim of the chapter is thus to provide an updated overview of the Ordovician of  
102 Belgium, France, and both western and northeastern Germany through a small number of  
103 selected regions, and to correlate their sedimentary successions with both global subdivisions  
104 (series ~~and~~ stages) of the International Chronostratigraphic Chart and the biostratigraphic  
105 time scales elaborated by Goldman *et al.* (2020) for the Ordovician.

106

## 107 **The Ordovician of Belgium and neighbouring areas (by Herbosch, Verniers** 108 **and Servais)**

109

### 110 *General framework*

111

112 Ordovician strata exposed in the Caledonian basement of Belgium and neighbouring areas ~~can~~  
113 ~~be observed in~~include two major units: those formed by the Brabant Massif (BM), the  
114 Brabant Parautochthon and the Haine-Sambre-Meuse Overturned Thrust Sheets (HSM-OTS)  
115 to the North, and ~~that by~~ the Ardenne Allochthon to the South (Fig. 2). The Midi-Eifel Thrust  
116 Fault displaced the Ardenne Allochthon several tens of km to the North (Hance *et al.* 1999;  
117 Mansy and Lacquement 2002) during the Variscan orogeny at the end of the Carboniferous.  
118 The Ordovician crops out also in the Condroz Inlier, a narrow strip along the Midi-Eifel Fault  
119 (Fig. 2). In the Ardenne Allochthon, the basement is exposed in four inliers: Stavelot-Venn,  
120 Rocroi, Givonne and Serpont. This Caledonian basement belongs to the Avalonia  
121 microcontinent (Fig. 1; Cocks and Torsvik 2002; Verniers *et al.* 2002). Overall, the  
122 sedimentation ~~of~~ during the Ordovician is terrigenous and deposited in a deep marine anoxic  
123 environment, which explains the scarcity of benthic macrofossils. As a result it is only since  
124 the 1970s that researches with graptolites, acritarchs and chitinozoans have allowed ~~to~~ the  
125 establishment of a reliable bio- and chronostratigraphy.

126

127 *The Brabant Massif*

128

129 The Brabant Massif (BM) consists of a largely concealed, NW-SE trending, low-grade  
130 metamorphic slate belt developed during [the](#) early Palaeozoic ~~times~~, in the sub-surface of  
131 northern Belgium (Verniers *et al.* 2002; Debacker *et al.* 2005). Along its southern edge, river  
132 incisions provide narrow outcrop areas through the cover (Fig. 2). To the South, the Brabant  
133 Parautochthon and/or the HSMS-OTS (Belanger *et al.* 2012) are overthrust by the Ardenne  
134 Allochthon along the Midi-Eifel Fault, which represents the Variscan Thrust Front (Fig. 2).  
135 The BM consists of a very thick siliciclastic sequence, ranging from the lower Cambrian in  
136 the core to the upper Silurian along the rims. The sedimentary record is continuous, except [for](#)  
137 a hiatus from the middle Tremadocian to the Dapingian (Fig. 3; Herbosch *et al.* 2008;  
138 Herbosch and Verniers 2014). The total thickness exceeds 13 km, with less than 1.8 km for  
139 the Ordovician (Linnemann *et al.* 2012, fig. 7).

140 The three sedimentary megasequences bounded by basin-wide unconformities recognized  
141 in the Welsh basin by Woodcock (1990) have also been recognized in the BM and the entire  
142 basement (Fig. 3; Woodcock 1991; Vanguetaine 1992; Verniers *et al.* 2002; Herbosch *et al.*  
143 2020). The Ordovician of the BM begins with the lower Tremadocian black slates of the  
144 uppermost part of the Mousty Formation dated by the graptolite *Rhabdinopora praeparabola*  
145 (Wang and Servais 2015). This formation passes gradually to the Chevlipont Formation  
146 formed by wavy bedded grey siltstones. It is dated by the graptolites *Rhabdinopora parabola*  
147 and *R. anglica* (Wang and Servais 2015) from about the lower half of the Tremadocian, i.e.  
148 from 485.5 to 482.7 Ma (Goldmann *et al.* 2020). Then, a stratigraphic hiatus of about 13  
149 [m.y.myr](#) precedes the deposition of the Abbaye de Villers Formation (Fig. 3). It is interpreted

150 as linked to the drifting of the Avalonia microplate from Gondwana and the opening of the  
151 Rheic Ocean (Cocks and Torsvik 2002; Verniers *et al.* 2002; Linnemann *et al.* 2012).

152 Megasequence 2 begins with the Abbaye de Villers and Tribotte formations consisting of  
153 silty slates (Herbosch and Verniers 2014). They are ~~dated~~correlated from with the upper  
154 Dapingian to the ~~early~~lower Darriwilian by acritarchs, which belong to the *F. hamata*-*S.*  
155 *rarrigulata* Biozone (Vanguetaine and Wauthoz 2011). This acritarch biozone can be  
156 correlated with the *I. gibberulus* and *A. cucullus* graptolite biozones (Cooper and Molyneux  
157 1990). The overlying Rigenée Formation, formed by dark silty slate, records a regional  
158 transgression (Paris *et al.* 2007) and is dated by graptolites (*D. artus* and *D. murchinsoni*;  
159 Maletz and Servais 1998) and acritarchs (*F. sartbernardensis*, *A. virgata*, *F. hamulata*;  
160 Servais 1991, 1993) to most of the Darriwilian (Herbosch and Verniers 2014, fig. 3). The Ittre  
161 Formation is formed by an alternation of fine sandstones, siltstones and slates interpreted as  
162 distal turbidites (Servais 1991). It was dated by graptolites (*N. gracilis* and *M. foliaceus*;  
163 Maletz and Servais 1998) and chitinozoans (*S. cervicornis* Biozone; Vanmeirhaeghe 2006) ~~to~~  
164 ~~the~~as Sandbian. The Bornival Formation essentially formed by mudstones was ~~dated~~  
165 correlated by chitinozoans (no index species) ~~to~~with the lowest Katian (Vanmeirhaeghe  
166 2006). The Cimetière de Grand-Manil Formation was not biostratigraphically dated. The Huet  
167 Formation is formed by greenish slates and siltstones with shelly facies level (e.g.  
168 brachiopods, bryozoans, corals and; crinoids; Mailleux 1926) interpreted as distal tempestites.  
169 It was dated by chitinozoans: *T. bergstroemi* (Van Grootel *et al.* 1997) and *F. spinifera*  
170 biozones (Vanmeirhaeghe 2006) ~~to~~and correlated with the middle Katian (Herbosch and  
171 Verniers 2014, fig. 3). The Fauquez Formation ~~shows~~includes graptolitic black slates  
172 interpreted as very distal turbidites deposited in a deep marine environment. It was ~~dated~~  
173 correlated by graptolites (*P. linearis*; Maletz and Servais 1998) and chitinozoans (*F. spinifera*;  
174 Vanmeirhaeghe, 2006) ~~to~~with the middle Katian.

175 Megasequence 3 begins with the Madot Formation showing many volcanic and volcano-  
176 sedimentary rocks interstratified with siltstones rich in shelly debris (Verniers *et al.* 2005)  
177 deposited in a shallow shelf. The transition from the Fauquez to the Madot Formation marks  
178 an abrupt and important change in bathymetry. The shallowing could be linked to the docking  
179 of Avalonia with Baltica (Cocks and Torsvik 2005; Torsvik and Cocks 2011; Linnemann *et*  
180 *al.* 2012) and was ~~dated~~correlated by the chitinozoans *T. bergstroemi* (Vanmeirhaeghe *et al.*  
181 2005) and *C. rugata* (Mortier 2014) ~~to~~with the upper half of the Katian (Fig. 3). The lower  
182 Gouteux Member of the Brutia Formation ~~shows~~includes grey bioturbated slates and was  
183 assigned by the chitinozoans *G. gamachiana* (Samuelsson and Verniers 2000) and *S. oulebsiri*  
184 (Mortier 2014) to the uppermost Katian ~~to the~~ Hirnantian and extends most probably even  
185 into the Silurian.

186

### 187 *The Condroz Inlier*

188

189 The Condroz Inlier (CI) is a narrow strip about 65 km long and 0.5 to 4 km wide composed of  
190 Ordovician and Silurian siliciclastic rocks, emplaced mostly as a series of tectonic blocks  
191 along the Midi-Eifel Fault Zone in the HSM-OTS (Fig. 2). The inlier is flanked tectonically to  
192 the North by Middle Devonian rocks and to the South by Lower Devonian rocks  
193 (Vanmeirhaeghe 2006). Given its position within the strongly faulted Variscan front zone and  
194 with a very poor degree of exposure, the stratigraphy of the CI is extremely difficult to  
195 establish and not fully resolved. In the main Central CI part, the sedimentary record extends  
196 from the base of the Ordovician to the upper part of the Silurian with an approximative  
197 thickness of over 1.1 km for the Ordovician (Vanmeirhaeghe 2006).

198 Megasequence 1 is reduced to the Chevlipont Formation observed in the Wépion borehole  
199 (Graulich 1961). Megasequence 2 begins with a 10 cm thick basal conglomerate, followed by

200 the upper slope dark mudstones of the Huy Formation ~~dated~~correlated by chitinozoans (*S.*  
201 *formosa* Biozone) ~~to~~with the middle Darriwilian (Vanmeirhaeghe 2006). The unconformity is  
202 well constrained, with a hiatus from the middle Tremadocian to the lower Darriwilian  
203 (Vanmeirhaeghe 2006; Wang and Servais 2015), i.e. a time interval close to that observed in  
204 the BM (Fig. 3). After an observational gap, the sedimentation continued with the Chevreuil  
205 and the Vitrival-Bruyère formations deposited on the outer shelf to upper slope. The  
206 Chevreuil Formation was dated by chitinozoans (typical association, no index species) ~~to~~  
207 theas late Darriwilian to ~~the~~early Sandbian (Vanmeirhaeghe 2006). The Vitrival-Bruyère  
208 Formation is subdivided into four members: the Giroux Member is dated by graptolites (*N.*  
209 *gracilis* Biozone) and chitinozoans (*L. stentor* Biozone) ~~to~~theas lower Sandbian, the poorly  
210 dated La Bruyère Member, the Sart-Bernard and the upper part of the Rue de Courrière  
211 members, both ~~dated~~correlated by chitinozoans (*S. cervicornis* Biozone) ~~to~~with the lower  
212 Katian (Vanmeirhaeghe 2006, fig. 47). The uppermost Rue de Courrière Member is  
213 composed of siltstones with shell beds, sandstones and microconglomerates, which mark the  
214 base of a short stratigraphic hiatus spanning the Onnian (mid-dle Katian) and points to the  
215 occurrence of a para-conformity. The repercussion of the drastic change of bathymetry  
216 observed in the BM at the end of Megasequence 2 (see above) lead in the Central CI to the  
217 emergence, as this region is globally shallower than the BM (Vanmeirhaeghe 2006, p. 202;  
218 Herbosch and Verniers 2014, fig. 11).

219 Megasequence 3 starts with the Fosses Formation, which is composed of shelf deposits and  
220 is subdivided into three members (Fig. 3). The two lower units were dated by chitinozoans:  
221 the Bois de Presles Member (*T. bergstroemi* and *C. rugata* biozones) ~~to~~theis middle to late  
222 Katian in age, and the Faux-les-Tombes Member (*B. umbilicata* Biozone), ~~to~~the  
223 uppermostlatest Katian (Vanmeirhaeghe 2006). The upper Tihange Member is ~~dated~~  
224 correlated with~~to~~ the Hirnantian by ~~deduction~~inference, as the overlying Bonne-Espérance

225 Formation ~~has-is an~~ early Rhuddanian in age (Vanmeirhaeghe 2006). However, the study of  
226 several fossil levels (Pereira *et al.* 2021, fig. 2) demonstrates that the upper half of the  
227 Tihange Member belongs to the *A. ascensus* and *P. acuminatus* graptolite biozones indicating  
228 an early Rhuddanian age (Silurian).

229

### 230 *The Stavelot-Venn Inlier*

231

232 The Stavelot-Venn Inlier (SVI) is located in the NE of the Ardenne Allochthon straddling the  
233 German border (Fig. 2). It is the inlier closest to the Variscan front and it was transported  
234 about 10 km northwards during the thrusting of the Ardenne Allochthon (Hance *et al.* 1999).  
235 It has undergone Variscan ~~orogeny~~orogenesis, but the presence of an earlier Caledonian  
236 tectonic event is still a matter of debate (Piessens and Sintubin 1997; Sintubin *et al.* 2009;  
237 Herbosch *et al.* 2020). The SVI shows a continuous, mainly terrigenous, sedimentation  
238 ranging from the early Cambrian to the Middle Ordovician.

239 The Ordovician comprises three formations: the Jalhay, Ottré and Bihain formations,  
240 together about 1.2 km thick. A detailed sedimentological study of the Jalhay Formation (Fig.  
241 2; Lamens 1985) has shown a regressive sequence documenting successively the basin plain  
242 (Solwaster Member), the slope (Spa Member) and finally the upper part of the slope  
243 (Lierneux Member). The two lower members have been dated by graptolites of the *R.*  
244 *praeparabola* to *R. flabelliformis anglica* biozones to the ~~early-lower~~ Tremadocian (Wang  
245 and Servais 2015). The Lieerneux Member contains acritarchs of the informal Zone 8  
246 (Vanguetaine 1992) corresponding to the upper part of the Tremadocian (Herbosch 2021).  
247 Hence, the Jalhay Formation spans most of the Tremadocian (Fig. 3).

248 The Ottré Formation contains three members that are completely devoid of microfossils  
249 due to their highly oxidizing ~~deep-water~~ depositional environment (Herbosch *et al.* 2016).

250 However, the discovery of conodonts of the *P. proteus* Biozone at the boundary between the  
251 Meuville and Les Plattes members (Vanguetaine *et al.* 2004) allows ~~to attribute~~assignment of  
252 this level to the uppermost Tremadocian or the lower Floian. Thus, the Ottré Formation  
253 extends from the top of the Tremadocian to approximately the upper part of the Floian (Fig.  
254 3).

255 The Bihain Formation comprises two members. The upper Salmchâteau Member contains  
256 acritarchs of the informal Zone 9 (Vanguetaine 1986, 1992), which can be correlated with  
257 the *I. gibberulus* and *A. cucullus* graptolite biozones (Cooper and Molyneux 1990; Servais *et*  
258 *al.* 2017; Herbosch, 2021). The Bihain Formation extends approximately from the uppermost  
259 part of the Floian to the lowermost Darriwilian (Fig. 3). The sedimentary record of the SVI  
260 ends with the Salmchâteau ~~member~~Member of the Bihain Formation.

### 262 *The Rocroi Inlier*

263  
264 The Rocroi Inlier (RI) is a E-W, about 20 by 70 km elongated area, located at the SW edge of  
265 the Ardenne Allochthon (Fig. 2). It is ~~surrounded~~bounded in by angular unconformity ~~by~~  
266 with above the uppermost Silurian and/or Lower Devonian except in the SW, where it  
267 disappears under the Mesozoic and Cenozoic cover. The RI shows a continuous siliciclastic  
268 sedimentation from the lower Cambrian to the Middle Ordovician (Beugnies 1963). The  
269 Ordovician is represented by a single formation and crops out in a very small area on the  
270 French-Belgian border.

