The Ordovician of France and neighbouring areas of Belgium and Germany

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

In western Europe, Ordovician sedimentary rocks generally occur as scattered, disconnected massifs or inliers surrounded and/or hidden by Mesozoic and/or Cenozoic cover. Reconstructing their original layout, stratigraphy and palaeogeographic relationships with other regions has been a major challenge during recent decades. All Ordovician strata occurring in present-day Belgium, France and Germany have undergone a long and complex journey. They were all deposited on a Cadomian or Cambrian basement with clear Gondwanan affinities. In the Furongian–Early Ordovician, the onset and diachronic propagation, from present-day SW to NE, of a major rift along the western margin of Gondwana gave rise to the Rheic Ocean (e.g. Linnemann *et al.* 2008; von Raumer and Stampfli 2008; Cocks and Fortey 2009; Stampfli *et al.* 2011). The opening of the Rheic Ocean resulted in the separation of Avalonia, as an independent microcontinent, away from Gondwana (e.g. Prigmore *et al.* 1997; Verniers *et al.* 2002; Cocks and Fortey 2009; Servais and Sintubin 2009; Domeier 2016). In the Ordovician, several areas included in this chapter thus belonged to the eastern part of Avalonia: the Brabant

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Massif and Ardenne inliers in Belgium, the Rhenish Massif in western Germany and Rügen Island, in northeastern Germany (Fig. 1). In contrast, the Armorican Massif, the Occitan Domain (Montagne Noire and Mouthoumet massifs), eastern Pyrenees, the Maures Massif, Corsica, and the Vosges (in France) and Schwarzwald (in western Germany) were located on the opposite (southern) side of the Rheic ocean, still forming part of Gondwana (Fig. 1). Avalonia drifted northwards, away from Gondwana, before docking with Baltica in the Late Ordovician (closure of the Tornquist Ocean; TS in Fig. 1), and slightly later with Laurentia, in the Silurian (closure of the Iapetus Ocean; IS in Fig. 1), resulting in the Caledonian orogeny (e.g. Verniers *et al.* 2002; Cocks and Fortey 2009).

In the late Paleozoic, the closure of the Rheic Ocean led to the formation of an extensive orogenic belt extending from the Appalachian and Ouachita mountains (to the West) to the Caucasus (to the East). This major continental collision or 'Variscan orogeny' between Laurentia–Avalonia–Baltica (to the north) and Gondwana (to the south) was particularly complex, because of the occurrence of numerous terranes, related to back-arc basins and narrow oceanic domains on both sides of the closing Rheic Ocean (Pharaoh 1999; Matte 2001; Ballèvre *et al.* 2009; Faure *et al.* 2009; Guillot and Ménot 2009; Stampfli *et al.* 2011; von Raumer *et al.* 2013). A consequence of the Variscan 'collage' was to bring together, on both sides of the Rheic suture (RS on Fig. 1), most pieces of the jigsaw forming present-day Europe, and extending from the Iberian Massif (see Gutiérrez-Marco *et al.* 2002) to the Alps (Ferretti *et al.* 2023) and the Bohemian Massif (Kraft *et al.* 2023). With the exception of the Brabant Massif, all Ordovician rocks now exposed in western Europe were more or less affected by the Variscan orogeny. For example, in some areas (e.g. Maures Massif, central zones of the Vosges and Schwarzwald), they endured medium- to high-grade metamorphism (Faure *et al.* 2014; Skrzypek *et al.* 2014; Gerbault *et al.* 2018). Even in regions exposing Ordovician rocks with very-low- to low-grade metamorphism, the strata were more or less affected by Variscan tectonics, involving folds and/or displacements along several tens of kilometres along shear zones or thrusts (e.g. Ballèvre *et al.* 2009; Pouclet *et al.* 2017; Herbosch *et al.* 2020). Moreover, in some areas (e.g. Alps, eastern Pyrenees), Variscan structures were reactivated and/or overprinted during the Alpine orogeny, thus making their interpretation even more difficult.

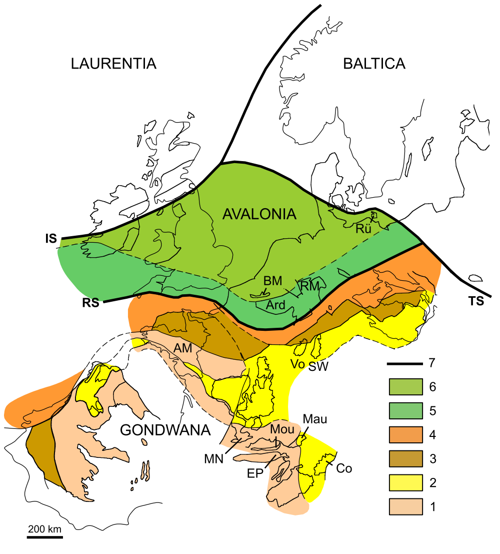

Fig. 1. Tectonostratigraphic subdivision of the Variscan Orogen in southwestern Europe with pre-Variscan setting of the Pyrenees between the Montagne Noire and Sardinia, and: (1) Variscan Parautochthon; (2) Moldanubian Zone and its prolongation; (3) Teplá–Barrandian Zone and its prolongations; (4) Saxothuringian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

During the last 30 years, the major advances in biostratigraphy (relying mostly on acritarchs, chitinozoans and graptolites, less frequently on brachiopods, conodonts and trilobites) have greatly contributed to the understanding of the complex pre-Variscan history of Western Europe, leading to a wide (and sometimes contradictory) set of palaeogeographic scenarios and regional syntheses, as e.g. in the Armorican Massif (Le Corre *et al.* 1991; Paris and Robardet 1994; Robardet *et al.* 1994*a*; Ballèvre *et al.* 2009), Belgium (Verniers *et al.* 2002; Linnemann *et al.* 2012; Herbosch *et al.* 2020), Corsica (Faure *et al.* 2014; Avigad *et al.* 2018), the Ebbe inlier (e.g. Maletz 2000; Koch *et al.* 2014), the Maures Massif (Bellot 2005), the Occitan Domain (Dégardin *et al.* 1995; Laumonier *et al.* 1995; Pouclet *et al.* 2017) or the Vosges (Skrzypek *et al.* 2014). A few more integrative reviews have been produced, and include syntheses on the Ordovician of France (Paris *et al.* 1999; Nardin *et al.* 2014), Belgium and France (Robardet *et al.* 1994*b*), or Belgium and Germany (Verniers *et al.* 2002; Servais *et al.* 2008). The aim of this chapter is thus to provide an updated overview of the Ordovician of Belgium, France and both western and northeastern Germany through a small number of selected regions, and to correlate their sedimentary successions with both global subdivisions (series and stages) of the International Chronostratigraphic Chart and the biostratigraphic time scales elaborated by Goldman *et al.* (2020) for the Ordovician.

The Ordovician of Belgium and neighbouring areas (by Herbosch, Verniers and Servais)

General framework

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

The Brabant Massif

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

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

Megasequence 2 begins with the Abbaye de Villers and Tribotte formations consisting of silty slates (Herbosch and Verniers 2014). They are correlated with the upper Dapingian to the lower Darriwilian by acritarchs, which belong to the *F. hamata–S. rarrigulata* Biozone (Vanguestaine and Wauthoz 2011). This acritarch biozone can be correlated with the *I. gibberulus* and *A. cucullus* graptolite biozones (Cooper and Molyneux 1990). The overlying Rigenée Formation, formed by dark silty slate, records a regional transgression (Paris *et al.* 2007) and is dated by graptolites (*Didymograptus artus* and *D. murchinsoni*; Maletz and Servais 1998) and acritarchs (*Arkonia virgata, Frankea hamulata and F. sartbernardensis*; Servais 1991, 1993) to most of the Darriwilian (Herbosch and Verniers 2014, fig. 3). The Ittre Formation is formed by an alternation of fine sandstones, siltstones and slates interpreted as distal turbidites (Servais 1991). It was dated by graptolites (*Nemagraptus gracilis* and *Mesocograptus foliaceus*; Maletz and Servais 1998) and chitinozoans (*S. cervicornis* Biozone; Vanmeirhaeghe 2006) as Sandbian. The Bornival Formation essentially formed by mudstones was correlated by chitinozoans (no index species) with the lowest Katian (Vanmeirhaeghe 2006). The Cimetière de Grand–Manil Formation was not biostratigraphically dated. The Huet Formation is formed by greenish slates and siltstones with shelly facies level (e.g. brachiopods, bryozoans, corals and crinoids; Mailleux 1926) interpreted as distal tempestites. It was dated by chitinozoans: *T. bergstroemi* (Van Grootel *et al.*