271 The sedimentary record begins with the Vieux-Moulin de Thilay Formation (Fig. 3), which  
272 was defined by Beugnies (1963) as the top of the Revin Group (Cambrian). It was  
273 lithostratigraphically redefined as belonging to the Ordovician by Geukens (1981), a  
274 hypothesis confirmed by Roche *et al.* (1986), who found acritarchs from the informal Zone 9

275 (Vanguetaine 1986, 1992). This zone was correlated with the *I. gibberulus* and *A. cucullus*  
276 graptolite biozones (see above) ~~dated~~ ranging from the uppermost Dapingian to the lowermost  
277 Darriwilian. Therefore, this formation ~~is to be situated in~~ has the same stratigraphic range as  
278 the Salmchâteau Member of the Bihain (SVI) and the Abbaye de Villers (BM) formations  
279 (Fig. 3).

280

### 281 *Givonne and Serpont inliers*

282

283 Ordovician strata are considered to be absent in these two less well known inliers (Beugnies  
284 1960; Geukens and Richter 1962; Herbosch 2021).

285

## 286 **The Ordovician of the Rhenish Massif and Rügen (by Lefebvre and** 287 **Servais)**

288

### 289 *General framework*

290

291 In western Germany, Ordovician strata are exposed in the eastern part of the Stavelot-Venn  
292 inlier (SVI; see above) and also in the two Ebbe and Solingen–Remscheid–Altena inliers, in  
293 the Rhenish Massif (Fig. 1; Servais *et al.* 2008, and references therein). Thick, tectonically  
294 complex Ordovician successions are also known from boreholes in Rügen (Fig. 1), in  
295 northeastern Germany, and Pomerania, in northwestern Poland (Trela, this volume). All these  
296 regions display relatively comparable successions of Ordovician rocks, yielding similar  
297 palynomorph assemblages and low-diversity deep shelf faunas dominated by graptolites and  
298 phyllocarids. Moreover, strong lithological, faunistic and geochemical similarities between  
299 Ordovician successions in Rügen, the Rhenish Massif, the SVI, the Brabant Massif, the



300 English Lake District and the Welsh Basin strongly support the view that all these areas were  
301 part of the same Avalonia microcontinent (Maletz and Servais 1993; Servais *et al.* 1998,  
302 2008; Pharaoh 1999; Samuelsson *et al.* 2002; Verniers *et al.* 2002). In the last thirty years,  
303 there has been significant progress in-on the bio- and chronostratigraphy was-achieved-inof  
304 the Ordovician of the Rhenish Massif and Rügen through a series of detailed studies based on  
305 acritarchs, chitinozoans and graptolites.

306

### 307 *Rhenish Massif*

308

309 In western Germany, the most extensive and fossiliferous succession of Ordovician  
310 sedimentary rocks is exposed in the Ebbe and Solingen–Remscheid–Altena inliers, both  
311 situated near Cologne (Sauerland), in the northeastern part of the Rhenish Massif  
312 (*Rheinisches Schiefergebirge*; Fig. 1). In this region, the about c. 800 m thick Herscheid Beds  
313 (*Herscheider Schichten*) consist of a monotonous, siliciclastic succession (mudstones and  
314 siltstones), classically subdivided into four units (Maletz 2000; Eiserhardt *et al.* 2001; Servais  
315 *et al.* 2008; Koch *et al.* 2014).

316 The stratigraphically lowermost unit (Plettenberg Bänderschiefer Formation) is represented  
317 by 65 m thick dark (grey-blue) compact mudstones with intercalated pyrite-rich siltstones.  
318 The base of this formation and putatively underlying strata are not exposed in the Sauerland.  
319 The Bänderschiefer Plettenberg Formation yielded a graptolite assemblage typical of the *H.*  
320 *lentus* Biozone, which corresponds to the lower part of the *D. artus* Biozone (Fig. 4; Maletz  
321 and Servais 1993; Maletz *in* Eiserhardt *et al.* 2001; Servais *et al.* 2008; Koch *et al.* 2014). The  
322 mid Darriwilian age of the Plettenberg Bänderschiefer Formation was also confirmed by  
323 chitinozoans (Samuelsson *et al.* 2002). Other faunal elements include foraminiferans,  
324 phyllocarids, trilobites, and trace fossils (Riegraf and Niemeyer 1996; Koch and Brauckmann

325 1998; Koch 1999; Eiserhardt *et al.* 2001). The age and faunal composition (graptolites,  
326 palynomorphs, trilobites) of the Plettenberger Bänderschifer are very similar to those of the  
327 Huy Formation in Belgium (Condroz inlier; CI, see above; Servais *et al.* 2008; Koch *et al.*  
328 2014). The trilobite assemblage indicates relatively deep, outer shelf to slope environmental  
329 conditions (Owens and Servais 1997; Servais *et al.* 2008).

330 The overlying Kiesbert Tonschiefer Formation (150–200 m thick) is composed of dark  
331 (grey-blue to black) mudstones with rare sandstone interbeds. This unit was ~~dated~~-correlated  
332 ~~to~~-with the middle Darriwilian, based on its graptolite content, which is typical of the *N.*  
333 *fasciculatus* Zone (upper part of the *D. artus* Biozone), and also on its chitinozoan assemblage  
334 (Fig. 4; Maletz and Servais 1993; Maletz *in* Eiserhardt *et al.* 2001; Samuelsson *et al.* 2002;  
335 Servais *et al.* 2008; Koch *et al.* 2014). The Kiesbert Tonschiefer Formation also yielded trace  
336 fossils (*Tomaculum problematicum*), brachiopods, conulariids, ostracods, phyllocarids, and  
337 trilobites (Beyer, 1941, 1943; Koch and Brauckmann 1998; Koch 1999; Schallreuter and  
338 Koch 1999; Servais *et al.* 2008).

339 The Rahlenberg Grauwackenschiefer Formation is a 300 m thick succession of thin bedded  
340 dark (blue to grey-black) silty mudstones with sandy layers that yielded siliceous concretions  
341 in its upper part. The fauna is ~~very~~-scarce and comprises conulariids, graptolites, phyllocarids,  
342 and trilobites (Eisenack 1939; Servais *et al.* 2008). Palynomorphs indicate an early Sandbian  
343 age (Fig. 4; Eisenack 1939; Samuelsson *et al.* 2002; Servais *et al.* 2008; Koch *et al.* 2014), in  
344 a good accordance with the presence of the graptolite *Pseudoclimacograptus* sp. (Maletz  
345 2000).

346 The Solingen Tonschiefer Formation is the stratigraphically highest Ordovician unit in the  
347 Sauerland area. It consists of about 200 m thick dark (grey to black) mudstones with silty to  
348 sandy levels, yielding very few fossil remains (graptolites, phyllocarids and, trilobites) and  
349 trace fossils (Servais *et al.* 2008; Koch *et al.* 2014). An early Katian age was proposed for this

350 formation, based on chitinozoans (Fig. 4; Samuelsson *et al.* 2002; Servais *et al.* 2008; Koch *et*  
351 *al.* 2014). The Solingen Tonschiefer Formation is overlain by late-upper Silurian (Pridoli)  
352 rocks of the Köbbinghauser Schichten Formation. This implies a stratigraphic gap extending  
353 from the middle Katian to most of the Silurian (Timm *et al.* 1981; Maletz and Servais 1993;  
354 Verniers *et al.* 2002).

355 In summary, the Ebbe and Solingen–Remscheid–Altena inliers expose a 800 m thick,  
356 discontinuous Ordovician succession comprising middle Darriwilian (Bänderschiefer  
357 Plettenberg and Kiesbert Tonschiefer formations), early-lower Sandbian (Rahlenberg  
358 Grauwackenschiefer Formation) and early-lower Katian (Solingen Tonschiefer Formation)  
359 rocks (Fig. 4). Palaeontological data suggest the persistence (at least from the mid Darriwilian  
360 to the early Katian) of low diversity assemblages associated to relatively deep (outer shelf,  
361 slope) palaeoenvironmental conditions (Owen and Servais 1997; Samuelsson *et al.* 2002;  
362 Verniers *et al.* 2002; Servais *et al.* 2008; Koch *et al.* 2014).

363

## 364 *Rügen*

365

366 In northeastern Germany, a particularly thick Ordovician sedimentary succession (over 3,000  
367 m) was described from numerous boreholes drilled in Rügen (Fig. 1) during the 1960s (Jaeger  
368 1967; Servais 1994; Maletz 1998, 2001; Beier *et al.* 2001; Verniers *et al.* 2002; Servais *et al.*  
369 2008). This mainly pelitic and clastic succession is synsedimentary deformed, locally  
370 overturned (Franke and Illers 1994; Beier and Katzung 1999; Samuelsson *et al.* 2000),  
371 suggesting unstable environmental conditions (slumps) along an active continental margin  
372 and/or tectonic stacking (Servais and Katzung 1993; Giese *et al.* 1994; McCann 1998). With  
373 the exception of graptolites and palynomorphs, only trace fossils (e.g. *Nereites*) and  
374 phyllocarid crustaceans were reported from the boreholes (Zagora 1997; Maletz 2001). A

375 subdivision of the Ordovician succession into three tectonically limited lithostratigraphic units  
376 was established, based on the thickest and most complete drilled successions (e.g. the Rügen  
377 5 borehole) (Giese *et al.* 1994; Beier *et al.* 2001; Servais *et al.* 2008).

378 The stratigraphically lowermost unit, the Varnkevitze Sandstein Formation, was only  
379 observed in the deepest and most complete borehole (Rügen 5). It consists of about 400 m of  
380 fine sandstones and intercalated shales (Giese *et al.* 1994; Beier *et al.* 2001). About 60 m  
381 above the base of this formation, a 200 m thick interval of black shales yielded both acritarchs  
382 typical of the *messaoudensis-trifidum* assemblage and a *Lagenochitina destombesi*-dominated  
383 chitinozoan association, both supporting a late Tremadocian age (Fig. 4; Servais and Katzung  
384 1993; Servais and Molyneux 1997; Samuelsson 1999; Beier *et al.* 2001; Samuelsson *et al.*  
385 2000; Servais *et al.* 2008).

386 The overlying Arkona Schwarzschiefer Formation is a very thick succession of black  
387 shales (over 1,000 m) observed in most boreholes (Samuelsson *et al.* 2000; Maletz 2001).  
388 Both its lower and upper parts are tectonically disturbed, and no borehole yielded a complete,  
389 continuous succession (many intervals are missing, probably due to tectonics). These levels  
390 have yielded graptolite assemblages characteristic of the middle and ~~late-upper~~ Darriwilian  
391 (equivalents of the *H. lentus*, *P. elegans*, *P. distichus*, and ?*H. teretiusculus* biozones) and the  
392 ~~early-lower~~ Sandbian (*N. gracilis* Biozone) (Fig. 4; Jaeger 1967; Maletz 1998, 2001).  
393 Palynomorphs provided a comparable range of ages for the Arkona Schwarzschiefer  
394 Formation (Burmam 1968, 1970; Samuelsson *et al.* 2000; Samuelsson and Servais 2001;  
395 Vecoli and Samuelsson 2001; Servais *et al.* 2008).

396 The Nobbin Grauwacken Formation is a very thick (~1,800 m) succession dominated by  
397 greywackes and shales, which was only observed in some boreholes. Graptolite and  
398 palynomorph assemblages yielded congruent ages (middle Darriwilian–early Sandbian) for  
399 this unit, similar to those obtained for the Arkona Schwarzschiefer Formation (Fig. 4; Maletz

400 1998, 2001; Samuelsson *et al.* 2000; Servais *et al.* 2001, 2008). This implies that both  
401 formations were probably concurrently deposited in distinct environmental conditions; the  
402 black shales of the Arkona Schwarzschiefer Formation were probably related to more distal  
403 (bathyal?) settings than the greywackes of the Nobbin Grauwacken Formation (Zagora 1997;  
404 Samuelsson *et al.* 2000).

405 In summary, the subsurface of the island of Rügen yielded a very thick, though tectonically  
406 complex, Ordovician succession comprising Tremadocian (Varnkevitze Sandstein Formation)  
407 and middle Darriwilian to lower Sandbian (Arkona Schwarzschiefer and Nobbin Grauwacken  
408 formations) sedimentary rocks (Fig. 4). The Ordovician successions of Rügen, the Rhenish  
409 Massif (see above) and Pomerania (NW Poland; see Trela, this volume) share many  
410 similarities in terms of lithologies, faunas, stratigraphy (e.g. latest Tremadocian–early  
411 Darriwilian gap), and associated environmental conditions (both were originally deposited in  
412 deep-water settings), thus supporting the view that they were parts of a same  
413 palaeogeographic area (Jaeger 1967), in eastern Avalonia (Maletz and Servais 1993; Servais  
414 *et al.* 1998; Pharaoh 1999; Maletz 2001; Samuelsson *et al.* 2002; Verniers *et al.* 2002).

415

## 416 **The Ordovician of the Armorican Massif (by Vidal, Loi, Lefebvre and** 417 **Ghienne)**

418

### 419 *General framework*

420

421 The Armorican Massif is usually subdivided into four tectonic units (Fig. 5): the Leon  
422 Domain (LD), the North Armorican Domain (NAD), the Median Armorican Domain (MAD,  
423 and the South Armorican Domain (SAD) (Matte 2001; Ballèvre *et al.* 2009). As the LD did  
424 not yield-contains any Palaeozoic sedimentary successions, only the three other domains will

425 be described below. A subdivision into similar tectonic units, with comparable stratigraphic  
426 and faunal features, occurs also in the Iberian Peninsula, thus strongly supporting the view  
427 that these two areas, united by a similar evolution during the Variscan orogeny, originally  
428 belonged to a same Ibero-Armorican palaeogeographic Domain (e.g. Paris and Robardet  
429 1977, 1994; Robardet and Gutiérrez-Marco 1990; Robardet *et al.* 1994b; Gutiérrez-Marco *et*  
430 *al.* this volume).

431 The NAD (i.e. Cotentin, zone bocaine and Mortain-Domfront area) and the MAD (i.e.  
432 Châteaulin, Menez-Belair and Laval synclines) are separated by the North Armorican Shear  
433 Zone (NASZ). Their Palaeozoic successions are quite similar from the Middle Ordovician  
434 onwards (Fig. 6). Based on their tectonic history and faunistic affinities, the NAD and MAD  
435 are often ~~considered as forming united together within~~ a single Median-North Armorican  
436 Domain (MNAD; Robardet *et al.* 1994a, b; Paris 2016). However, recent U-Pb analyses on  
437 detrital zircons coupled with Sm-Nd and Lu-Hf isotope analyses identified different zircon  
438 populations in the Lower Palaeozoic successions of these two areas (Dabard *et al.* 2021). In  
439 the Ordovician sediments of the MAD, the occurrence of Stenian detrital zircons suggests  
440 affinities with the eastern African "Sahara metacraton" and the Arabian Nubian Shield. In  
441 contrast, the absence of Stenian and late Tonian ages in zircon populations ~~of in~~ Ordovician  
442 sediments from the NAD implies distinct source areas, probably located more westward.  
443 These features suggest that the NAD and the MAD were geographically separated during the  
444 Early Ordovician, and getting-moving closer only during the Variscan orogeny (Fig. 1).

445 The MAD is separated from the SAD (i.e. south of Saint-Julien-de-Vouvantes unit, Saint-  
446 Georges-sur-Loire and Ancenis synclines in southern Brittany, and Chantonay syncline in  
447 Vendée) by the northern branch of the South Armorican Shear Zone (SASZ, Fig. 5). The  
448 Palaeozoic succession of the SAD is markedly distinct from those of the NAD and MAD. The  
449 three domains differ by-in the lower part of the Ordovician succession, which rests

450 conformably on the lower Cambrian in Normandy-NAD (Séries de Carteret; Doré 1969;  
451 Pillola 1993), on the Miaolingian in the SAD (Schistes à *Paradoxides* of Cléré-sur-Layon,  
452 South of Angers; Cavet *et al.* 1966) and unconformably on the Brioverian basement (see  
453 below) in central Brittany (MAD; Robardet *et al.* 1994a, b).

454 The main features of the Ordovician succession in the MAD, which is the most complete  
455 and best exposed within the Armorican Massif, will be presented in detail below, with brief  
456 references about differences with the NAD and the SAD.

457

### 458 *Lower Ordovician*

459

460 In central Brittany, the Ordovician succession (Fig. 6) begins with basal red-bed sequences  
461 (*Séries Rouges Initiales* or Cap de la Chèvre Formation), followed by the geographically  
462 widespread Grès Armoricain Formation. These ~~earliest~~ lowest Ordovician deposits rest  
463 unconformably on the Brioverian succession (Gougeon *et al.* 2022 and references therein).

464 Probably diachronous within the Armorican Massif, the Brioverian ~~series~~ succession are is  
465 dated as ~~Precambrian~~ late Neoproterozoic in both the NAD and the SAD, based on their  
466 stratigraphic situation below the Cambrian. In the MNAD, recent investigations on historical  
467 and new outcrops have yielded low diversity trace-fossils with not penetrative but simple  
468 horizontal patterns, microbially induced sedimentary structures and very rare soft-body fossils  
469 (*Nimbia* sp.), coupled with U-Pb detrital zircon datings, all suggesting an Ediacaran–  
470 Fortunian age for at least the upper part of the Brioverian succession (Gougeon *et al.* 2018,  
471 2022; Néraudeau *et al.* 2018).

472 Within the Armorican Massif, red-bed sequences are discontinuous (named Cap de la  
473 Chèvre Formation in the southern part of the Crozon Peninsula, Pont-Réan Formation in the  
474 synclines south of Rennes, and Margat Formation close to Angers in the SAD) and start with

475 a ~~polygenie~~polymict conglomerate overlain by sandstones interbedded with silty-clayey  
476 sediments (Bonjour 1988; Bonjour and Chauvel 1988; Suire *et al.* 1991; Egal *et al.* 1996).  
477 The source area of sediments is the Brioverian basement (Dabard *et al.* 2021). The main  
478 sedimentary structures are either plane- or cross-bedding laminations in the sandstones and  
479 desiccation cracks in some siltstones or argillites levels. The geometry of sedimentary bodies  
480 and the associated structures suggest depositional conditions ranging from alluvial fans and  
481 braided river systems to alluvial plains (Bonjour 1988; Durand 1989; Suire *et al.* 1991; Egal  
482 *et al.* 1996). Then, the environment progressively opened up to the marine domain, probably  
483 as a storm-influenced delta. Despite the lack of body fossils, the occurrence of living  
484 organisms is evidenced in the upper part of these red-bed sequences by the record of  
485 bioturbation traces such as *Cruziana*-type bilobed tracks and *Skolithos*-type vertical burrows.

486 This lowermost part of the Ordovician succession was deposited on a rather unstable  
487 continental basement, as evidenced by the contemporary effusive volcanism and variations in  
488 lateral thickness, which are indicative of a significant tectonically induced subsidence (Noblet  
489 1983; Bonjour 1988). This instability was controlled by extensional tectonics (Ballard *et al.*  
490 1986; Dauteuil *et al.* 1987; Brun *et al.* 1991) expressed ~~from~~in the Cambrian and likely  
491 associated with the opening of the Rheic Ocean between Avalonia and Gondwana (Paris and  
492 Robardet 1990; Prigmore *et al.* 1997; Linnemann *et al.* 2008). The dating applied to  
493 interbedded volcanoclastic levels have assigned numeric ages ranging from  $472 \pm 5$  Ma (Rb-  
494 Sr, red-bed sequences of northern Brittany; Auvray *et al.* 1980) to  $465 \pm 1$  Ma (U-Pb, Cap de  
495 la Chèvre Formation; Bonjour *et al.* 1988). The latter age is, however, inconsistent with the  
496 biostratigraphic ~~allocations~~correlations of overlying formations (e.g. the chitinozoan *E.*  
497 *brevis* Zone, indicative of the upper Floian; see Paris 1990, 2016).

498 The thickness of the Grès Armorica Formation varies from tens to several hundreds of  
499 meters. Depending on the regions, it consists of either an undivided mass of sandstones with



500 silty-clayey intercalations (e.g. Domfront in Normandy, northern flank of Menez-Bélaïr,  
501 Vendée), or is heterogeneous and subdivided into three members: (i) a lower member  
502 including thick sandstone beds only separated by few silty to clayey intercalations. This  
503 member can comprise iron layers, which used to be exploited in the synclines south of Rennes  
504 (Chauvel 1968); (ii) a middle member (variously named "Congrier Member", "Gador  
505 Member" or *Schistes intermédiaires* depending on the region), consisting of siltstones,  
506 argillites with micaceous sandstone intercalations; and (iii) an upper member, which is mainly  
507 composed of sandstones with silty-clayey intercalations (e.g. in the Châteaulin synclinorium,  
508 synclines south of Rennes), and containing sometimes highly radioactive placers (rutile and  
509 zircon grains; Faure 1978; Pistis *et al.* 2008, 2016).

510 One of the main features of the Grès Armoricaïn Formation is the occurrence of numerous  
511 sedimentary structures (e.g. hummocky cross stratification, planar laminations, erosive basal  
512 surfaces, wave and current ripples and clay drapes). They reflect shallow environments  
513 subject to tidal influence and to fair-weather and storm waves (Durand 1985; Guillocheau *et*  
514 *al.* 2009). Moreover, this formation has yielded highly diverse trace fossils, which are either  
515 vertical or horizontal (Durand 1985). Surprisingly, the fauna is sparse with few *Lingula*-rich  
516 levels and, in some areas, brachiopods such as "*Dinobolus*". In the uppermost part of the Grès  
517 Armoricaïn Formation, rare bivalves (Babin 1966) and trilobites (Henry 1980) ~~can be~~  
518 ~~found~~ occur.

519 Sedimentological and palaeobiological data suggest that the Grès Armoricaïn Formation  
520 was deposited in environments ranging from a protected marine domain (bay, lagoon) to the  
521 top of the continental shelf (shoreface to upper offshore; Durand 1985; Dabard *et al.* 2007;  
522 Guillocheau *et al.* 2009; Pistis *et al.* 2016). Some clayey levels of the middle and upper  
523 members yielded chitinozoans typical of the *E. brevis* Biozone (Paris 1990), thus supporting a  
524 Floian age for these levels. The huge variations in thickness observed for the Grès Armoricaïn

525 Formation (i.e. from a few meters up to 600 meters) suggest very high local rates of  
526 sedimentation (Noblet 1983; Durand 1985; Dabard *et al.* 2009), probably driven by an active  
527 tectonic influence.

528

### 529 *Middle Ordovician – Sandbian*

530

531 During the Dapingian–Sandbian ~~time~~ interval, the sedimentation was dominated by the  
532 deposition of dark, fine-grained shales, previously called "Schistes à *Calymene*", and  
533 currently named the Postolonnec, Andouillé, Le Pissot and Traveusot formations in the  
534 MNAD, Schistes d'Angers and Fresne formations in the SAD (Fig. 6). These 150 to 400 m  
535 thick silty-clayey units bear levels with siliceous, phosphatic and/or calcareous concretions,  
536 shell-beds, and sandstone intercalations that can be several dozen of meters thick (e.g.  
537 Kerarvail Member, about 30 meters thick in the Crozon Peninsula). The main sedimentary  
538 structures are planar bedding, hummocky cross stratifications and numerous A- or B-type  
539 shell beds (type A: thin, lenticular single coquinas; type B: thick, continuous and composite  
540 shell concentrations, *sensu* Botquelen *et al.* 2004). All these structures are indicative of a  
541 storm wave-influenced sedimentation. These deposits are related to various environments  
542 ranging from beach (shoreface) to lower offshore (Guillocheau 1983; Loi *et al.* 1999; Loi and  
543 Dabard 2002; Dabard *et al.* 2007; Guillocheau *et al.* 2009). Phosphatic and siliceous  
544 concretions, and crusts are common in Middle Ordovician sedimentary rocks. They are of ~~an~~  
545 early diagenetic origin and are promoted by the stability of the water/sediment interface.  
546 These concretions, which are indicative of sedimentary condensation, were mainly produced  
547 during sea-level rises (Loi *et al.* 1999; Loi and Dabard 2002; Dabard *et al.* 2007; Dabard and  
548 Loi 2012). Depending on the depositional environment, they can be linked to composite shell  
549 beds (B-type shell beds *sensu* Botquelen *et al.* 2004).

550 Bioturbation is less diverse than in the underlying Grès Armorica Formation, with  
551 *Planolites* and *Teichichnus* as the most common trace fossils (Guillocheau 1983; Guillocheau  
552 *et al.* 2009). In contrast, invertebrate faunas are particularly diverse and consist mainly of  
553 benthic taxa (Vidal *et al.* 2011a): bivalves (Babin 1966), brachiopods (Mélou 1973, 1975,  
554 1976; Botquelen and Mélou 2007), echinoderms with aristocystitid diploporites, crinoids,  
555 ophiuroids and stylophorans (Chauvel 1941, 1980, 1981; Chauvel and Nion 1969, 1977;  
556 Lefebvre and Vizcaïno 1999; Lefebvre 2000, 2007; Hunter *et al.* 2007; Lefebvre *et al.* 2015,  
557 2022; Blake *et al.* 2016), ostracods (Vannier 1986a, b), and trilobites (Henry 1980; Vidal *et*  
558 *al.* 2011b; Gendry *et al.* 2013; Courville and Gendry 2016). Additional faunal elements  
559 include conodonts (Lindström *et al.* 1974), graptolites (Philippot 1950; Henry *et al.* 1976;  
560 Paris and Skevington 1979) and organic-walled microfossils, such as acritarchs (Deunff 1951,  
561 1954, 1958; Henry 1969; Paris and Le Hérisse 1992) and chitinozoans (Paris 1981, 2016).  
562 The ages inferred from graptolites and chitinozoans range from the Dapingian (*D. ornensis*  
563 Biozone) to the Sandbian (*L. dalbyensis* Biozone; Paris 1981, 1990, 2016; Dabard *et al.*  
564 2015). The benthic faunas display clear Gondwanan affinities, particularly with the  
565 Mediterranean Province and are particularly similar to those described in the Iberian  
566 Peninsula, highlighting the long-established similarities between the faunas of these two  
567 regions (e.g. Henry *et al.* 1976; Paris and Robardet 1977, 1994). Some levels record the  
568 immigration of some taxa following transgressive events, as for example the *Marrolithus*  
569 *bureaui* level (Henry *et al.* 1993), which coincides with the appearance of Balto-Scandinavian  
570 chitinozoans in the lower Sandbian (e.g. *L. stentor*; Paris 2016).

571 Differences in the composition of Darriwilian–Sandbian benthic assemblages in the North-  
572 Median- and South Armorican Domains mainly result from distinct environmental settings,  
573 with a southward (in present-day geography) deepening trend (Henry 1989; Henry *et al.* 1997;  
574 Lefebvre 2007; Courville and Gendry 2016). In the NAD, benthic diversity is low and

575 trilobite taxa (Homalonotinae) are typically related to shallow-[water](#) environments (Henry  
576 1989 and references therein). In the MAD, the faunal diversity is significantly higher and  
577 characteristic of median shelf settings with both typical endobenthic and epibenthic trilobite  
578 taxa (Henry 1989; Courville and Gendry 2016). In the SAD, fossils are rare and outcrops  
579 discontinuous. However, assemblages are typical of distal shelf environments, with strong  
580 affinities with coeval Bohemian faunas from the Prague Basin (Henry *et al.* 1997; Lefebvre *et*  
581 *al.* 2010). Assemblages correspond to the "atheloptic biofacies" with benthic trilobites either  
582 blind or with atrophied eyes, associated ~~to~~-[with](#) pelagic forms with hypertrophied eyes  
583 (Cyclopygidae), along with lagynocystid mitrate stylophorans, typical of deep-[water](#)  
584 environments of the Prague Basin (Henry *et al.* 1997; Lefebvre 2007; Lardeux *et al.* 2008;  
585 Jouhier and Gendry 2017).

586 Several transgressive and regressive cycles were identified in the Middle Ordovician–  
587 Sandbian succession of the MAD, based on the detailed analysis of sedimentary facies  
588 (Dabard *et al.* 2007, 2009; Paris *et al.* 2007). Transgressive episodes identified within the *D.*  
589 *bullata*, *S. formosa* and *L. pissotensis* chitinozoan biozones were recognized in other peri-  
590 Gondwanan areas (Dabard *et al.* 2007; Paris *et al.* 2007; Videt *et al.* 2010). In the Crozon  
591 Peninsula, two key sections within the Postolonnec Formation provided a continuous, well-  
592 exposed framework for the detailed analysis of sedimentary facies and gamma-ray record  
593 (Dabard *et al.* 2015). There, a back stripping procedure was applied and calibrated with  
594 graptolite biozones to convert the observed variations in depositional environments into  
595 apparent sea-level changes, and ultimately into eustatic variations. The resulting pattern (Fig.  
596 7) supports the occurrence of short-term and high amplitude cyclic sea-level fluctuations,  
597 suggesting glacio-eustasy. This pattern suggests the occurrence of ice-house conditions and  
598 oscillations of an ice-sheet on Gondwana at least at the lower limit of the *L. pissotensis*  
599 Biozone (at the transition between the Kerarvail and Morgat members), but probably as early

600 as the *S. formosa* chitinozoan Biozone during the Darriwilian (i.e. at the transition between  
601 the Kerloc'h and Corréjou members, Fig. 6; Dabard *et al.* 2015). The expression of these ice  
602 sheet-induced oscillations differs in other parts of the Armorican Massif, depending on their  
603 initial location along a proximo-distal transect and/or erosional surfaces. For example, no  
604 sandy member (equivalent to Kerarvail in Crozon) is observed in the central and eastern parts  
605 of the MAD, characterized by more distal palaeoenvironmental conditions. In Normandy, the  
606 coarser, shallow-water facies of the Middle Ordovician successions and the quality of the  
607 outcrops do not make it possible to achieve any ~~continuous~~coherent, detailed analysis.  
608 Nevertheless, all these different sections ~~could~~may be correlated based on their chitinozoan  
609 assemblages (Fig. 6; Paris 2016).