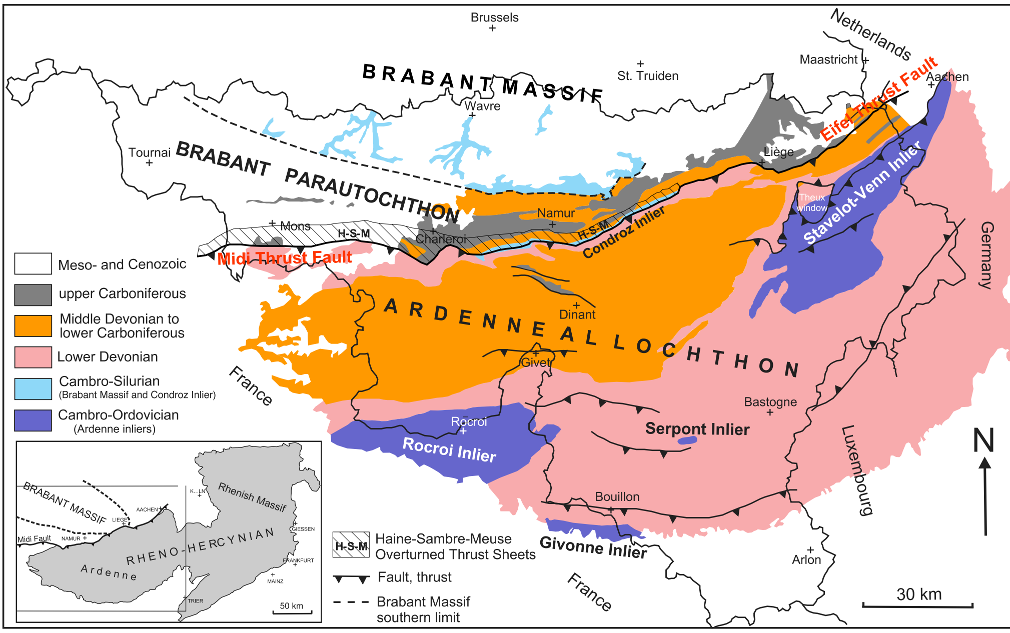

Fig. 2. Simplified geological map of the southern part of Belgium and neighbouring countries showing the main tectonostratigraphic units of the Paleozoic. To the north, the Brabant Parautochthon and/or the Haine–Sambre–Meuse Overturned Thrust sheets are separated from the Ardenne Allochthon by the Midi-Aachen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

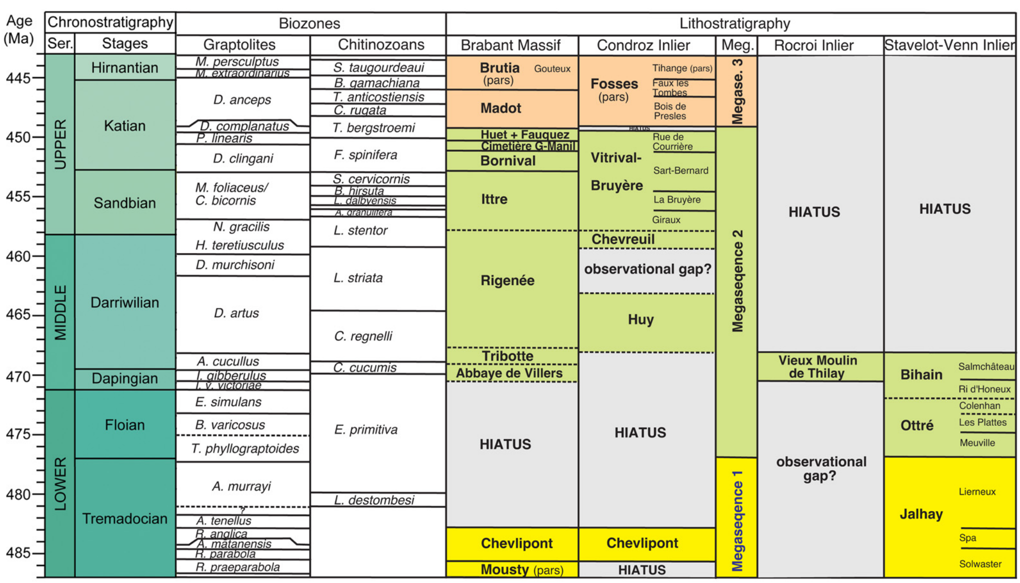

Fig. 3. Ordovician basement stratigraphy of Belgium and neighbouring countries. Chronostratigraphy, graptolite and chitinozoan biozones after Goldman *et al.* (2020, fig. 20.2–4, Baltica column for chitinozoan and Britain for graptolite zonations). References are provided in the text for the bio- and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

1997) and *F. spinifera* biozones (Vanmeirhaeghe 2006) and correlated with the middle Katian (Herbosch and Verniers 2014, fig. 3). The Fauquez Formation includes graptolitic black slates interpreted as very distal turbidites deposited in a deep marine environment. It was correlated by graptolites (*Pleurograptus linearis*; Maletz and Servais 1998) and chitinozoans (*Fungochitina spinifera*; Vanmeirhaeghe 2006) with the middle Katian.

Megasequence 3 begins with the Madot Formation showing many volcanic and volcano-sedimentary rocks interstratified with siltstones rich in shelly debris (Verniers *et al.* 2005) deposited in a shallow shelf. The transition from the Fauquez to the Madot Formation marks an abrupt and important change in bathymetry. The shallowing could be linked to the docking of Avalonia with Baltica (Cocks and Torsvik 2005; Torsvik and Cocks 2011; Linnemann *et al.* 2012) and was correlated by the chitinozoans *Tanuchitina bergstroemi* (Vanmeirhaeghe *et al.* 2005) and *Conochitina rugata* (Mortier 2014) with the upper half of the Katian (Fig. 3). The lower Gouteux Member of the Brutia Formation includes grey bioturbated slates and was assigned by the chitinozoans *Belonechitina gamachiana* (Samuelsson and Verniers 2000) and *Spinachitina oulebsiri* (Mortier 2014) to the uppermost Katian–Hirnantian. It extends most probably even into the Silurian.

The Condroz Inlier

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

Megasequence 1 is reduced to the Chevlipont Formation observed in the Wépion borehole (Graulich 1961). Megasequence 2 begins with a 10 cm thick basal conglomerate, followed by the upper slope dark mudstones of the Huy Formation correlated by chitinozoans (*S. formosa* Biozone) with the middle Darriwilian (Vanmeirhaeghe 2006). The unconformity is well constrained, with a hiatus from the middle Tremadocian to the lower Darriwilian (Vanmeirhaeghe 2006; Wang and Servais 2015), i.e. a time interval close to that observed in the BM (Fig. 3). After an observational gap, the sedimentation continued with the Chevreuil and the Vitrival–Bruyère formations deposited on the outer shelf to upper slope. The Chevreuil Formation was dated by chitinozoans (typical association, no index species) as late Darriwilian to early Sandbian (Vanmeirhaeghe 2006). The Vitrival–Bruyère Formation is subdivided into four members: the Giraux Member, dated by graptolites (*N. gracilis* Biozone) and chitinozoans (*L. stentor* Biozone) as lower Sandbian; the poorly dated La Bruyère Member; and the Sart-Bernard and the upper part of the Rue de Courrière members, both correlated by chitinozoans (*S. cervicornis* Biozone) with the lower Katian (Vanmeirhaeghe 2006, fig. 47). The uppermost Rue de Courrière Member is composed of siltstones with shell beds, sandstones and microconglomerates, which mark the base of a short stratigraphic hiatus spanning the Onnian (middle Katian), and points to the occurrence of a paraconformity. The repercussions of the drastic change in bathymetry observed in the BM at the end of Megasequence 2 (see above) lead in the central CI to the emergence, as this region is globally shallower than the BM (Vanmeirhaeghe 2006, p. 202; Herbosch and Verniers 2014, fig. 11).