610

### 611 *Katian*

612

613 In many regions of the Mediterranean Province, the Sandbian–Katian transition coincides  
614 with a major regressive episode (Robardet 1981; Villas 1992; Bourahrouh *et al.* 2004). In  
615 both Brittany and Portugal, the recovery of the sedimentation process is expressed by a  
616 conglomeratic layer bearing phosphatic pebbles, ferruginous and phosphatic oolites (Paris  
617 1979, 1981; Young 1988). Palynological data suggest a stratigraphic gap, at least locally:  
618 chitinozoans extracted from the conglomerate matrix belong to the *E. tanvillensis* Biozone  
619 (Katian), whilst those from the pebbles belong to the *L. deunffi* or *L. dalbyensis* biozones  
620 (Sandbian; Paris 1981; Bourahrouh 2002; Vidal *et al.* 2011b; Dabard *et al.* 2015). This level  
621 represents the base of a new transgressive sequence corresponding to the Kermeur and Saint-  
622 Germain-sur-Ille formations in the median synclinorium, and to the Grès du Châtellier  
623 followed by the Schistes de Riadan south of Rennes (Dabard *et al.* 2009; Paris 2016). These  
624 sandstone-dominated deposits contain silty and clayey intercalations, as well as phosphatic,

625 siliceous and/or calcareous concretions. The sedimentary structures (hummocky cross-  
626 stratifications, symmetric ripples [=oscillation ripples], clay drapes, inversion in palaeocurrent  
627 directions, evidence of emersion) reflect shallow environments subjected to tidal influence  
628 and to fair-weather and storm waves. The depositional environments inferred from the  
629 sedimentary facies range from barrier and back-barrier nearshore settings (e.g. base of the  
630 Kermeur Formation; Vidal *et al.* 2011b) to the upper offshore (Gorini *et al.* 2008) throughout  
631 several transgression-regression cycles (Gorini *et al.* 2008; Dabard *et al.* 2009). Some levels  
632 are extensively bioturbated by well-diversified trace fossils (Mélou and Plusquellec 1975),  
633 including *Equilibrichnia* (Vidal *et al.* 2011b) and spiral-shaped traces. Benthic assemblages  
634 are scattered in the arenaceous base of the Kermeur Formation, and in the siltstones of the  
635 upper part of this formation, equivalent to the Châtellier and Riadan formations. The main  
636 taxa are trilobites (see Henry 1980; Vidal *et al.* 2011b), diploporite echinoderms, brachiopods,  
637 and rare bivalves (Babin and Mélou 1972; Mélou 1985; Botquelen and Mélou 2007) and  
638 gastropods. These faunas are less diversified than those occurring in underlying formations.  
639 Accurate stratigraphic assignments of the Kermeur, Saint-Germain-sur-Ille, Châtellier and  
640 Riadan formations are provided by chitinozoans, indicative of the Katian (from the *E.*  
641 *tanvillensis* to the *A. barbata* biozones; Paris 1981, 2016; Bourahrouh 2002; Gorini *et al.*  
642 2008). No coeval section is exposed in the SAD, but concretions yielded a typical Katian  
643 trilobite fauna with *Deanaspis*, *Dreyfussina*, *Eudolatites* and *Prionocheilus* (Beaulieu *et al.*  
644 2014).

645 The widespread upper Katian calcareous facies occurring in most regions of the  
646 Mediterranean Province have been interpreted as reflecting a period of global warming, the  
647 so-called "Boda Event" (Fortey and Cocks 2005). In the Armorican Massif, they correspond  
648 to spatially restricted units, such as the Vaux Limestone in Normandy (Sées syncline) and the  
649 Rosan Formation in the Crozon Peninsula (Fig. 6). In the latter region, calcareous facies are

650 associated with an anorogenic volcanic complex suggestive of shallow submarine to partially  
651 aerial conditions (Juteau *et al.* 2007; Caroff *et al.* 2009). The limestones yielded brachiopods  
652 (Mélou 1990), bryozoans, conodonts (Lindström and Pelhate 1971; Paris *et al.* 1981), and  
653 locally also echinoderms (crinoids and cystoids; Chauvel and Le Menn 1972) and trilobites  
654 together with the oldest occurrence of rugose corals in the Armorican Massif. Brachiopods  
655 and conodonts correspond to a mid-late Katian age (*A. ordovicicus* Zone; Paris *et al.* 1981;  
656 Mélou 1990; Ferretti *et al.* 2014).

657

### 658 *Hirnantian*

659

660 In the Mediterranean Province, the Hirnantian stage is characterized by a well-recorded  
661 glacial episode in most regions (e.g. Algeria, Morocco, Portugal, Spain; Ghienne *et al.* 2007;  
662 Loi *et al.* 2010; Videt *et al.* 2010). In the Armorican Massif, this interval corresponds to silty-  
663 clayey deposits containing scattered clasts of various sizes and lithologies, which are  
664 interpreted as dropstones (Dangeard and Doré 1971; Robardet and Doré 1988). Hirnantian  
665 deposits constitute the *Pélites à fragments* (diamictites) in Normandy, the uppermost part of  
666 the Saint-Germain-sur-Ille Formation in Menez-Bélaire and Laval synclines, the Chesnaie  
667 Formation in the South of Rennes or the Cosquer Formation in western Brittany. These facies  
668 are interpreted as glaciomarine deposits composed of fragments originating from melting ice  
669 (Hamoumi 1981; Hamoumi *et al.* 1981; Robardet and Doré 1988).

670 In these levels, the macrofauna is scarce. In the Crozon Peninsula, a single locality  
671 (Trégarvan section) yielded a fossiliferous sandstone level with *Hirnantia sagittifera* (Mélou  
672 1990), elsewhere named “Grès de Roudou Hir” or “Grès de Lamm Saoz” at the top of the  
673 Cosquer Formation (Plusquellec *et al.* 1999). This ubiquitous brachiopod is emblematic of the

674 Hirnantian ~~stage~~Stage. This age is also supported by palynological data (chitinozoans), which  
675 are typical of the *T. elongata* Biozone (Bourahrouh 2002; Paris 2016).

676

## 677 **The Ordovician of the Montagne Noire, Mouthoumet and eastern Pyrenees** 678 **massifs (by Álvaro, Casas, Lefebvre, Monceret and Vizcaïno)**

679

### 680 *General framework*

681

682 After the late Cambrian rifting along the so-called Atlas – Ossa-Morena – North Armorican  
683 rift (western Gondwana; Álvaro *et al.* 2021), drifting conditions prevailed during the  
684 Ordovician~~times~~. However, the evolution of the high-latitude western Gondwanan margin  
685 was not a monotonous journey, but an eventful drift episodically punctuated by major  
686 geodynamics events. One of them was initially identified as a Middle Ordovician gap in  
687 southwestern Sardinia: the Sardic Phase (see Loi *et al.* this volume) has ~~represented~~during  
688 the last few decades become a standard for correlation in western Europe, in spite of  
689 sometimes different conceptions ~~among areas and scholars~~both regionally and academically.  
690 Similar gaps involving the absence of Middle Ordovician strata have been invoked to  
691 recognize Sardic-style geodynamic events in the eastern Pyrenees, southern France and the  
692 Alps (Stampfli *et al.* 2002). There, the Sardic phase is characterized by generalized cortical  
693 uplift, denudation of exposed uplifted areas under subaerial conditions, Middle Ordovician  
694 stratigraphic gaps of about 25–30 million years, broad felsic plutonism and volcanism (with  
695 felsic rocks now orthogneisses after Variscan deformation and metamorphism) with calc-  
696 alkaline affinity, and the record of alluvial to fluvial deposits onlapping ~~on~~an inherited  
697 palaeorelief (Álvaro *et al.* 2018; Casas and Álvaro 2019). However, the record of Ordovician



698 volcanism associated with local listric faults in neighbouring areas cannot be used as a  
699 supporting argument for contemporaneous Sardinian events ~~in those areas~~ (Álvaro *et al.* 2020).

700

### 701 *Tectonostratigraphic framework*

702

703 The southern Massif Central or Occitan Domain (Poulet *et al.* 2017) comprises the Variscan  
704 tectonostratigraphic units of the Thiviers-Payzac and Génis, Rouergue-Albigeois, Montagne  
705 Noire, southern Cévennes and Mouthoumet massifs (Álvaro *et al.* 2021) (Fig. 8). Two massifs  
706 are selected for comparison here: the Montagne Noire and the Mouthoumet massifs. The  
707 former has been traditionally subdivided into three tectonostratigraphic units, which are: (i)  
708 the northern Montagne Noire with a southward tectonic vergence, comprising Lower  
709 Palaeozoic low-grade metasedimentary units; (ii) the Axial Zone, an elongated dome of  
710 migmatized orthogneisses displaying an Ordovician age for their protoliths (460–450 Ma;  
711 Roger *et al.* 2004), which also includes migmatites and micaschists derived from  
712 Neoproterozoic–Ordovician rocks; and (iii) the southern Montagne Noire made up of south-  
713 facing nappes including a complete and fossiliferous Cambrian–Ordovician succession  
714 (Vizcaino *et al.* 2001) (Fig. 8). The emplacement of recumbent folds during the Viséan–early  
715 Namurian altered the pre–Variscan (palaeogeographic) proximal-to-distal trend, placing the  
716 (proximal) Axial Zone in an intermediate position between the southern and (distal) northern  
717 Montagne Noire flanks. The Mouthoumet Massif or inlier, located between the Montagne  
718 Noire and the North Pyrenean frontal thrust, represents the southernmost prolongation of the  
719 southern Massif Central (Berger *et al.* 1997).

720 The pre-Variscan rocks exposed in the eastern Pyrenees form an elongated strip that crops  
721 out along the backbone of the Alpine cordillera. The main tectonostratigraphic units  
722 recognized in the eastern Pyrenees are the Canigó/Canigou unit, comprising the Canigó and

723 the Roc de Frausa/Roc de France massifs, the Albera/Albères-Cap de Creus unit and the  
724 Aspres-Conflent unit (Fig. 8). The Canigó unit exhibits the most complete pre-Variscan  
725 sequence and separates palaeogeographically a proximal sector to the South (Puigmal sector)  
726 from a distal one to the North (Conflent unit), which mimics the same bathymetric trend  
727 displayed ~~from~~-across the southern to northern sides of the Montagne Noire (Álvaro *et al.*  
728 2021).

729

### 730 *Lower Ordovician (pre-Sardic) stratigraphic and volcano-sedimentary features*

731

732 A comprehensive summary of the Lower Ordovician stratigraphic nomenclatures from the  
733 southern Montagne Noire was ~~synthetized~~-presented by Vizcaïno *et al.* (2001), who  
734 subdivided the succession into several formations, alternating between shale-dominant (La  
735 Gardie, Saint-Chinian, Setso and Landeyran units) and sandstone-dominant (La Dentelle, La  
736 Maurerie, Cluse de l'Orb and Foulon) intervals, locally punctuated by carbonate interbeds  
737 (Val d'Homs and Mounio formations). A regional biozonation (Fig. 9) of the Tremadocian-  
738 Floian interval was proposed by Vizcaïno and Álvaro (2003), Tortello *et al.* (2006), and  
739 Álvaro *et al.* (2007, 2008). This biostratigraphic scheme relies on correlations with Avalonia  
740 (England), Baltica (Scandinavia) and high-latitude (peri-)Gondwana areas (e.g. Bohemian  
741 Massif- and Sardinia). It comprises, from bottom to top the *P. geinitzi* (mid Tremadocian), *S.*  
742 (*C.*) *pusilla* (mid-late Tremadocian), *E. filacovi* (late Tremadocian), *T. miqueli* (latest  
743 Tremadocian-early Floian), *T. shui landayranensis* (mid Floian), *C. maynardensis* (mid  
744 Floian), *N. (N.) arenosus* (mid Floian), *A. incisus* (late Floian) and *H. primitivus* biozones  
745 (late Floian) (Fig. 9). The base of each biozone corresponds to the FAD of its index taxon  
746 (Vizcaïno and Álvaro 2003). This biozonation was completed by the presence of conodonts in

747 the limestone interbeds belonging to the Tremadocian *P. deltifer* Biozone (Álvaro *et al.* 2007;  
748 Serpagli *et al.* 2007).

749 In the southern Montagne Noire, the Lower Ordovician succession records a shallowing-  
750 upward trend from the moderately deep shelf settings, below storm wave base, of the Saint-  
751 Chinian Formation (upper Tremadocian) to the more proximal, storm-dominated deposits of  
752 the Cluse de l'Orb and Foulon formations (mid Floian) (Courtessole *et al.* 1985; Dabard and  
753 Chauvel 1991; Vidal 1996; Vizcaïno and Lefebvre 1999; Vizcaïno *et al.* 2001). In the upper  
754 part of the Lower Ordovician succession, the black shales of the Landeyran Formation  
755 correspond to the onset of more distal, deeper offshore settings (Vizcaïno and Lefebvre 1999;  
756 Vizcaïno *et al.* 2001; Van Iten and Lefebvre 2020).

757 These changes in environmental conditions are also reflected by turnovers in the  
758 composition and diversity of faunal assemblages. The Saint-Chinian and La Maurerie  
759 formations yielded very diverse and abundant faunas comprising brachiopods, conulariids,  
760 echinoderms, graptolites, hyolithids, machaeridians, molluscs, and trilobites (e.g. Thoral  
761 1935; Courtessole *et al.* 1983; Vizcaïno *et al.* 2001). In marked contrast, the overlying Cluse  
762 de l'Orb and Foulon formations are characterized by low-diversity assemblages dominated by  
763 ~~inarticulate~~ non articulate brachiopods, eocrinoids and trilobites (e.g. Thoral 1935;  
764 Courtessole *et al.* 1985; Dabard and Chauvel 1991; Vidal 1996; Vizcaïno and Lefebvre 1999;  
765 Vizcaïno *et al.* 2001). The Landeyran Formation documents a major faunal transition, with  
766 the re-occurrence of several groups absent from the underlying Cluse de l'Orb and Foulon  
767 formations, but present in the Saint-Chinian and La Maurerie formations (e.g. agnostids,  
768 raphiophorid trilobites, stylophoran echinoderms). The Landeyran Formation yielded very  
769 diverse faunal assemblages including brachiopods, conulariids, echinoderms, graptolites,  
770 hyolithids, machaeridians, molluscs, ostracods, and trilobites (e.g. Dean 1966; Capéra *et al.*

771 1978; Courtessole *et al.* 1983; Vizcaïno *et al.* 2001; Tortello *et al.* 2006; Van Iten and  
772 Lefebvre 2020).

773 Two Early Ordovician magmatic events took place in the Occitan Domain that are absent  
774 in the Pyrenees. First, a Tremadocian felsic magmatism is recorded in the parautochthonous  
775 and allochthonous units of the Albigeois and northern Montagne Noire (Larroque Volcanic  
776 Formation), and the Mouthoumet Massif (Davejean acidic volcanics). This Tremadocian  
777 felsic magmatism was interpreted as the result of an episode of massive crustal melting  
778 related to asthenospheric upwelling leading to lithospheric doming and continental extension  
779 and decompressionally driven mantle melting, which triggered abundant subaerial explosive  
780 and effusive rhyolitic eruptions similar in facies and composition to the “Ollo de Sapo” event  
781 reported in the Iberian Massif. The second Early Ordovician magmatic event is illustrated by  
782 the Peyrebrune and Davejean Volcanic Complex basalts, which display a typical initial rift  
783 tholeiite (IRT) magmatic signature indicating contributions of both asthenospheric and  
784 lithospheric mantle sources (Pouclet *et al.* 2017).

785 In contrast, Furongian–Lower Ordovician strata in the eastern Pyrenees represent a  
786 quiescent time span with no remarkable tectonic activity. An unconformity-free monotonous  
787 succession of shales with subsidiary sandstone interbeds is recognized there as the Serdinya  
788 Formation (Jujols Group), with a characteristic sandstone-dominated upper part distinguished  
789 as the Font Frède Member (Padel *et al.* 2018). Acritarchs recovered from the uppermost part  
790 of the formation in the southern Canigó ~~massif~~Massif have yielded a broad Furongian–Early  
791 Ordovician age (Casas and Palacios 2012).

792

793 *The Sardic Phase*

794

795 In the Montagne Noire, Mouthoumet and eastern Pyrenees massifs, a late Early–Mid  
796 Ordovician phase of uplift, tilting and denudation of a broad palaeorelief subaerially exposed  
797 led to the Sardinian unconformity. Uplift was associated with magmatic activity, which  
798 continued until [the](#) Late Ordovician ~~times~~. Consequently, Middle Ordovician strata are absent  
799 in the Occitan Domain.

800 In the eastern Pyrenees, the Sardinian Phase is linked to a time gap of about 16–23 million  
801 years. Coeval with the late Early–Mid Ordovician phase of generalized uplift and denudation,  
802 key magmatic activity led to the intrusion of voluminous granitoids, about 500 to 3000 m  
803 thick and encased in Ediacaran–Cambrian Series 2 strata. These granitoids constitute the  
804 protoliths of the large orthogneissic laccoliths that punctuate the backbone of the central and  
805 eastern Pyrenees. They form, from west to east, the Aston (467–470 Ma; Denèle *et al.* 2009;  
806 Mezger and Gerdes 2016), Hospitalet (about 472 Ma; Denèle *et al.* 2009), Canigó (472–462  
807 Ma; Cocherie *et al.* 2005; Navidad *et al.* 2018), Roc de Frausa (477–476 Ma; Cocherie *et al.*  
808 2005; Castiñeiras *et al.* 2008) and Albera (about 470 Ma; Liesa *et al.* 2011) massifs (Fig. 8),  
809 which ~~comprise~~ [imply](#) a dominant Floian–Dapingian age. Felsic volcanic equivalents have  
810 been documented in the Albera Massif, where subvolcanic rhyolitic porphyroid rocks have  
811 yielded similar ages to those of the main gneissic bodies (about 474–465 Ma; Liesa *et al.*  
812 2011) (Fig. 2).

813 Magmatism persisted until the Late Ordovician, yielding another set of magmatic rocks  
814 that constitute the protoliths of the Cadí (ca. 456 Ma; Casas *et al.* 2010), Casemí (446 to 452  
815 Ma; Casas *et al.* 2010), Núria (ca. 457 Ma; Martínez *et al.* 2011) and Canigó G1-type (ca. 457  
816 Ma; Navidad *et al.* 2018) gneisses. The lowermost part of the Ediacaran Canaveilles Group  
817 hosts metre-scale thick bodies of metadiorite sills related to an Upper Ordovician protolith,  
818 (ca. 453 Ma; Casas *et al.* 2010). Coeval calc-alkaline ignimbrites, andesites and volcanoclastic  
819 rocks are interbedded in the Upper Ordovician succession of the Bruguera and Ribes de

820 Freser areas (Martí *et al.* 2019). In the latter area, a granitic body with granophyric texture,  
821 dated at ca. 458 Ma by Martínez *et al.* (2011), intruded at the base of the Upper Ordovician  
822 succession. In the La Pallaresa dome, some metre-scale rhyodacitic to dacitic subvolcanic  
823 sills, Late Ordovician in age (ca. 453 Ma; Clariana *et al.* 2018), occur interbedded within the  
824 pre-unconformity strata and close to the base of the Upper Ordovician.

825 In the Axial Zone of the Montagne Noire, some migmatitic orthogneisses, derived from  
826 Ordovician metagranites bearing large K-feldspar phenocrysts, were emplaced at about 471  
827 Ma (Somail orthogneiss; Cocherie *et al.* 2005), 456 to 450 Ma (Pont de-Larn and Gorges  
828 d'Héric gneisses, Roger *et al.* 2004) and ca. 455 Ma (Saint Eutrope Gneiss; Pitra *et al.* 2012)  
829 (Fig. 9). They intruded the “Schistes X” or St-Pons-Cabardès Group, a poorly constrained  
830 Ediacaran stratigraphic succession capped by the Sériès tuff, dated at about 545 Ma (Lescuyer  
831 and Cocherie 1992). The age of migmatization has been inferred from U–Pb dates on  
832 monazite from migmatites and anatectic granites at 333 to 327 Ma (Charles *et al.* 2009),  
833 which would represent a Variscan crustal melting event.

834

### 835 *Late Ordovician (post-Sardic) stratigraphic and volcano-sedimentary features*

836

837 Upper Ordovician–Lower Devonian rocks of the southern Montagne Noire and the  
838 Mouthoumet Massif rest paraconformably or with angular discordance on an inherited (pre–  
839 Sardic) Cambrian–Lower Ordovician palaeorelief. Upper Ordovician sedimentation started  
840 along the Cabrières and Mouthoumet rift branches with deposition of basaltic lava flows and  
841 lahar deposits (Roque de Bandies and Villerouge formations) of continental tholeiitic  
842 signature, indicative of continental fracturing (Álvaro *et al.* 2016). Overlying this tholeiitic  
843 volcanic episode, an Upper Ordovician stratigraphic succession can be recognized infilling  
844 these grabens and half-grabens, composed, from bottom to top, of sandstone complexes

845 (lower Katian Glauzy and Gascagne formations), shales interbedded by carbonate (upper  
846 Katian Gabian and Montjoi formations), and an incisive glaciogenic unconformity,  
847 represented by the Hirnantian Marmaraine diamictites in the Mouthoumet Massif (Fig. 9).  
848 The chronostratigraphic control of this succession is mainly based on brachiopods belonging  
849 to the *Nicolella* and *Hirnantia* communities (Colmenar *et al.* 2013; Álvaro *et al.* 2016).

850 A similar stratigraphic arrangement can be recognized in the eastern Pyrenees, where the  
851 Sardinian Phase was succeeded by a Late Ordovician extensional interval responsible for the  
852 opening of (half-)grabens infilled with the basal Upper Ordovician alluvial-to-fluvial  
853 conglomerates (La Rabassa Conglomerate Formation; Hartevelt 1970). Major variations in  
854 the thickness of the Upper Ordovician strata, along with drastic variations in grain size and  
855 thickness, are probably related to the development of palaeotopographies controlled by faults  
856 and the subsequent erosion of uplifted palaeoreliefs. Katian strata are represented by the  
857 shale-dominant Cava Formation and the shale/carbonate alternations of the Estana and El  
858 Baell formations (Puddu *et al.* 2018 and references therein). The Hirnantian glaciation is  
859 identified as an erosive unconformity, capped by the Ansovell black shales and the Quartzite  
860 Barr formations, finally sealed by Silurian sediments (Puddu *et al.* 2019).

861

## 862 **The Ordovician of the Maures Massif and Corsica (by Lefebvre)**

863

### 864 *General framework*

865

866 Before the opening of Mediterranean back-arc basins in the Miocene, both Corsica and  
867 Sardinia were adjacent to the Maures Massif in Provence, southern France (Corsini and  
868 Rolland 2009; Schneider *et al.* 2014; Franke *et al.* 2017; Gerbault *et al.* 2018; Álvaro *et al.*

869 2021; Loi *et al.* this volume). These three regions belong to the southern Variscan belt of  
870 Europe and share a particularly complex geology (Bellot 2005; Gerbault *et al.* 2018).

871

## 872 *Maures Massif*

873

874 The low-grade metamorphic Palaeozoic rocks exposed in the western part of the Maures  
875 Massif (Cap Sicié, Fenouillet, Maurette and Loli units) are generally considered as equivalent  
876 to those occurring in other areas of the external zone of the southern Variscan belt (e.g.  
877 Montagne Noire, Mouthoumet, Pyrenees, southwestern Sardinia) and belonging to the  
878 Gondwanan margin (Bellot 2005; Franke *et al.* 2017; Gerbault *et al.* 2018).

879 In the western part of the Maures Massif, the Palaeozoic succession is poorly preserved,  
880 and the precise age of most units, poorly constrained (Bellot 2005). Most units were assigned  
881 to the Silurian and, so far, no evidence of Ordovician rocks could be demonstrated (Gueirard  
882 *et al.* 1970; Paris *et al.* 1999; Bellot 2005; Schneider *et al.* 2014).

883 The central part of the Maures Massif (Collobrières, Bormes, Cap Nègre and Cavalaire  
884 units) consists of highly metamorphosed late Precambrian–lowermost Ordovician volcano-  
885 sedimentary and magmatic rocks, possibly equivalent to the succession occurring in the  
886 Nappe Zone of central Sardinia (Bellot 2005; Gerbault *et al.* 2018).

887 The eastern part of the Maures Massif (Cavalières and Petites Maures units) is  
888 characterized by high-grade (high pressure) Cambro-Ordovician metagranites and Ordovician  
889 metagabbros, suggesting possible affinities to northeastern Sardinia and southwestern Corsica  
890 (Bellot 2005; Faure *et al.* 2014; Schneider *et al.* 2014; Franke *et al.* 2017; Gerbault *et al.*  
891 2018; Álvaro *et al.* 2021; Loi *et al.* this volume).

892

## 893 *Corsica*



894

895 In northwestern Corsica, a non-metamorphic Palaeozoic succession is preserved in the  
896 Argentella area, near Galeria (Baudelot *et al.* 1976; Rossi *et al.* 1995, 2009; Barca *et al.* 1996;  
897 Paris *et al.* 1999; Faure *et al.* 2014; Avigad *et al.* 2018). This succession ~~is lying~~  
898 unconformably on a more than 1,000 m thick basement of Neoproterozoic rocks  
899 (amphibolites, ~~metagraywackes~~~~metagreywackes~~, micaschists and quartzites), polydeformed  
900 and metamorphosed during the Cadomian orogeny (Barca *et al.* 1996; Faure *et al.* 2014;  
901 Avigad *et al.* 2018). The overlying Palaeozoic succession comprises several informal  
902 stratigraphic units.

903 The lowermost unit (Ciuttone Sandstone) corresponds to about 50 m thick unfossiliferous,  
904 folded, grey sandstones containing clasts and pebbles reworked from underlying Precambrian  
905 micaschists (Baudelot *et al.* 1976; Barca *et al.* 1996; Avigad *et al.* 2018). A Cambrian to  
906 Early Ordovician age is generally assigned to the Ciuttone Sandstone based on a comparison  
907 with the eastern Pyrenees and Sardinia, both characterized by similar sandstone-dominated,  
908 pre-Sardic deposits (Barca *et al.* 1996; Paris *et al.* 1999; Rossi *et al.* 2009). This age is also  
909 congruent with dating obtained from detrital zircons (~555 Ma) suggesting that the Ciuttone  
910 Sandstone ~~resulted~~~~was derived~~ from weathered ~~late-upper~~ Ediacaran rocks (Avigad *et al.*  
911 2018).

912 The upper part of this unit is channeled by the eroding base of the overlying Monte  
913 Martinu Conglomerate, which corresponds to a ~100 m thick succession of conglomerates (at  
914 the bottom) and quartzites (at the top) (Baudelot *et al.* 1976; Barca *et al.* 1996; Avigad *et al.*  
915 2018). Zircons extracted in the upper part of the Monte Martinu Conglomerate yielded an age  
916 of 476 ±26 Ma (Early Ordovician), supporting the interpretation of this unit as consisting of  
917 Upper Ordovician (post-Sardic) deposits (Rossi *et al.* 1995; Avigad *et al.* 2018).

918 The Monte Martinu Conglomerate is tectonically overlain by the ~150 m thick Campu  
919 Orbu Shale. The dark shales of this unit were originally tentatively assigned to the Silurian,  
920 based on poorly preserved acritarchs and chitinozoans ~~remains~~ (Beaudelot *et al.* 1976).  
921 However, the occurrence of typical Hirnantian diamictites (20–30 m thick) in the overlying  
922 Capu Russellu unit suggests that the age of the Campu Orbu Shale has to be reconsidered and  
923 assigned to the Upper Ordovician (Rossi *et al.* 1995; Paris *et al.* 1999; Avigad *et al.* 2018).

924 The upper part of the Capu Russellu unit consists of 5–6 m of dark shales and lydites  
925 yielding typical assemblages of Rhuddanian and Aeronian (Silurian) graptolites (Štorch 1994;  
926 Barca *et al.* 1996; Paris *et al.* 1999; Avigad *et al.* 2018).

927

## 928 **The Ordovician of Schwarzwald and the Vosges (by Lefebvre and Servais)**

929

### 930 *General framework*

931

932 The Vosges Massif in eastern France and Schwarzwald (Black Forest) in southwestern  
933 Germany are parts of ~~a~~ the same geologically complex structure inherited from the Variscan  
934 orogeny and now occurring on the western and eastern shoulders of the Cenozoic Rhine  
935 Graben, respectively (Wickert and Eisbacher 1988; Servais *et al.* 2008). These two isolated  
936 massifs occupy a central position in the Variscan orogen belt between the French Massif  
937 Central and the Bohemian Massif (Vaida *et al.* 2004; Servais *et al.* 2008; Skrzypek *et al.*  
938 2014; Franke *et al.* 2017). The Vosges and Schwarzwald both consist of three similar litho-  
939 tectonic complexes, all intruded by Variscan granitic plutons: (1) a northern complex with  
940 successive belts of (meta-) sedimentary Palaeozoic rocks; (2) a central complex characterized  
941 by HP/HT polymetamorphic gneisses; and (3) a southern complex comprising a mélange of  
942 autochthonous and allochthonous (volcano-) sedimentary Palaeozoic rocks.