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

The Stavelot–Venn Inlier

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

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

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

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

The Rocroi Inlier

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

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

Givonne and Serpont inliers

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

The Ordovician of the Rhenish Massif and Rügen (by Lefebvre and Servais)

General framework

In western Germany, Ordovician strata are exposed in the eastern part of the Stavelot–Venn inlier (see above) and also in the two Ebbe and Solingen–Remscheid–Altena inliers, in the Rhenish Massif (Fig. 1; Servais *et al.* 2008, and references therein). Thick, tectonically complex Ordovician successions are also known from boreholes in Rügen (Fig. 1), in northeastern Germany, and Pomerania, in northwestern Poland (Trela 2022). All of these regions display relatively comparable successions of Ordovician rocks, yielding similar palynomorph assemblages and low-diversity deep-shelf faunas dominated by graptolites and phyllocarids. Moreover, strong lithological, faunistic and geochemical similarities between Ordovician successions in Rügen, the Rhenish Massif, the SVI, the Brabant Massif, the English Lake District and the Welsh Basin strongly support the view that all these areas were part of the same Avalonia microcontinent (Maletz and Servais 1993; Servais *et al.* 1998, 2008; Pharaoh 1999; Samuelsson *et al.* 2002; Verniers *et al.* 2002). In the last 30 years, there has been significant progress on the bio- and chronostratigraphy of the Ordovician of the Rhenish Massif and Rügen through a series of detailed studies based on acritarchs, chitinozoans and graptolites.

Rhenish Massif

In western Germany, the most extensive and fossiliferous succession of Ordovician sedimentary rocks is exposed in the Ebbe and Solingen–Remscheid–Altena inliers, both situated near Cologne (Sauerland), in the northeastern part of the Rhenish Massif (*Rheinisches Schiefergebirge*; Fig. 1). In this region, the *c.* 800 m thick Herscheid Beds (*Herscheider*

Schichten) consist of a monotonous, siliciclastic succession (mudstones and siltstones), classically subdivided into four units (Maletz 2000; Eiserhardt *et al.* 2001; Servais *et al.* 2008; Koch *et al.* 2014).

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

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

The Rahlenberg Grauwackenschiefer Formation is a 300 m thick succession of thin-bedded dark (blue to grey-black) silty mudstones with sandy layers that yielded siliceous concretions in its upper

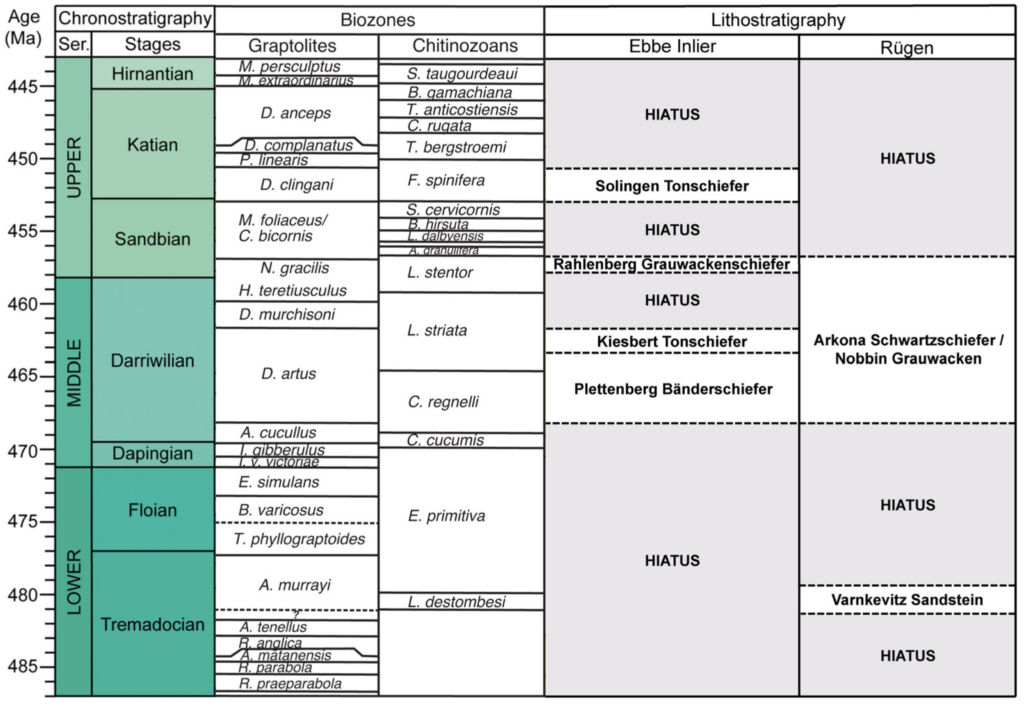

Fig. 4. Ordovician stratigraphy of the Ebbe Inlier (Rhenish Massif, western Germany) and the subsurface of Rügen Island (northeastern Germany). Chronostratigraphy, graptolite and chitinozoan biozones after Goldman *et al.* (2020, fig. 20.2–4, Baltica column for chitinozoan and Britain for graptolite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

part. The fauna is scarce and comprises conulariids, graptolites, phyllocarids and trilobites (Eisenack 1939; Servais *et al.* 2008). Palynomorphs indicate an early Sandbian age (Fig. 4; Eisenack 1939; Samuelsson *et al.* 2002; Servais *et al.* 2008; Koch *et al.* 2014), in good accordance with the presence of the graptolite *Pseudoclimacograptus* sp. (Maletz 2000).

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

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

Rügen

In northeastern Germany, a particularly thick Ordovician sedimentary succession (over 3000 m) was described from numerous boreholes drilled in Rügen (Fig. 1) during the 1960s (Jaeger 1967; Servais 1994; Maletz 1998, 2001; Beier *et al.* 2001; Verniers *et al.* 2002; Servais *et al.* 2008). This mainly pelitic and clastic succession is synsedimentary deformed, locally overturned (Franke and Illers 1994; Beier and Katzung 1999; Samuelsson *et al.* 2000), suggesting unstable environmental conditions (slumps) along an active continental margin and/or tectonic stacking (Servais and Katzung 1993; Giese *et al.* 1994; McCann 1998). With the exception of graptolites and palynomorphs, only trace fossils (e.g. *Nereites*) and phyllocarid crustaceans were reported from the boreholes (Zagora 1997; Maletz 2001). A subdivision of the Ordovician succession into three tectonically limited lithostratigraphic units was established, based on the thickest and most complete drilled successions (e.g. the Rügen 5 borehole; Giese *et al.* 1994; Beier *et al.* 2001; Servais *et al.* 2008).

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

The overlying Arkona Schwarzschiefer Formation is a very thick succession of black shales (over 1000 m) observed in most boreholes (Samuelsson *et al.* 2000; Maletz 2001). Both its lower and upper parts are tectonically disturbed, and no borehole yielded a complete, continuous succession (many intervals are missing, probably due to tectonics). These levels have yielded graptolite assemblages characteristic of the middle and upper Darriwilian (equivalents of the *H. lentus*, *P. elegans*, *P. distichus*, and ?*H. teretiusculus* biozones) and the lower Sandbian (*N. gracilis* Biozone) (Fig. 4; Jaeger 1967; Maletz 1998, 2001). Palynomorphs provided a comparable range of ages for the Arkona Schwarzschiefer Formation (Burmann 1968, 1970; Samuelsson *et al.* 2000; Samuelsson and Servais 2001; Vecoli and Samuelsson 2001; Servais *et al.* 2008).

The Nobbin Grauwacken Formation is a very thick (*c.* 1800 m) succession dominated by greywackes and shales, which was only observed in some boreholes. Graptolite and palynomorph assemblages yielded congruent ages (middle Darriwilian–early Sandbian) for this unit, similar to those obtained for the Arkona Schwarzschiefer Formation (Fig. 4; Maletz 1998, 2001; Samuelsson *et al.* 2000; Servais *et al.* 2001, 2008). This implies that both formations were probably concurrently deposited in distinct environmental conditions; the black shales of the Arkona Schwarzschiefer Formation were probably related to more distal (bathyal?) settings than the greywackes of the Nobbin Grauwacken Formation (Zagora 1997; Samuelsson *et al.* 2000).

In summary, the subsurface of the island of Rügen yielded a very thick, although tectonically complex, Ordovician succession comprising Tremadocian (Varnkevitz Sandstein Formation) and middle Darriwilian to lower Sandbian (Arkona Schwarzschiefer and Nobbin Grauwacken formations) sedimentary rocks (Fig. 4). The Ordovician successions of Rügen, the Rhenish Massif (see above) and Pomerania (NW Poland; see Trela 2022) share many similarities in terms of lithologies, faunas, stratigraphy (e.g. latest Tremadocian–early Darriwilian gap), and associated environmental conditions (both were originally deposited in deep-water settings), thus supporting the view that they were parts of a same palaeogeographic area (Jaeger 1967), in eastern Avalonia (Maletz and Servais 1993; Servais *et al.* 1998; Pharaoh 1999; Maletz 2001; Samuelsson *et al.* 2002; Verniers *et al.* 2002).