943 Since the pioneer work of Kossmat (1927), the northern complex has been traditionally  
944 assigned to the Saxothuringian Zone, whereas Moldanubian affinities were suggested for both  
945 the central and southern complexes (Wickert and Eisbacher 1988; Eisbacher *et al.* 1989;  
946 Fluck *et al.* 1991; Paris *et al.* 1999; Montenari *et al.* 2000; Vaida *et al.* 2004; Servais *et al.*  
947 2008). In this scheme, the Lalaye-Lubine fault zone (LLFZ) in the Vosges and its  
948 prolongation in Schwarzwald, the Baden-Baden fault zone, were interpreted as corresponding  
949 to the former boundary between the subducted Saxothuringian passive margin and the  
950 overlying Moldanubian plate (Wickert and Eisbacher 1988; Eisbacher *et al.* 1989). Recently,  
951 strong lithological affinities with the Teplá-Barrandian Zone (Bohemian Massif) were  
952 ~~evidenced~~ established for the northern complexes of the Vosges and Schwarzwald, suggesting  
953 that they were originally located south of the Saxothuringian-Moldanubian suture, and thus not  
954 part of the Saxothuringian Zone (Fig. 1; Skrzypek *et al.* 2014; Álvaro *et al.* 2021).

955

### 956 *Northern complex*

957

958 The northern complex is best exposed in the Vosges, where it consists of three successive,  
959 fault-bounded units: from North to South, the Bruche, Steige and Villé units (Wickert and  
960 Eisbacher 1988; Skrzypek *et al.* 2014). These three units ~~are showing~~ show an increasing  
961 metamorphic grade towards the LLFZ, which forms the southern boundary of the Northern  
962 Vosges (Wickert and Eisbacher 1988). The Bruche unit is composed of Middle Devonian to  
963 lower Carboniferous unmetamorphosed sedimentary rocks and volcanics (Skrzypek *et al.*  
964 2014). The Steige unit mainly consists of very low-grade (greenschist facies), red-coloured  
965 metasediments (mainly phyllites and metasandstones) preserving primary bedding structures  
966 and trace fossils (Clauer 1970; Reitz and Wickert 1989). In Andlau and Biarville, the Steige  
967 unit yielded two distinct, relatively diverse, poorly preserved chitinozoan assemblages

968 (Doubinger 1963; Doubinger and von Eller 1963). The Andlau assemblage includes  
969 *Conochitina brevis*, thus supporting a late Early to early Middle Ordovician age, whereas the  
970 Biarville fauna comprises younger (Silurian) chitinozoans (Doubinger 1963; Doubinger and  
971 von Eller 1963). The structurally deeper Villé unit corresponds to a thick series of low-grade  
972 (greenschist facies) monotonous clastic series of meta-sediments (phyllites, meta-sandstones,  
973 and meta-conglomerates), locally interrupted by a several-meter thick rhyolitic tuff (Reitz and  
974 Wickert 1989). Although primary sedimentary structures are not preserved due to incipient  
975 schistosity, trace fossils (*Tomaculum*) and putative sponge remains were reported in the Villé  
976 unit (Doubinger and von Eller 1963, 1967; Ross 1964). This unit also yielded a poorly  
977 preserved, though diverse acritarch assemblage suggesting a late Cambrian to Early  
978 Ordovician age (Reitz and Wickert 1989).

979 In the Schwarzwald, the northern complex is less extensive and restricted to the Baden-  
980 Baden–Gaggenau area (Eisbacher *et al.* 1989; Servais *et al.* 2008). It comprises a thick  
981 succession of low-grade (greenschist facies) rocks generally considered as equivalent to both  
982 the Villé and Steige units (Maas 1981; Montenari *et al.* 2000). This succession is usually  
983 subdivided into four 'series' (Sittig 1965; Montenari *et al.* 2000; Servais *et al.* 2008). The  
984 lowermost one (*Untere Schindelklamm Serie*) mainly consists of quartzitic siltstones and  
985 quartzites. It is overlain by the diabases (meta-volcanics) of the *Basischer Zug*. The third part  
986 (*Obere Schindelklamm Serie*) comprises greywackes, and it is overlain by the uppermost  
987 *Traischbach Serie* formed by a pelitic succession with intercalated carbonate olistoliths. An  
988 extensive palynological sampling through all four series yielded a single, poorly preserved  
989 acritarch assemblage in the uppermost part of the *Traischbach Serie* (Montenari and Servais  
990 2000; Montenari *et al.* 2000). This assemblage comprises the genera *Acanthodiacrodium*,  
991 *Caldariola*, *Cymatiogalea*, *Dasydiacrodium* and *Stelliferidium*, thus suggesting a late

992 Cambrian to Early Ordovician age, in good accordance with that obtained for the Villé unit in  
993 the Vosges (Montenari and Servais 2000; Montenari *et al.* 2000; Servais *et al.* 2008).

994

### 995 *Central and southern complexes*

996

997 Although the high-grade gneisses of the central complex in the Vosges and Schwarzwald  
998 have not yielded any fossils, an Early Palaeozoic (late Cambrian to Silurian) age for their  
999 sedimentary protoliths could be deduced from the included detrital zircons dated at 490 and  
1000 450 Ma (Skrzypek *et al.* 2014; Franke *et al.* 2017).

1001 In the Vosges, the southern complex comprises a thick series of autochthonous and  
1002 allochthonous Upper Devonian–lower Carboniferous sedimentary rocks intruded by granitic  
1003 plutons (Eisbacher *et al.* 1989; Skrzypek *et al.* 2014).

1004 In the Badenweiler-Lenzkirch area (Schwarzwald), the southern complex also includes  
1005 low-grade meta-sediments (greenschist facies; phyllites) structurally below the  
1006 unmetamorphosed Upper Devonian–lower Carboniferous succession. The age of the phyllites  
1007 from the Badenweiler-Lenzkirch area long remained poorly known, and they were tentatively  
1008 assigned to the Devonian (Schäfer 1957; Metz and Rein 1958) or to the lower Carboniferous  
1009 (Maas 1961; Krohe and Eisbacher 1988). The occurrence of putative remains of Ordovician  
1010 palynomorphs in this unit was first suggested based on poorly preserved material (Hann *et al.*  
1011 1995; Sawatzki *et al.* 1997). However, the description of better-preserved palynomorphs  
1012 finally confirmed the Early Palaeozoic age of the phyllites (Montenari and Maas 1996;  
1013 Montenari *et al.* 2000; Vaida *et al.* 2004; Servais *et al.* 2008). At least two distinct  
1014 fossiliferous lithostratigraphic units were documented in the Badenweiler-Lenzkirch area.  
1015 One consists of meta-greywackes and meta-pelites yielding an [Lower](#)~~Early~~–Middle  
1016 Ordovician acritarch assemblage, including the eponymous taxa of the *E. brevis* (~~late~~~~-~~[upper](#)

1017 Floian) and ?*C. protocalix* (Darriwilian) biozones (Vaida *et al.* 2004). The second level  
1018 corresponds to meta-siltstones containing a relatively diverse palynomorph assemblage  
1019 (acritarchs, chitinozoans and cryptospores) suggesting a Silurian (Wenlock–Ludlow) age  
1020 (Montenari and Maas 1996; Montenari *et al.* 2000; Vaida *et al.* 2004).

1021

## 1022 **Concluding remarks and perspectives**

1023

1024 In spite of a complex Variscan history and overprinting, the original depositional contexts  
1025 and palaeogeographic affinities of [the](#) Ordovician strata, now exposed in scattered massifs  
1026 throughout western Europe can be reconstructed, based on various and complementary  
1027 approaches relying e.g. on geodynamics (magmatic affinities, metamorphism, palaeomagnetic  
1028 data [and](#), tectonics), palaeontology (biostratigraphy, [and](#) faunal affinities), and sedimentology  
1029 (stratigraphic gaps, [and](#) proximal-distal gradients).

1030 In [the](#) Ordovician ~~times~~, the regions situated in present-day Belgium and neighbouring  
1031 areas of France and Germany (Brabant Massif, Condroz and Ardenne inliers), western  
1032 Germany (Rhenish Massif) and in the subsurface of northeastern Germany (Rügen) were all  
1033 part of the Avalonia microcontinent. In most of these areas, the Ordovician succession is  
1034 characterized by a ~13 [m.y.myr](#) hiatus extending from the mid Tremadocian and embracing  
1035 the whole Floian (Fig. 3), which is generally interpreted as resulting from the drift of  
1036 Avalonia away from Gondwana (Cocks and Torsvik 2002; Verniers *et al.* 2002; Linnemann *et*  
1037 *al.* 2012). This interpretation is confirmed by the major and relatively sudden shift in  
1038 depositional environments recorded in the Stavelot-Venn inlier, at the transition between  
1039 megasequences 1 and 2 (Fig. 3; Vanguetaine 1992; Verniers *et al.* 2002; Linnemann *et al.*  
1040 2012; Herbosch *et al.* 2016, 2020). In this area, instead of a hiatus, the Lower Ordovician  
1041 succession records a sharp transition from shallow shelf environmental conditions (upper

1042 Lierneux Member of the Jalhay Formation, upper Tremadocian) to bathyal settings (Otré  
1043 Formation, Floian). In the Brabant Massif deep and often anoxic environmental conditions  
1044 persisted throughout the Mid Ordovician and part of the Late Ordovician, whereas in the  
1045 Condroz, Rhenish Massif and Rügen the depositional environment was shallower on the shelf  
1046 (megasequence 2, Fig. 3; Verniers *et al.* 2002; Servais *et al.* 2008; Linnemann *et al.* 2012). In  
1047 most areas, the transition between megasequences 2 and 3 around mid-Katian times (Fig. 3)  
1048 marks a shift from deep and/or outer shelf settings to shallower environments and even  
1049 emersion in the Condroz, very likely related to the initial stages of the soft docking of eastern  
1050 Avalonia with Baltica (e.g. Verniers *et al.* 2002; Linnemann *et al.* 2012). The three  
1051 megasequences documented in Belgium, western and northeastern Germany are ~~very~~  
1052 comparable to those originally described ~~in~~ from the Welsh Basin and the English Lake  
1053 ~~district~~ District (Woodcock 1990; Molyneux *et al.* this volume), thus supporting close  
1054 palaeogeographic Avalonian affinities for all these regions (Katzung *et al.* 1995; Maletz,  
1055 2000; Verniers *et al.* 2002; Servais *et al.* 2008; Linnemann *et al.* 2012).

1056 With ~~the~~ only exception of the southern part of the Ardenne inliers, all Ordovician rocks  
1057 exposed in present-day France and western Germany (Schwarzwald) were originally  
1058 deposited on the southern margin of the Rheic Ocean, on the Gondwanan passive margin or  
1059 nearby (Fig. 1; Robardet *et al.* 1994b; Paris *et al.* 1999; Servais and Sintubin 2009; Pouclet *et*  
1060 *al.* 2017; Álvaro *et al.* 2021; Caroff, in press). However, reconstructing their original  
1061 depositional setting, relative position and relationship with other (peri-)Gondwanan regions  
1062 remains difficult. A close palaeogeographic link between the Armorican Massif and the  
1063 Iberian Peninsula has been long established based on strong lithological and faunistic  
1064 similarities (e.g. Paris and Robardet 1977, 1994; Robardet and Gutiérrez-Marco 1990;  
1065 Robardet *et al.* 1994b; ~~Gutiérrez-Marco *et al.* this volume~~). Probable affinities between these  
1066 two regions are further supported by the occurrence of a similar Furongian–mid Floian hiatus

1067 (Fig. 6). Designated ~~the~~ "lacune normande" in the Armorican Massif (Le Corre *et al.* 1991)  
1068 and "Toledonian gap" in the Iberian Massif (Alvaro *et al.* 2021), this ~22 ~~m.y.~~myr long hiatus  
1069 possibly resulted, in both areas, from the uplift of Ediacaran–Cambrian rocks (Sánchez-  
1070 García *et al.* 2019). This "Norman hiatus" has no equivalent in Corsica, the Occitan Domain  
1071 (Montagne Noire and Mouthoumet massifs), eastern Pyrenees and in the Vosges-  
1072 Schwarzwald massifs.

1073 In marked contrast with the ~~closeness~~ close proximity of the Armorican Massif and the  
1074 Iberian Massif, ~~a~~ continuous shallow shelf sedimentation is recorded in the Lower Ordovician  
1075 successions of northwestern Corsica (Avigad *et al.* 2018), the northern and southern  
1076 Montagne Noire (Fig. 9; Vizcaïno *et al.* 2001), the Mouthoumet Massif (Colmenar *et al.*  
1077 2013), as well as in other areas (e.g. Maures Massif ~~and~~; ~~western~~ Western Alps), yielding less  
1078 well-preserved, metamorphosed rocks (Bellot 2005; Guillot and Ménot 2009). Moreover, the  
1079 Ordovician successions of Corsica and all regions of the Occitan Domain are characterized by  
1080 a major hiatus extending from the Dapingian to the early Katian boundary (Fig. 9). Several  
1081 authors (Pouclet *et al.* 2017; Alvaro *et al.* 2021) proposed to link this stratigraphic gap to the  
1082 Sardinic phase, originally described in the Sulcis-Iglesiente area of southwestern Sardinia (see  
1083 Loi *et al.* this volume). The occurrence of a similar stratigraphic unconformity (~~mimicking~~  
1084 mirroring the Sardinic unconformity) in northwestern Corsica, the Occitan Domain, eastern  
1085 Pyrenees and southwestern Sardinia thus strongly suggests that all these areas were  
1086 geodynamically linked and palaeogeographically close in ~~the~~ Ordovician ~~times~~. In marked  
1087 contrast with the situation in these areas, a continuous shelf sedimentation is recorded in the  
1088 Middle and Upper Ordovician successions of the Armorican Massif (Fig. 6).

1089 It is thus possible to identify two main "clusters" of Ordovician rocks in France. The first  
1090 one, which comprises the Armorican Massif, is characterized by strong affinities with the  
1091 Iberian Massif, a Furongian–mid Floian "Norman gap", and a continuous shelf sedimentation



1092 in the Mid and Late Ordovician (Fig. 6). The second cluster, which includes northwestern  
1093 Corsica, the Montagne Noire, the Mouthoumet Massif, the eastern Pyrenees and possibly the  
1094 Maures Massif, can be identified based on strong affinities with Sardinia (Sulcis-Iglesiente  
1095 unit, Sarrabus and Gerrei units), a continuous shelf sedimentation during the Early  
1096 Ordovician, and an extensive Dapingian–early Katian Sardinic(?) gap (Fig. 9). However, in [the](#)  
1097 late Katian ~~times~~, regions belonging to both clusters are characterized by a similar shift from  
1098 siliciclastic to carbonate sedimentation, and a major faunal transition from cool-water adapted  
1099 assemblages to more temperate ones (see e.g. Dégardin *et al.* 1995; Colmenar *et al.* 2013;  
1100 Lefebvre *et al.* 2022). This major environmental change, which is also recorded in other high-  
1101 latitude (peri-)Gondwanan regions (e.g. Algeria, Carnic Alps, Iberian Peninsula, eastern  
1102 Tafilalt area in Morocco, ~~and~~ [Sardinia](#)), probably results from a short episode of global  
1103 warming: the "Boda Event" (Villas *et al.* 2002; Fortey and Cocks 2005; Ferretti *et al.* this  
1104 volume).