The Ordovician of the Armorican Massif (by Vidal, Loi, Lefebvre and Ghienne)

General framework

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

The NAD (i.e. Cotentin, zone bocaine and Domfront area) and the MAD (i.e. Châteaulin, Menez–Belair and Laval synclines) are separated by the North Armorican Shear Zone. Their Paleozoic successions are quite similar from the Middle Ordovician onwards (Fig. 6). Based on their tectonic history and faunistic affinities, the NAD and MAD are often united within a single Median-North Armorican Domain (MNAD; Robardet *et al.* 1994*a*, *b*; Paris 2016). However, recent U–Pb analyses on detrital zircons coupled with Sm–Nd and Lu–Hf isotope analyses identified different zircon populations in the Lower Paleozoic successions of these two areas (Dabard *et al.* 2021). In the Ordovician sediments of the MAD, the occurrence of Stenian detrital zircons suggests affinities with the eastern African 'Sahara metacraton' and the Arabian Nubian Shield. In contrast, the absence of Stenian and late Tonian ages in zircon populations in Ordovician sediments from the NAD implies distinct source areas,

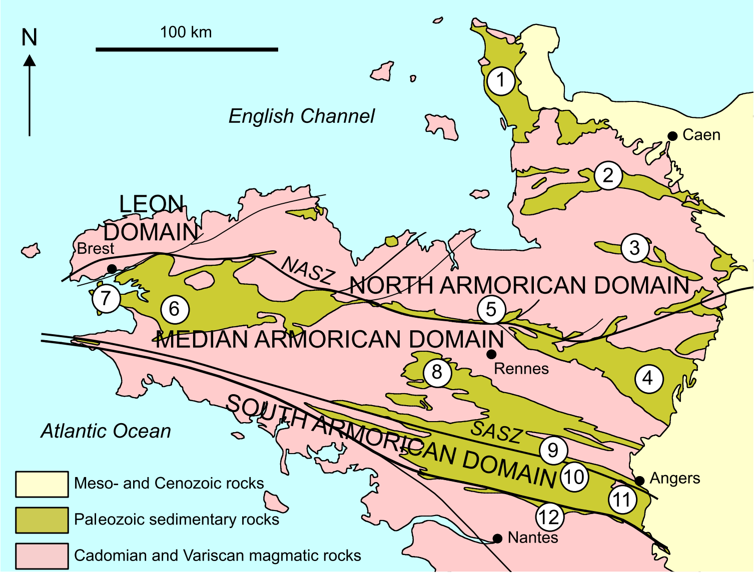

Fig. 5. Simplified geological map of the Armorican Massif (western France) illustrating its division into four distinct domains separated by two main late Carboniferous (Variscan) shear zones: the North Armorican Shear Zone (NASZ) and the South Armorican Shear Zone (SASZ). All main areas yielding Ordovician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

probably located more westward. These features suggest that the NAD and the MAD were geographically separated during the Early Ordovician, and moving closer only during the Variscan orogeny (Fig. 1).

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

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

Lower Ordovician

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

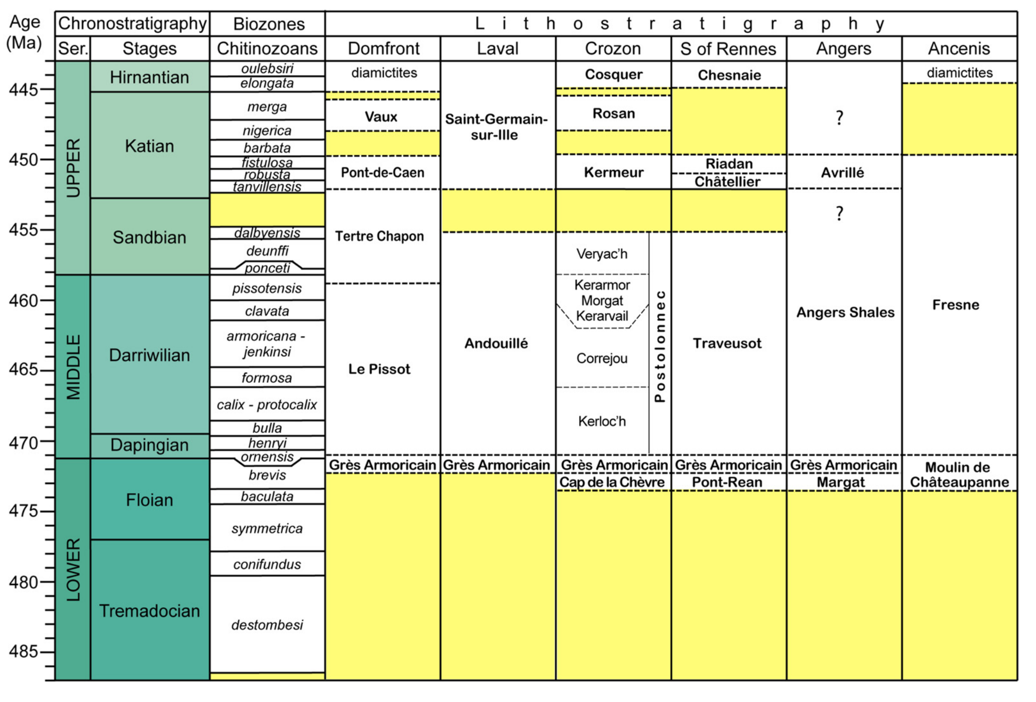

Fig. 6. Ordovician stratigraphy of the Armorican Massif (western France). Chronostratigraphy and high latitude Gondwanan chitinozoan biozones after Goldman *et al.* (2020). References are provided in the text for the bio- and chronostratigraphic position of the formations belonging to the North Armorican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

of the Brioverian succession (Gougeon *et al.* 2018, 2022; Néraudeau *et al.* 2018).

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

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

The thickness of the Grès Armoricain Formation varies from tens to several hundreds of metres. Depending on the regions, it consists of either an undivided mass of sandstones with silty–clayey intercalations (e.g. Domfront in Normandy, northern flank of Menez–Bélair), or is heterogeneous and subdivided into three members: (1) a lower member including thick sandstone beds only separated by few silty to clayey intercalations – this member can comprise iron layers, which used to be exploited in the synclines south of Rennes (Chauvel 1968); (2) a middle member (variously named ‘Congrier Member’, ‘Gador Member’ or *Schistes intermédiaires* depending on the region), consisting of siltstones, argillites with micaceous sandstone intercalations; and (3) an upper member, which is mainly composed of sandstones with silty–clayey intercalations (e.g. in the Châteaulin synclinorium, synclines south of Rennes), and containing sometimes highly radioactive placers (rutile and zircon grains; Faure 1978; Pistis *et al.* 2008, 2016).

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

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

Middle Ordovician–Sandbian

During the Dapingian–Sandbian interval, the sedimentation was dominated by the deposition of dark, fine-grained shales, previously called ‘Schistes à *Calymene*’, and currently named the Postolonnec, Andouillé, Le Pissot and Traveusot formations in the MNAD, Schistes d’Angers and Fresne formations in the SAD (Fig. 6). These 150–400 m thick silty–clayey units bear levels with siliceous, phosphatic and/or calcareous concretions, shell-beds and sandstone intercalations that can be several dozen of metres thick (e.g. Kerarvail Member, about 30 m thick in the Crozon Peninsula). The main sedimentary structures are planar bedding, hummocky cross-stratifications and numerous

A- or B-type shell beds (type A, thin, lenticular single coquinas; type B, thick, continuous and composite shell concentrations, *sensu* Botquelen *et al.* 2004). All of these structures are indicative of a storm wave-influenced sedimentation. These deposits are related to various environments ranging from beach (shoreface) to lower offshore (Guillocheau 1983; Loi *et al.* 1999; Loi and Dabard 2002; Dabard *et al.* 2007, 2015; Guillocheau *et al.* 2009). Phosphatic and siliceous concretions and crusts are common in Middle Ordovician sedimentary rocks. They are of early diagenetic origin and are promoted by the stability of the water/sediment interface. These concretions, which are indicative of sedimentary condensation, were mainly produced during sea-level rises (Loi *et al.* 1999; Loi and Dabard 2002; Dabard *et al.* 2007; Dabard and Loi 2012). Depending on the depositional environment, they can be linked to composite shell beds (B-type shell beds *sensu* Botquelen *et al.* 2004).

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

Differences in the composition of Darriwilian–Sandbian benthic assemblages in the North, Median- and South Armorican Domains mainly result from distinct environmental settings, with a southward (in present-day geography) deepening trend (Henry 1989; Henry *et al.* 1997; Lefebvre 2007; Courville and Gendry 2016). In the NAD, benthic diversity is low and trilobite taxa (Homalonotinae) are typically related to shallow-water environments (Henry 1989 and references therein). In the MAD, the faunal diversity is significantly higher and characteristic of median shelf settings with both typical endobenthic and epibenthic trilobite taxa (Henry 1989; Courville and Gendry 2016). In the SAD, fossils are rare and outcrops discontinuous. However, assemblages are typical of distal shelf environments, with strong affinities with coeval Bohemian faunas from the Prague Basin (Henry *et al.* 1997; Lefebvre *et al.* 2010). Assemblages correspond to the 'atheloptic biofacies' with benthic trilobites either blind or with atrophied eyes, associated with pelagic forms with hypertrophied eyes (Cyclopygidae), along with lagynocystid mitrate stylophorans (Henry *et al.* 1997; Lefebvre 2007; Lardeux *et al.* 2008; Jouhier and Gendry 2017).

Several transgressive and regressive cycles were identified in the Middle Ordovician–Sandbian succession of the MAD, based on the detailed analysis of sedimentary facies (Dabard *et al.* 2007, 2009; Paris *et al.* 2007). Transgressive episodes identified within the *D. bulla*, *S. formosa* and *L. pissotensis* chitinozoan biozones were recognized in other peri-Gondwanan areas (Dabard *et al.* 2007; Paris *et al.* 2007; Videt *et al.* 2010). In the Crozon Peninsula, two key sections within the Postolonnec Formation provided a continuous, well-exposed framework for the detailed analysis of sedimentary facies and gamma-ray record (Dabard *et al.* 2015). There, a back-stripping procedure was applied and calibrated with graptolite biozones to convert the observed variations in depositional environments into apparent sea-level changes, and ultimately into eustatic variations. The resulting pattern (Fig. 7) supports the occurrence of short-term and high-amplitude cyclic sea-level fluctuations, suggesting glacio-eustasy. This pattern suggests the occurrence of ice-house conditions and oscillations of an ice-sheet on Gondwana at least at the lower limit of the *L. pissotensis* Biozone (at the transition between the Kerarvail and Morgat members), but probably as early as the *S. formosa* chitinozoan Biozone during the Darriwilian (i.e. at the transition between the Kerloc'h and Corréjou members, Fig. 6; Dabard *et al.* 2015). The expression of these ice sheet-induced oscillations differs in other parts of the Armorican Massif, depending on their initial location along a proximo-distal transect and/or erosional surfaces. For example, no sandy member (equivalent to Kerarvail in

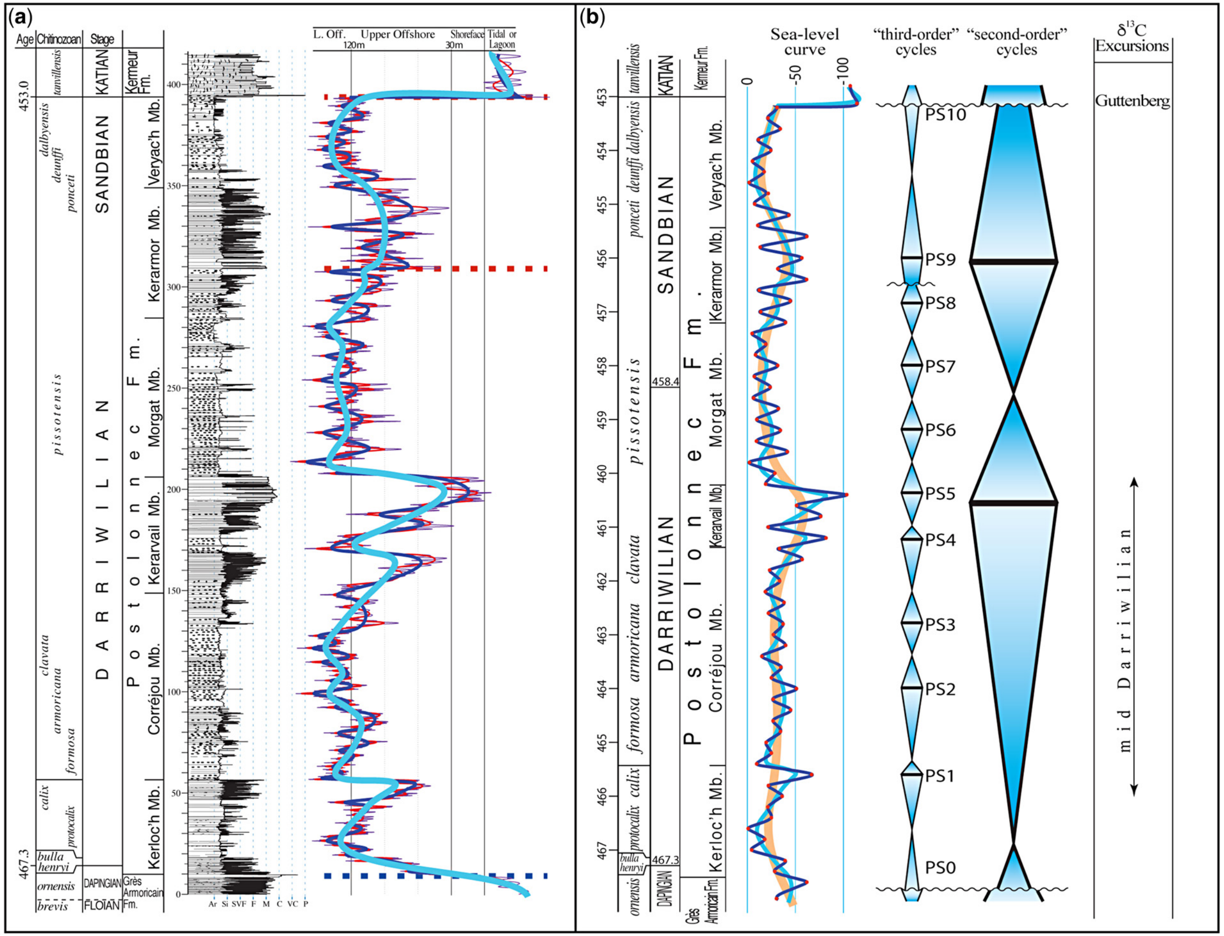

Fig. 7. Sequence stratigraphy of the Postolonnec Formation, Crozon, western part of the MAD (from Dabard *et al.* 2015). (**a**) Lithostratigraphic column with depositional environments deduced from sedimentological features and gamma-ray data. Very high frequency (purple and red curves), high frequency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

Crozon) is observed in the central and eastern parts of the MAD, characterized by more distal palaeoenvironmental conditions. In Normandy, the coarser, shallow-water facies of the Middle Ordovician successions and the quality of the outcrops do not make it possible to achieve any coherent, detailed analysis. Nevertheless, all of these different sections may be correlated based on their chitinozoan assemblages (Fig. 6; Paris 2016).

Katian

In many regions of the Mediterranean Province, the Sandbian–Katian transition coincides with a major regressive episode (Robardet 1981; Villas 1992; Bourahrouh *et al.* 2004). In both Brittany and Portugal, the recovery of the sedimentation process is expressed by a conglomeratic layer bearing phosphatic pebbles, ferruginous and phosphatic oolites (Paris 1979, 1981; Young 1988). Palynological data suggest a stratigraphic gap, at least locally: chitinozoans extracted from the conglomerate matrix belong to the *E. tanvillensis* Biozone (Katian), whilst those from the pebbles belong to the *L. deunffi* or *L. dalbyensis* biozones (Sandbian; Paris 1981; Bourahrouh 2002; Vidal *et al.* 2011*b*; Dabard *et al.* 2015). This level represents the base of a new transgressive sequence corresponding to the Kermeur and Saint-Germain-sur-Ille formations in the median synclinorium, and to the Grès du Châtellier followed by the Schistes de Riadan south of Rennes (Dabard *et al.* 2009; Paris 2016). These sandstone-dominated deposits contain silty and clayey intercalations, as well as phosphatic, siliceous and/or calcareous concretions. The sedimentary structures (hummocky cross-stratifications, symmetric ripples (= oscillation ripples), clay drapes, inversion in palaeocurrent directions, evidence of emersion) reflect shallow environments subjected to tidal influence and to fair-weather and storm waves. The depositional environments inferred from the sedimentary facies range from barrier and back-barrier nearshore settings (e.g. base of the Kermeur Formation; Vidal *et al.* 2011*b*) to the upper offshore (Gorini *et al.* 2008) throughout several transgression–regression cycles (Gorini *et al.* 2008; Dabard *et al.* 2009). Some levels are extensively bioturbated by well-diversified trace fossils (Mélou and Plusquellec 1975), including Equilibrichnia (Vidal *et al.* 2011*b*) and spiral-shaped traces. Benthic assemblages are scattered in the arenaceous base of the Kermeur Formation, and in the siltstones of the upper part of this formation, equivalent to the Châtellier and Riadan formations. The main taxa are trilobites (see Henry 1980; Vidal *et al.* 2011*b*), diploporite echinoderms, brachiopods and rare bivalves (Babin and Mélou 1972; Mélou 1985; Botquelen and Mélou 2007) and gastropods. These faunas are less diverse than those occurring in underlying formations. Accurate stratigraphic assignments of the Kermeur, Saint-Germain-sur-Ille, Châtellier and Riadan formations are provided by chitinozoans, indicative of the Katian (from the *E. tanvillensis* to the *A. barbata* biozones; Paris 1981, 2016; Bourahrouh 2002; Gorini *et al.* 2008). No coeval section is exposed in the SAD, but concretions yielded a typical Katian trilobite fauna with *Deanaspis*, *Dreyfussina*, *Eudolatites* and *Prionocheilus* (Beaulieu *et al.* 2014).