1105 In the Vosges and Schwarzwald massifs, Ordovician rocks are too poorly preserved to  
1106 establish any clear relationship with better-known successions from the Armorican-Iberian or  
1107 the Occitan-Sardinian clusters. In the northern complex of the Vosges-Schwarzwald massifs,  
1108 the identification of Furongian–Early Ordovician acritarch assemblages in both the Villé  
1109 Formation and the Traisbach Serie indicates, at least, the absence of a "Norman gap" in  
1110 these levels. Similarly, the occurrence of typical late Floian and Darriwilian chitinozoans in  
1111 the Badenweiler-Lenzkirch area suggests that no "Sardinic gap" occurs in the Ordovician  
1112 succession of southern Schwarzwald. Correlation of these two isolated massifs with other  
1113 Variscan areas thus remains problematic, and future work will probably help clarify if the  
1114 northern complex of the Vosges-Schwarzwald massifs belongs to the Saxothuringian Zone  
1115 (Wickert and Eisbacher 1988; Eisbacher *et al.* 1989) or to the Teplá-Barrandian Zone  
1116 (Skrzypek *et al.* 2014; Álvaro *et al.* 2021). Moldanubian affinities for the central and southern

1117 Vosges-Schwarzwald massifs seem to be better established (Eisbacher *et al.* 1989; Skrzypek  
1118 *et al.* 2014).

1119 Other West-European areas, not treated here, are in a serious need of reevaluation of their  
1120 Ordovician rocks. One of the most extensive Ordovician successions is probably the  
1121 particularly thick one occurring in the subsurface of the Aquitaine Basin (Paris and Le Pochat,  
1122 1994; Robardet *et al.*, 1994b). Preliminary investigations of boreholes suggests the occurrence  
1123 of over 1,200 m of siliciclastic rocks dated from the Lower and Middle Ordovician, based on  
1124 acritarchs, chitinozoans, graptolites, and trilobites. The presence of Upper Ordovician rocks is  
1125 less poorly constrained, but typical Hirnantian diamictites ~~would be~~ present (Robardet *et*  
1126 *al.* 1994b). Other poorly known regions in serious need of a reevaluation of their Ordovician  
1127 rocks include e.g. the central Pyrenees (see e.g. Robardet *et al.* 1994b; Dégardin *et al.* 1995),  
1128 the southeastern-most extremity of the SAD (Chantonnay and Vendée areas; see Pouclet *et al.*  
1129 2017) or the meta-sediments of the (French) western Alps (Guillot and Ménot 2009).  
1130 However, the Montagne Noire is an iconic example illustrating that, even in better-known and  
1131 "classical" areas, much work still remains to be done. In this region, the trilobite-based  
1132 biostratigraphic framework (Fig. 9) still needs to be carefully correlated with more "standard"  
1133 regional or international biostratigraphic charts based on acritarchs, chitinozoans, conodonts  
1134 or graptolites.

1135

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1137

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1150

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2059 368.

2060

2061

## 2062 **Figure captions**

2063

2064 **Fig. 1.** Tectonostratigraphic subdivision of the Variscan Orogen in southwestern Europe with  
2065 pre-Variscan setting of the Pyrenees between the Montagne Noire and Sardinia, and: (1)  
2066 Variscan parautochthon; (2) Moldanubian Zone and its prolongation; (3) Teplá-Barrandian  
2067 Zone and its prolongations; (4) Saxothuringian Zone; (5) Rheno-Hercynian Zone; (6)  
2068 Avalonian parautochthon; (7) oceanic sutures; Abbreviations: AM, Armorican Massif; Ard,



2069 Ardenne; BM, Brabant Massif; Co, Corsica; EP, Eastern Pyrenees; IS; Iapetus suture; Mau,  
2070 Maures Massif; MN, Montagne Noire; Mou: Mouthoumet Massif; RM: Rhenish Massif; RS:  
2071 Rheic suture; Rü: Rügen; SW: Schwartzwald; TS: Thor suture; Vo: Vosges. Adapted and  
2072 modified from Ballèvre *et al.* (2009), Torsvik and Cocks (2013), Skrzypek *et al.* (2014), and  
2073 Álvaro *et al.* (2021).

2074

2075 **Fig. 2.** Simplified geological map of the southern part of Belgium and neighbouring countries  
2076 showing the main tectonostratigraphic units of the Palaeozoic. To the North, the Brabant  
2077 Parautochthon and/or the Haine-Sambre-Meuse Overturned Thrust sheets are separated from  
2078 the Ardenne Allochthon by the Midi-Aachen Thrust Fault. The four Ardenne inliers, Stavelot-  
2079 Venn, Rocroi, Givonne and Serpont, are represented in purple. Modified from Herbosch *et al.*  
2080 (2020).

2081

2082 **Fig. 3.** Ordovician basement stratigraphy of Belgium and neighbouring countries.  
2083 Chronostratigraphy, graptolite and chitinozoan biozones after Goldman *et al.* (2020, fig. 20.2–  
2084 4, column Baltica for chitinozoan and Britain for graptolite zonations). References are  
2085 provided in the text for the bio- and chronostratigraphic position of the formations belonging  
2086 to the Brabant Massif, the Condroz, Rocroi and Stavelot-Venn inliers. Dotted line: boundary  
2087 between formations less precise. Lithostratigraphic units indicated in plain style are members,  
2088 and those in bold, formations. Abbreviations: Ser., Series; Meg., Megasequences.

2089

2090 **Fig. 4.** Ordovician stratigraphy of the Ebbe Inlier (Rhenish Massif, western Germany) and the  
2091 subsurface of Rügen Island (northeastern Germany). Chronostratigraphy, graptolite and  
2092 chitinozoan biozones after Goldman *et al.* (2020, fig. 20.2–4, column Baltica for chitinozoan  
2093 and Britain for graptolite zonations). References are provided in the text for the bio- and

2094 chronostratigraphic position of the formations. Dotted line: boundary between formations less  
2095 precise. All lithostratigraphic units reported (in bold) are formations. Abbreviation: Ser.,  
2096 Series.

2097

2098 **Fig. 5.** Simplified geological map of the Armorican Massif (western France) illustrating its  
2099 structuration into four distinct domains separated by two main late Carboniferous (Variscan)  
2100 shear zones: the North Armorican Shear Zone (NASZ) and the South Armorican Shear Zone  
2101 (SASZ). All main areas yielding Ordovician rocks mentioned in the text and/or reported on  
2102 Fig. 6 (chronostratigraphic chart) are indicated on the map by numbers, with: (1) Cotentin; (2)  
2103 Zone bocaine; (3) Domfront; (4) Laval; (5) Menez-Belair; (6) Châteaulin; (7) Crozon; (8)  
2104 South of Rennes; (9) Angers; (10) Saint-Julien-de-Vouvantes; (11) Saint-Georges-sur-Loire;  
2105 (12) Ancenis. Modified from Henry (1980), Ballèvre *et al.* (2009), Vidal *et al.* (2011b), and  
2106 Paris (2016).

2107

2108 **Fig. 6.** Ordovician stratigraphy of the Armorican Massif (western France).  
2109 Chronostratigraphy and high latitude Gondwanan chitinozoan biozones after Goldman *et al.*  
2110 (2020). References are provided in the text for the bio- and chronostratigraphic position of the  
2111 formations belonging to the NAD (Domfront), the MAD (Laval, Crozon, South of Rennes,  
2112 Angers) and the SAD (Ancenis). Dotted line: boundary between formations less precise.  
2113 Lithostratigraphic units indicated in plain style are members, and those in bold, formations.  
2114 Abbreviation: Ser., Series.

2115

2116 **Fig. 7.** Sequence stratigraphy of the Postolonnec Formation, Crozon, western part of the  
2117 Median Armorican Domain (from Dabard *et al.* 2015). (A) Lithostratigraphic column with  
2118 depositional environments deduced from sedimentological features and Gamma-ray data.

2119 Very high frequency (purple and red curves), high frequency (blue curve) and low frequency  
2120 (pale-blue curve) sequences. **(B)** Time-linear sea-level curve calibrated from the graptolites  
2121 biozones (i.e. *D. artus* at the top of the Kerloc'h Member and *G. linnarsoni* in Morgat  
2122 Member). On this figure the report of the chitinozoan biozones highlights the long-time  
2123 duration of the *L. pissotensis* Biozone, as recorded in the Crozon succession. Abbreviations:  
2124 Fm., Formation; L. Off., Lower Offshore; Mb., Member.

2125

2126 **Fig. 8.** Simplified geological map of the Occitan Domain (Albigeois, Montagne Noire,  
2127 Mouthoumet Massif), and eastern Pyrenees in southern France and neighbouring countries,  
2128 with: (1) Albera (Albères) Massif; (2) Roc de Frausa (Roc de France) Massif; (3) Canigó  
2129 (Canigou) Massif; (4) Hospitalet Massif; (5) Aston Massif. Modified from Touzeau *et al.*  
2130 (2012), Pouclet *et al.* (2017), and Padel *et al.* (2018). Other regions of the Occitan Domain  
2131 mentioned in the text are not shown here, because they are lying further North (i.e. Thiviers-  
2132 Payzac and Génis, Rouergue, and southern Cévennes massifs).

2133

2134 **Fig. 9.** Ordovician stratigraphy of the Montagne (southern France). Chronostratigraphy after  
2135 Goldman *et al.* (2020), and trilobite biostratigraphy based on Vizcaïno and Álvaro (2003),  
2136 Tortello *et al.* (2006), Serpagli *et al.* (2007), and Colmenar *et al.* (2013). References are  
2137 provided in the text for the bio- and chronostratigraphic position of the formations. Dotted  
2138 line: boundary between formations less precise. Key to trilobite generic names: *Proteuloma*  
2139 *geinitzi*, *Shumardia* (*C.*) *pusilla*, *Euloma* *filacovi*, *Taihungshania* *miqueli*, *Thaihungshania*  
2140 *shui landeyranensis*, *Colpocoryphe* *maynardensis*, *Neseuretus* (*N.*) *arenosus*, *Apatokephalus*  
2141 *incisus*, *Hangchungolithus* *primitivus*, and *Calymenella* *boisseli*. Lithostratigraphic units  
2142 indicated in plain style are members, and those in bold, formations. Abbreviation: Ser., Series.

Figure 1

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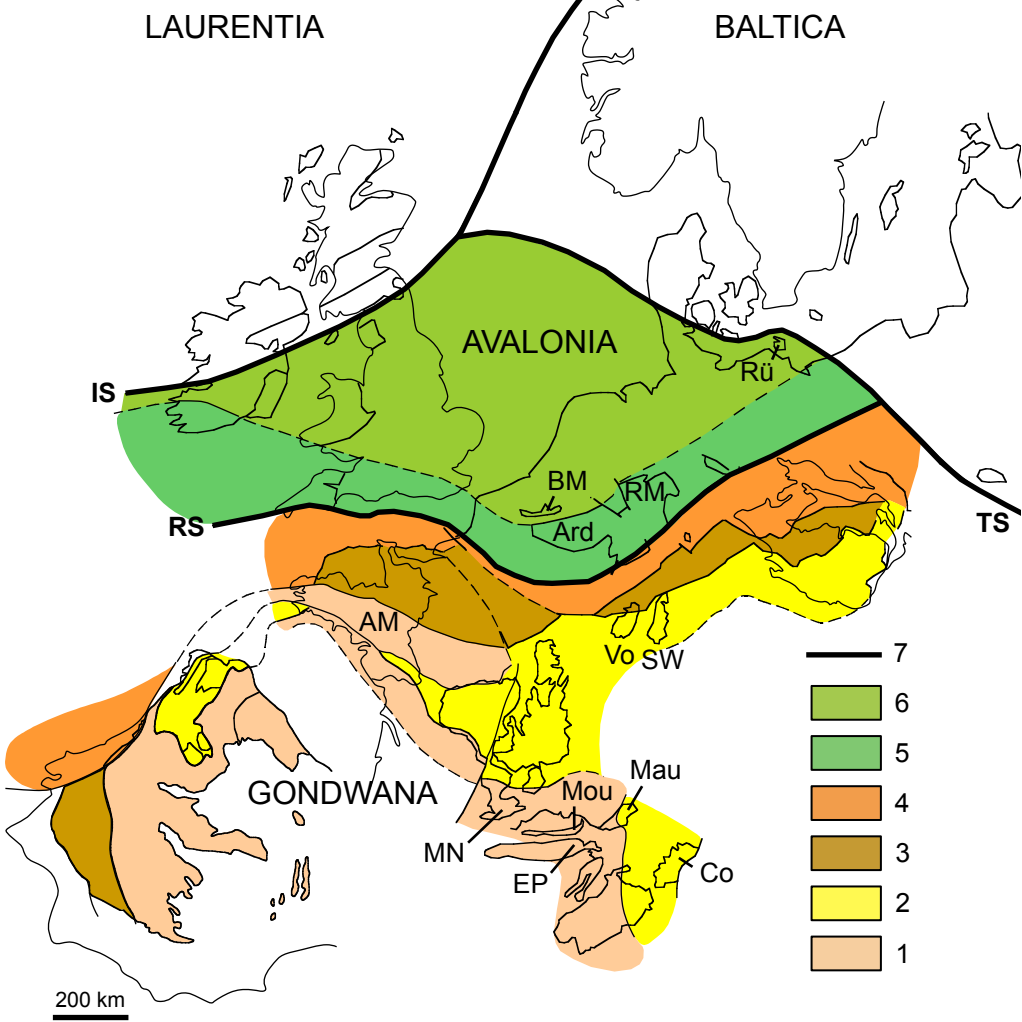