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

Hirnantian

In the Mediterranean Province, the Hirnantian Stage is characterized by a well-recorded glacial episode in most regions (e.g. Algeria, Morocco, Portugal, Spain; Ghienne *et al.* 2007; Loi *et al.* 2010; Videt *et al.* 2010). In the Armorican Massif Hirnantian deposits constitute the *Pélites à fragments* (diamictites) in Normandy, the uppermost part of the Saint-Germain-sur-Ille Formation in Menez-Bélair and Laval synclines, the Chesnaie Formation in the South of Rennes or the Cosquer Formation in western Brittany. These facies are interpreted as glaciomarine deposits composed of fragments originating from melting ice (Dangeard and Doré 1971; Hamoumi 1981; Hamoumi *et al.* 1981; Robardet and Doré 1988).

In these levels, the macrofauna is scarce. In the Crozon Peninsula, a single locality (Trégarvan section) yielded a fossiliferous sandstone level with *Hirnantia sagittifera* (Mélou 1990), elsewhere named ‘Grès de Roudou Hir’ or ‘Grès de Lamm Saoz’ at the top of the Cosquer Formation (Plusquellec *et al.* 1999). This ubiquitous brachiopod is emblematic of the Hirnantian Stage. This age is also supported by palynological data (chitinozoans), which are typical of the *T. elongata* Biozone (Bourahrouh 2002; Paris 2016).

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

General framework

After the late Cambrian rifting along the so-called Atlas–Ossa-Morena–North Armorican rift (western Gondwana; Álvaro *et al.* 2021), drifting conditions prevailed during the Ordovician. However, the evolution of the high-latitude western Gondwanan margin was not a monotonous journey, but an eventful drift episodically punctuated by major geodynamics events. One of them was initially identified as a Middle Ordovician gap in southwestern Sardinia: the Sardic Phase (see Loi *et al.* 2022) has during the last few decades become a standard for correlation in Western Europe, in spite of sometimes different conceptions both regionally and academically. Similar gaps involving the absence of Middle Ordovician strata have been invoked to recognize Sardic-style geodynamic events in the eastern Pyrenees, southern France and the Alps (Stampfli *et al.* 2002). There, the Sardic phase is characterized by generalized cortical uplift, denudation of exposed uplifted areas under subaerial conditions, Middle Ordovician stratigraphic gaps of about 25–30 myr, broad felsic plutonism and volcanism (with felsic rocks now orthogneisses after Variscan deformation and metamorphism) with calc-alkaline affinity, and the record of alluvial to fluvial deposits onlapping an inherited palaeorelief (Álvaro *et al.* 2018; Casas and Álvaro 2019). However, the record of Ordovician volcanism associated with local listric faults in neighbouring areas cannot be used as a supporting argument for contemporaneous Sardic events (Álvaro *et al.* 2020).

Tectonostratigraphic framework

The southern Massif Central or Occitan Domain (Pouclet *et al.* 2017) comprises the Variscan tectonostratigraphic units of the Thiviers–Payzac and Génis, Rouergue–Albigeois, Montagne Noire, southern Cévennes and Mouthoumet massifs (Álvaro *et al.* 2021) (Fig. 8). Two massifs are selected for comparison here: the Montagne Noire and the Mouthoumet massifs. The former has been traditionally subdivided into three tectonostratigraphic units, which are: (1) the northern Montagne Noire with a southward tectonic vergence, comprising Lower Paleozoic low-grade metasedimentary units; (2) the Axial Zone, an elongated dome of migmatized orthogneisses displaying an Ordovician age for their protoliths (460–450 Ma; Roger *et al.* 2004), which also includes migmatites and micaschists derived from Neoproterozoic–Ordovician rocks; and (3) the southern Montagne Noire made up of south-facing nappes including a complete and fossiliferous Cambrian–Ordovician succession (Vizcaïno *et al.* 2001) (Fig. 8). The emplacement of recumbent folds during the Visean–early Namurian altered the pre–Variscan (palaeogeographic) proximal-to-distal trend, placing the (proximal) Axial Zone in an intermediate position between the southern and (distal) northern Montagne Noire flanks. The Mouthoumet Massif or inlier, located between the Montagne Noire and the North Pyrenean frontal thrust, represents the southernmost prolongation of the southern Massif Central (Berger *et al.* 1997).

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

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

A comprehensive summary of the Lower Ordovician stratigraphic nomenclatures from the southern Montagne Noire was presented by Vizcaïno *et al.* (2001), who subdivided the succession into several formations, alternating between shale-dominant (La Gardie, Saint-Chinian, Setso and Landeyran units) and sandstone-dominant (La Dentelle, La Maurerie, Cluse de l'Orb and Foulon) intervals, locally punctuated by carbonate interbeds (Val d'Homs and Mounio formations). A regional biozonation (Fig. 9) of the Tremadocian–Floian interval was proposed by Vizcaïno and Álvaro (2003), Tortello *et al.* (2006) and Álvaro *et al.* (2007, 2008). This biostratigraphic scheme relies on correlations with Avalonia (England), Baltica (Scandinavia) and high-latitude (peri-)Gondwana areas (e.g. Bohemian Massif and Sardinia). It comprises, from bottom to top the *P. geinitzi* (mid Tremadocian), *S.* (*C.*) *pusilla* (mid–late Tremadocian), *E. filacovi* (late Tremadocian), *T. miqueli* (latest Tremadocian–early Floian), *T. shui landayranensis* (mid Floian), *C. maynardensis*

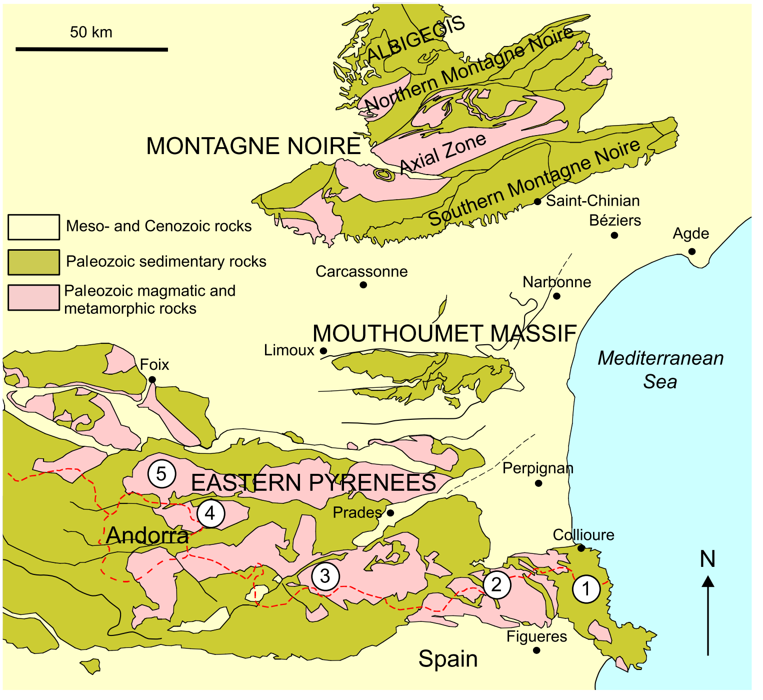

Fig. 8. Simplified geological map of the Occitan Domain (Albigeois, Montagne Noire, Mouthoumet Massif), and eastern Pyrenees in southern France and neighbouring countries, with: (1) Albera (Albères) Massif; (2) Roc de Frausa (Roc de France) Massif; (3) Canigó (Canigou) Massif; (4) Hospitalet Massif;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

(mid Floian), *N.* (*N.*) *arenosus* (mid Floian), *A. incisus* (late Floian) and *H. primitivus* biozones (late Floian) (Fig. 9). The base of each biozone corresponds to the first appearance datum of its index taxon (Vizcaïno and Álvaro 2003). This biozonation was completed by the presence of conodonts in the limestone interbeds belonging to the Tremadocian *P. deltifer* Biozone (Álvaro *et al.* 2007; Serpagli *et al.* 2007).

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

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

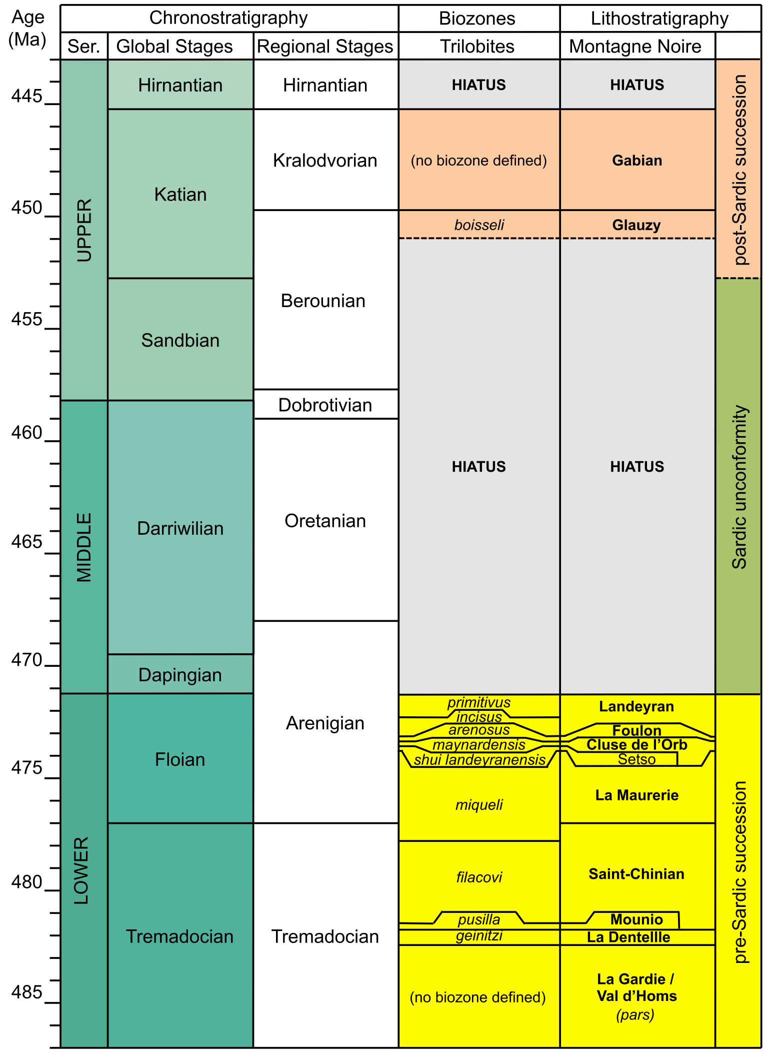

Fig. 9. Ordovician stratigraphy of the Montagne Noire (southern France). Chronostratigraphy after Goldman *et al.* (2020), and trilobite biostratigraphy based on Vizcaïno and Álvaro (2003), Tortello *et al.* (2006), Serpagli *et al.* (2007) and Colmenar *et al.* (2013). References are provided in the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

trilobites (e.g. Dean 1966; Capéra *et al.* 1978; Courtessole *et al.* 1983; Vizcaïno *et al.* 2001; Tortello *et al.* 2006; Van Iten and Lefebvre 2020).

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

Massif. The second Early Ordovician magmatic event is illustrated by the Peyrebrune and Davejean Volcanic Complex basalts, which display a typical initial rift tholeiite magmatic signature indicating contributions of both asthenospheric and lithospheric mantle sources (Pouclet *et al.* 2017).

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

The Sardic Phase

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

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

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

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

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

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

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

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

General framework

Before the opening of Mediterranean back-arc basins in the Miocene, both Corsica and Sardinia were adjacent to the Maures Massif in Provence, southern France (Corsini and Rolland 2009; Schneider *et al.* 2014; Franke *et al.* 2017; Gerbault *et al.* 2018; Álvaro *et al.* 2021; Loi *et al.* 2022). These three regions belong to the southern Variscan belt of Europe and share a particularly complex geology (Bellot 2005; Gerbault *et al.* 2018).

Maures Massif

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

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

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

The eastern part of the Maures Massif (Cavalières and Petites Maures units) is characterized by high-grade (high pressure) Cambro-Ordovician metagranites and Ordovician metagabbros, suggesting possible affinities to northeastern Sardinia and southwestern Corsica (Bellot 2005; Faure *et al.* 2014; Schneider *et al.* 2014; Franke *et al.* 2017; Gerbault *et al.* 2018; Álvaro *et al.* 2021; Loi *et al.* 2022).

Corsica

In northwestern Corsica, a non-metamorphic Paleozoic succession is preserved in the Argentella area, near Galeria (Baudelot *et al.* 1976; Rossi *et al.* 1995, 2009; Barca *et al.* 1996; Paris *et al.* 1999; Faure *et al.* 2014; Avigad *et al.* 2018). This succession lies unconformably on a more than 1000 m thick basement of Neoproterozoic rocks (amphibolites, metagreywackes, micaschists and quartzites), polydeformed and metamorphosed during the Cadomian orogeny (Barca *et al.* 1996; Faure *et al.* 2014; Avigad *et al.* 2018). The overlying Paleozoic succession comprises several informal stratigraphic units.

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

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

The Monte Martinu Conglomerate is tectonically overlain by the *c.* 150 m thick Campu Orbu Shale.

The dark shales of this unit were originally tentatively assigned to the Silurian, based on poorly preserved acritarchs and chitinozoans (Baudelot *et al.* 1976). However, the occurrence of typical Hirnantian diamictites (20–30 m thick) in the overlying Capu Russellu unit suggests that the age of the Campu Orbu Shale has to be reconsidered and assigned to the Upper Ordovician (Rossi *et al.* 1995; Paris *et al.* 1999; Avigad *et al.* 2018).

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

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

General framework

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

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

Northern complex

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

In the Schwarzwald, the northern complex is less extensive and restricted to the Baden-Baden–Gaggenau area (Eisbacher *et al.* 1989; Servais *et al.* 2008). It comprises a thick succession of low-grade (greenschist facies) rocks generally considered as equivalent to both the Villé and Steige units (Maass 1981; Montenari *et al.* 2000). This succession is usually subdivided into four ‘series’ (Sittig 1965; Montenari *et al.* 2000; Servais *et al.* 2008). The lowermost one (*Untere Schindelklamm Serie*) mainly consists of quartzitic siltstones and quartzites. It is overlain by the diabases (meta-volcanics) of the *Basischer Zug*.

The third part (*Obere Schindelklamme Serie*) comprises greywackes, and it is overlain by the uppermost *Traischbach Serie* formed by a pelitic succession with intercalated carbonate olistoliths. An extensive palynological sampling through all four series yielded a single, poorly preserved acritarch assemblage in the uppermost part of the *Traischbach Serie* (Montenari *et al.* 2000; Montenari and Servais 2000). This assemblage comprises the genera *Acanthodiacrodium*, *Caldariola*, *Cymatiogalea*, *Dasydiacrodium* and *Stelliferidium*, thus suggesting a late Cambrian to Early Ordovician age, in good accordance with that obtained for the Villé unit in the Vosges (Montenari *et al.* 2000; Montenari and Servais 2000; Servais *et al.* 2008).