Figure 2

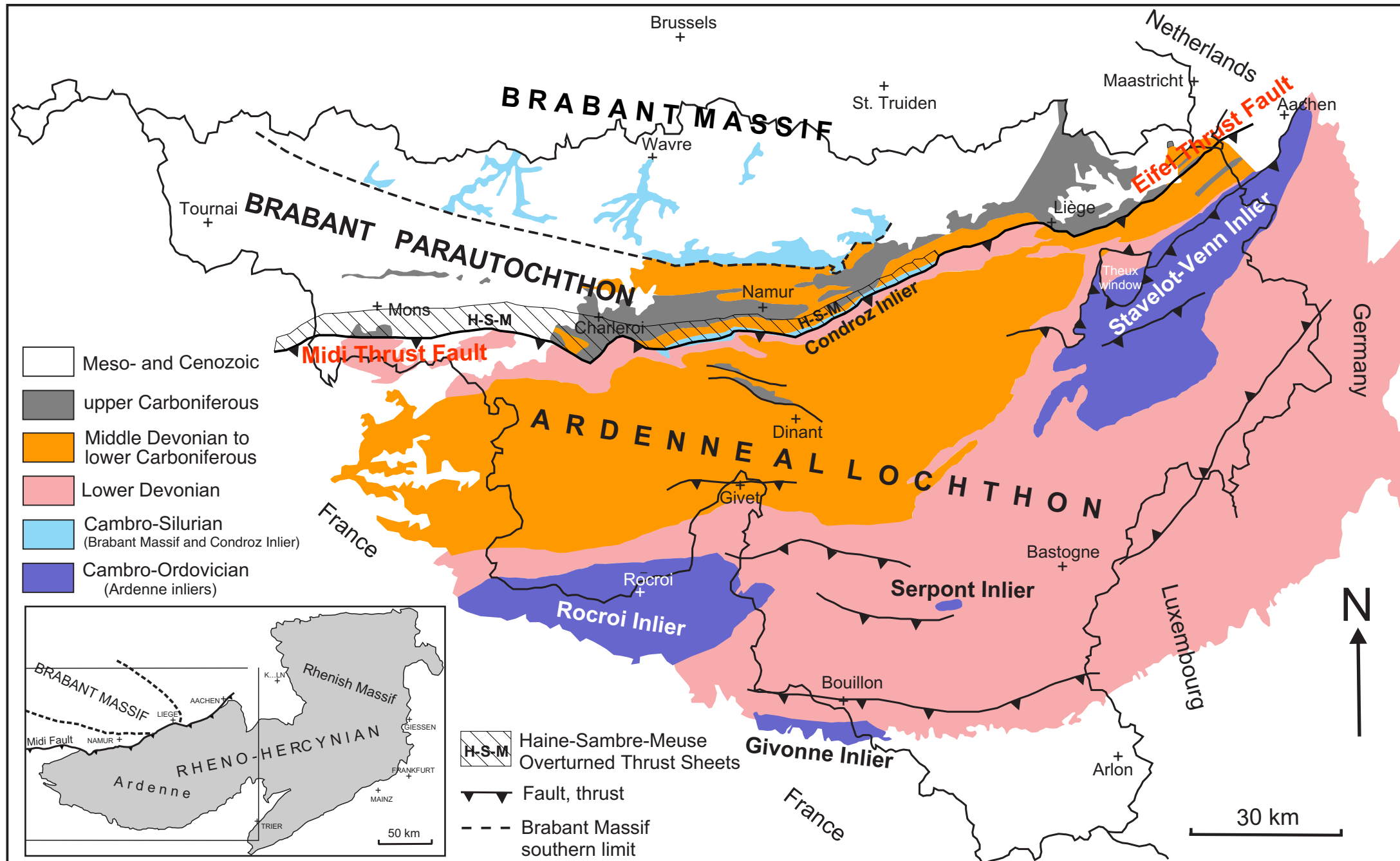


Figure 3  
Age (Ma)

Age (Ma)	Chronostratigraphy		Biozones		Lithostratigraphy						
	Ser.	Stages	Graptolites	Chitinozoans	Brabant Massif	Condroz Inlier	Meg.	Rocroi Inlier	Stavelot-Venn Inlier		
445	UPPER	Hirnantian	<i>M. persculptus</i>	<i>S. taugourdeai</i>	Brutia (pars) Gouteux	Fosses (pars)	Megaseq. 3	HIATUS	HIATUS		
			<i>M. extraordinarius</i>	<i>B. gamachiana</i>						Tihange (pars)	
		Katian	<i>D. anceps</i>	<i>T. anticostiensis</i>	Madot	Faux les Tombes					
			<i>C. rugata</i>	<i>C. rugata</i>						Bois de Presles	
450	Katian	<i>D. complanatus</i>	<i>T. bergstroemi</i>	Huet + Fauquez	Vitrival-Bruyère	Megasequence 2	HIATUS	HIATUS			
		<i>P. linearis</i>	<i>F. spinifera</i>	Cimetière G-Manil					Rue de Courrière		
455	Sandbian	<i>M. foliaceus/ C. bicornis</i>	<i>S. cervicornis</i>	Ittre	Sart-Bernard				La Bruyère	HIATUS	HIATUS
						<i>L. dalbyensis</i>					
						<i>A. granulifera</i>					
460	Darriwilian	<i>N. gracilis</i>	<i>L. stentor</i>	Rigenée	Chevreuil	Giroux	HIATUS	HIATUS			
									<i>H. teretiusculus</i>	L. striata	observational gap?
465	Darriwilian	<i>D. murchisoni</i>	<i>D. artus</i>	Tribotte	Huy	Vieux Moulin de Thilay	Bihain	Salmchâteau			
								<i>C. regnelli</i>	<i>C. cucumis</i>	Abbaye de Villers	Ri d'Honeux
470	Dapingian	<i>A. cucullus</i>	<i>C. cucumis</i>	HIATUS	HIATUS	HIATUS	Ottre	Colenhan			
								<i>T. gibberulus</i>	<i>E. primitiva</i>	HIATUS	Ottre
475	Floian	<i>I. v. victoriae</i>	<i>E. primitiva</i>	HIATUS	HIATUS	HIATUS	Ottre	Meuville			
		<i>E. simulans</i>						<i>T. phyllograptoides</i>	HIATUS	Ottre	
480	Tremadocian	<i>B. varicosus</i>	<i>E. primitiva</i>	HIATUS	HIATUS	HIATUS	Ottre	Jalhay			
		<i>A. murrayi</i>							<i>L. destombesi</i>	Chevripont	Chevripont
485	Tremadocian	<i>A. tenellus</i>	<i>L. destombesi</i>	Chevripont	Chevripont	Megasequence 1	observational gap?	Jalhay			
		<i>R. anglica</i>							Spa		
		<i>A. matanensis</i>								Solwaster	
		<i>R. parabola</i>									
<i>R. praeparabola</i>	Mousty (pars)	HIATUS									

Figure 4

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Age (Ma)	Chronostratigraphy		Biozones		Lithostratigraphy		
	Ser.	Stages	Graptolites	Chitinozoans	Ebbe Inlier	Rügen	
445	UPPER	Hirnantian	<i>M. persculptus</i>	<i>S. taugourdeoui</i>	Ebbe Inlier	Rügen	
			<i>M. extraordinarius</i>	<i>B. gamachiana</i>			
		Katian	<i>D. anceps</i>	<i>T. anticostiensis</i>			HIATUS
			<i>D. complanatus</i>	<i>C. rugata</i>			
			<i>P. linearis</i>	<i>T. bergstroemi</i>			
			<i>D. clingani</i>	<i>F. spinifera</i>			
			<i>M. foliaceus/ C. bicornis</i>	<i>S. cervicornis</i>			
		Sandbian	<i>B. hirsuta</i>	HIATUS			
			<i>L. dalbyensis</i>				
			<i>A. granulifera</i>				
450	MIDDLE	Darriwilian	<i>N. gracilis</i>	<i>L. stentor</i>	Rahlenberg Grauwackenschiefer	Arkona Schwartzschiefer / Nobbin Grauwacken	
			<i>H. teretiusculus</i>	L. striata	HIATUS		
		<i>D. murchisoni</i>	Kiesbert Tonschiefer				
		<i>D. artus</i>	<i>C. regnelli</i>	Plettenberg Bänderschiefer			
		Dapingian	<i>A. cucullus</i>	<i>C. cucumis</i>	HIATUS		
			<i>I. gibberulus</i>	Ebbe Inlier			
			<i>I. v. victoriae</i>				
		Floian	<i>E. simulans</i>		Ebbe Inlier		
			<i>B. varicosus</i>				
			<i>T. phyllograptoides</i>				
<i>E. primitiva</i>							
<i>A. murrayi</i>							
475	LOWER	Tremadocian	<i>L. destombesi</i>	HIATUS			
			<i>A. tenellus</i>				
		<i>R. anglica</i>	Varnkevitz Sandstein				
		<i>A. matanensis</i>					
		<i>R. parabola</i>					
		<i>R. praeparabola</i>					
		<i>R. praeparabola</i>					
480							
485							

Figure 5

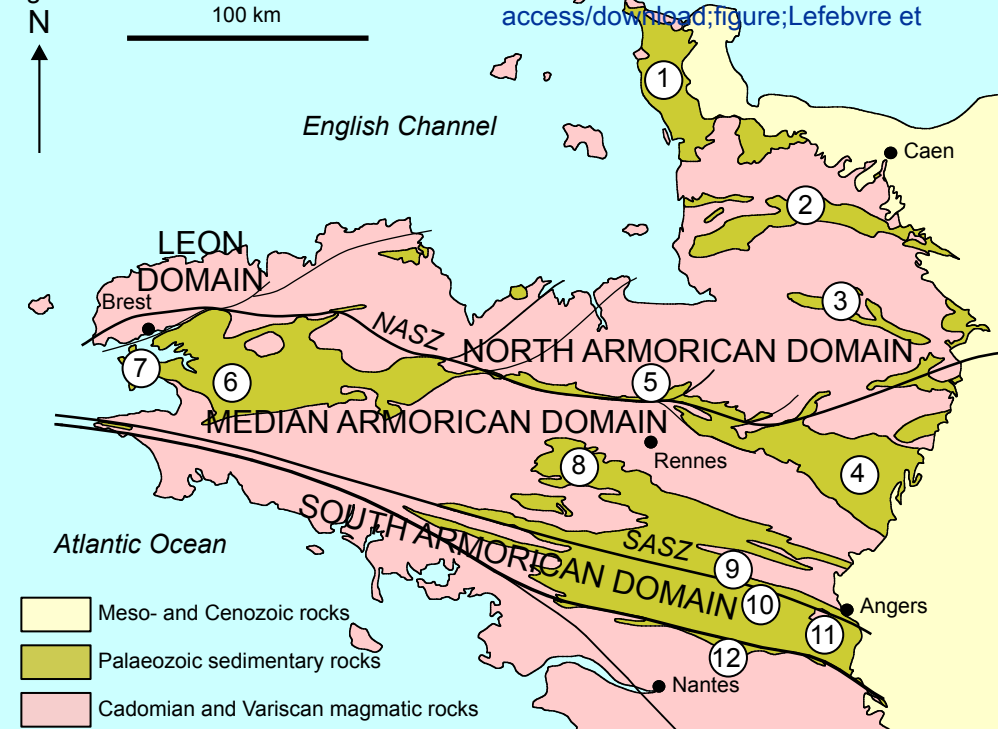
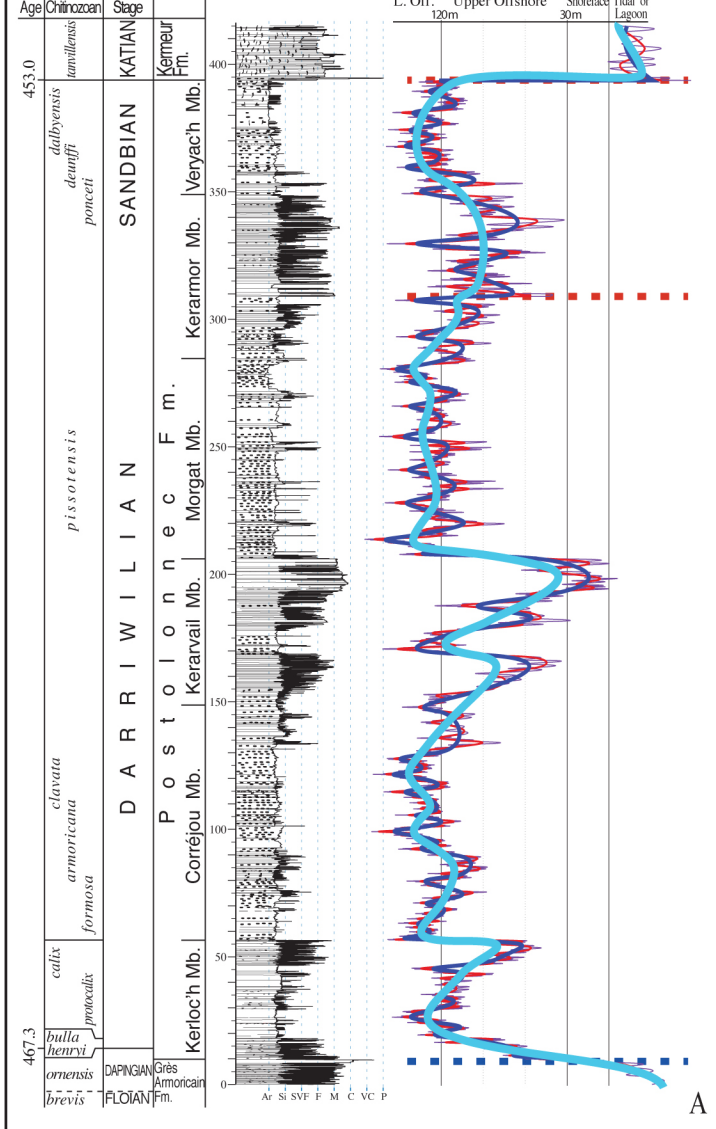


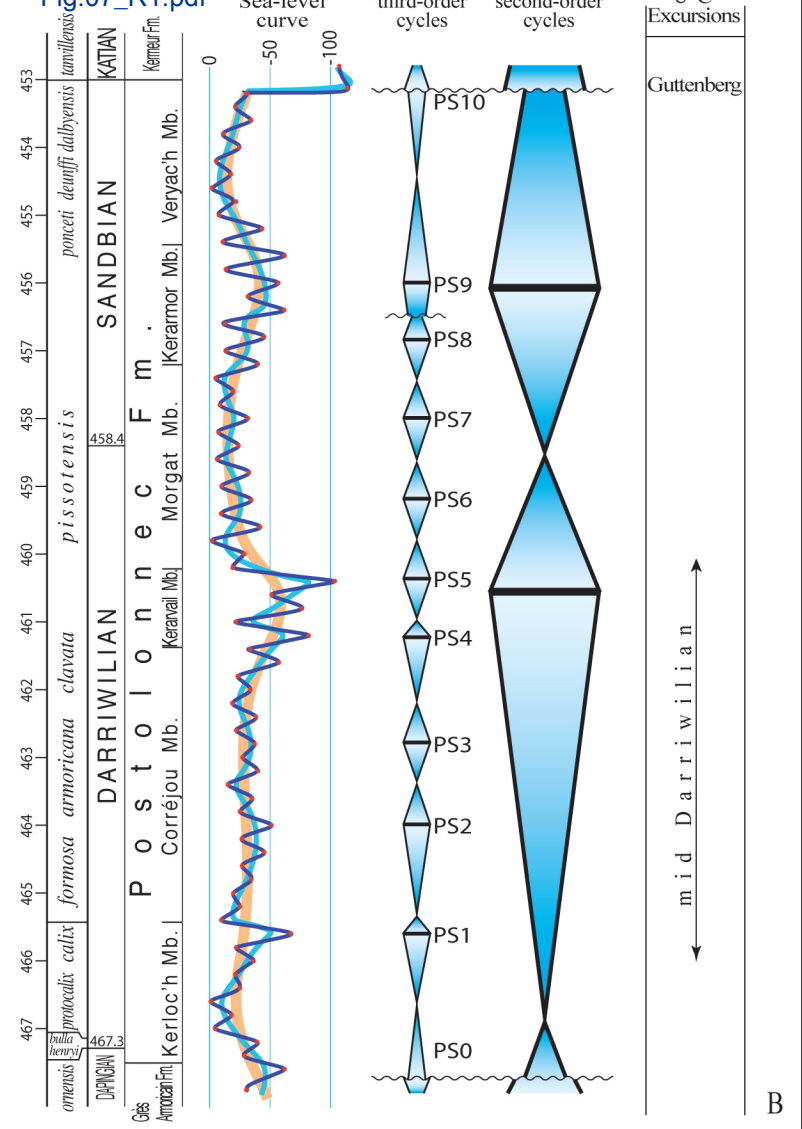




Figure 7



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B

Figure 8  
50 km

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