Central and southern complexes

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

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

In the Badenweiler–Lenzkirch area (Schwarzwald), the southern complex also includes low-grade meta-sediments (greenschist facies; phyllites) structurally below the unmetamorphosed Upper Devonian–lower Carboniferous succession. The age of the phyllites from the Badenweiler–Lenzkirch area long remained poorly known, and they were tentatively assigned to the Devonian (Schäfer 1957; Metz and Rein 1958) or to the lower Carboniferous (Maass 1961; Krohe and Eisbacher 1988). The occurrence of putative remains of Ordovician palynomorphs in this unit was first suggested based on poorly preserved material (Hann *et al.* 1995; Sawatzki *et al.* 1997). However, the description of better-preserved palynomorphs finally confirmed the Early Paleozoic age of the phyllites (Montenari and Maass 1996; Montenari *et al.* 2000; Vaida *et al.* 2004; Servais *et al.* 2008). At least two distinct fossiliferous lithostratigraphic units were documented in the Badenweiler–Lenzkirch area. One consists of meta-greywackes and meta-pelites yielding a Lower–Middle Ordovician acritarch assemblage, including the eponymous taxa of the *E. brevis* (upper Floian) and ?*C. protocalix* (Darriwilian) biozones (Vaida *et al.* 2004). The second level corresponds to meta-siltstones containing a relatively diverse palynomorph assemblage (acritarchs, chitinozoans and cryptospores) suggesting a Silurian (Wenlock–Ludlow) age (Montenari and Maass 1996; Montenari *et al.* 2000; Vaida *et al.* 2004).

Concluding remarks and perspectives

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

In the Ordovician, the regions situated in present-day Belgium and neighbouring areas of France and Germany (Brabant Massif, Condroz and Ardenne inliers), western Germany (Rhenish Massif) and in the subsurface of northeastern Germany (Rügen) were all part of the Avalonia microcontinent. In most of these areas, the Ordovician succession is characterized by a *c.* 13 myr hiatus extending from the mid Tremadocian and embracing the whole Floian (Fig. 3), which is generally interpreted as resulting from the drift of Avalonia away from Gondwana (Cocks and Torsvik 2002; Verniers *et al.* 2002; Linnemann *et al.* 2012). This interpretation is confirmed by the major and relatively sudden shift in depositional environments recorded in the Stavelot–Venn inlier, at the transition between megasequences 1 and 2 (Fig. 3; Vanguestaine 1992; Verniers *et al.* 2002; Linnemann *et al.* 2012; Herbosch *et al.* 2016, 2020). In this area, instead of a hiatus, the Lower Ordovician succession records a sharp transition from shallow shelf environmental conditions (upper Lierneux Member of the Jalhay Formation, upper Tremadocian) to bathyal settings (Ottré Formation, Floian). In the Brabant Massif deep and often anoxic environmental conditions persisted throughout the Mid Ordovician and part of the Late Ordovician, whereas in the Condroz, Rhenish Massif and Rügen the depositional environment was shallower on the shelf (megasequence 2, Fig. 3; Verniers *et al.* 2002; Servais *et al.* 2008; Linnemann *et al.* 2012). In most areas, the transition between megasequences 2 and 3 around mid-Katian times (Fig. 3) marks a shift from deep and/or outer shelf settings to shallower environments and even emersion in the Condroz, very probably related to the initial stages of the soft docking of eastern Avalonia with Baltica (e.g. Verniers *et al.* 2002; Linnemann *et al.* 2012). The three megasequences documented in Belgium and western and northeastern Germany are comparable with those originally described from the Welsh Basin and the English Lake District

(Woodcock 1990; Molyneux *et al.* 2023), thus supporting close palaeogeographic Avalonian affinities for all these regions (Katzung *et al.* 1995; Maletz 2000; Verniers *et al.* 2002; Servais *et al.* 2008; Linnemann *et al.* 2012).

With only exception of the southern part of the Ardenne inliers, all Ordovician rocks exposed in present-day France and western Germany (Schwarzwald) were originally deposited on the southern margin of the Rheic Ocean, on the Gondwanan passive margin or nearby (Fig. 1; Robardet *et al.* 1994*b*; Paris *et al.* 1999; Servais and Sintubin 2009; Pouclet *et al.* 2017; Álvaro *et al.* 2021; Caroff 2022). However, reconstructing their original depositional setting, relative position and relationship with other (peri-)Gondwanan regions remains difficult. A close palaeogeographic link between the Armorican Massif and the Iberian Peninsula has been long established based on strong lithological and faunistic similarities (e.g. Paris and Robardet 1977, 1994; Robardet and Gutiérrez-Marco 1990; Robardet *et al.* 1994*b*). Probable affinities between these two regions are further supported by the occurrence of a similar Furongian–mid Floian hiatus (Fig. 6). Designated the 'lacune normande' in the Armorican Massif (Le Corre *et al.* 1991) and 'Toledonian gap' in the Iberian Massif (Álvaro *et al.* 2021), this *c.* 22 myr long hiatus possibly resulted, in both areas, from the uplift of Ediacaran–Cambrian rocks (Sánchez-García *et al.* 2019). This 'Norman hiatus' has no equivalent in Corsica, the Occitan Domain (Montagne Noire and Mouthoumet massifs), eastern Pyrenees and the Vosges–Schwarzwald massifs.

In marked contrast with the close proximity of the Armorican Massif and the Iberian Massif, continuous shallow shelf sedimentation is recorded in the Lower Ordovician successions of northwestern Corsica (Avigad *et al.* 2018), the northern and southern Montagne Noire (Fig. 9; Vizcaïno *et al.* 2001) and the Mouthoumet Massif (Colmenar *et al.* 2013), as well as in other areas (e.g. Maures Massif and Western Alps), yielding less well-preserved, metamorphosed rocks (Bellot 2005; Guillot and Ménot 2009). Moreover, the Ordovician successions of Corsica and all regions of the Occitan Domain are characterized by a major hiatus extending from the Dapingian to the early Katian boundary (Fig. 9). Several authors (Pouclet *et al.* 2017; Álvaro *et al.* 2021) proposed to link this stratigraphic gap to the Sardic phase, originally described in the Sulcis–Iglesiente area of southwestern Sardinia (see Loi *et al.* 2022). The occurrence of a similar stratigraphic unconformity (mirroring the Sardic unconformity) in northwestern Corsica, the Occitan Domain, eastern Pyrenees and southwestern Sardinia thus strongly suggests that all these areas were geodynamically linked and palaeogeographically close in the Ordovician. In marked contrast with the situation in these areas, a continuous shelf sedimentation is recorded in the Middle and Upper Ordovician successions of the Armorican Massif (Fig. 6).

It is thus possible to identify two main 'clusters' of Ordovician rocks in France. The first one, which comprises the Armorican Massif, is characterized by strong affinities with the Iberian Massif, a Furongian–mid Floian 'Norman gap' and a continuous shelf sedimentation in the Mid and Late Ordovician (Fig. 6). The second cluster, which includes northwestern Corsica, the Montagne Noire, the Mouthoumet Massif, the eastern Pyrenees and possibly the Maures Massif, can be identified based on strong affinities with Sardinia (Sulcis–Iglesiente unit, Sarrabus and Gerrei units), a continuous shelf sedimentation during the Early Ordovician, and an extensive Dapingian–early Katian Sardic(?) gap (Fig. 9). However, in the late Katian, regions belonging to both clusters are characterized by a similar shift from siliciclastic to carbonate sedimentation, and a major faunal transition from cool-water adapted assemblages to more temperate ones (see e.g. Dégardin *et al.* 1995; Colmenar *et al.* 2013; Lefebvre *et al.* 2022). This major environmental change, which is also recorded in other high-latitude (peri-)Gondwanan regions (e.g. Algeria, Carnic Alps, Iberian Peninsula, eastern Tafilalt area in Morocco and Sardinia), probably results from a short episode of global warming: the 'Boda Event' (Villas *et al.* 2002; Fortey and Cocks 2005; Ferretti *et al.* 2023).

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

Other Western European areas, not treated here, are in a serious need of reevaluation of their Ordovician rocks. One of the most extensive Ordovician successions is probably the particularly thick one occurring in the subsurface of the Aquitaine Basin (Paris and Le Pochat 1994; Robardet *et al.* 1994*b*). Preliminary investigations of boreholes suggests the occurrence of over 1200 m of siliciclastic rocks dated from the Lower and Middle Ordovician, based on acritarchs, chitinozoans, graptolites and trilobites. The presence of Upper Ordovician rocks is less poorly constrained, but typical Hirnantian diamictites are present (Robardet *et al.* 1994*b*). Other poorly known regions in serious need of a reevaluation of their Ordovician rocks include e.g. the central Pyrenees (see e.g. Robardet *et al.* 1994*b*; Dégardin *et al.* 1995), the southeastern-most extremity of the SAD (Chantonnay and Vendée areas; see Pouclet *et al.* 2017) or the meta-sediments of the (French) western Alps (Guillot and Ménot 2009). However, the Montagne Noire is an iconic example illustrating that, even in better known and 'classical' areas, much work still remains to be done. In this region, the trilobite-based biostratigraphic framework (Fig. 9) still needs to be carefully correlated with more 'standard' regional or international biostratigraphic charts based on acritarchs, chitinozoans, conodonts and graptolites.

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