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

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Abstract The non-cylindrical character of the Variscan Belt is established via a description of its most salient geological characteristics and correlations among the main and relatively minor massifs from Bohemia to Morocco on the basis of a bibliographical synthesis. Using the firmly established tectonostratigraphic domains, the continuity of Variscan zones is discussed, as well as the western termination of the one that occupies most of the central part of the belt. This zone is characterized by its allochthony, and breaks the continuity of the zonation, that is, the cylindrism of the belt. A proposal is made to call this zone the Mid-Variscan Allochthon and its wedge shape in map view is explained performing a step by step restoration of late Variscan deformation. Once the geometry of the belt is interpreted by a succession of regionally significant deformation events, a kinematic reconstruction is proposed. It is based on recent developments on the knowledge of Paleozoic plate tectonics and intended to show that the deformation model is compatible with current paleogeographical developments.

Keywords: Variscan Belt, Zone correlation, Mid-Variscan suture, Late Variscan restoration, North Gondwana, Paleozoic

Contents

1. Introduction

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

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

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

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

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

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

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

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

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

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

2. Geological map of the Variscan Belt

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

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

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

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

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

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

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

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

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

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

More details for the reconstruction are given in the following sections. They describe the main outcrops of the Variscan basement in central and western Europe, the Variscan Belt in a strict sense, and in Morocco. The description is intended to highlight the aspects meaningful for correlation among the massifs and the evolution of the belt as a whole. The sections corresponding to the Iberian Peninsula, the French massifs, the Bohemian Massif and Morocco include description of the established tectonostratigraphic zonation and other details pertinent to the reconstruction of the Variscan Belt.

3. Iberian Peninsula

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

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

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

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

3.1. Cantabrian Zone (CZ)

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

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

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

3.2. West Asturian-Leonese Zone (WALZ)

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

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

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

3.3. Central Iberian Zone (CIZ)

Also a slate belt, the CIZ derived form more distal parts of the northern Gondwana platform than the WALZ. Its stratigraphy is similar to that of the CZ and WALZ, although diferences in the Ediacaran succesion, volcanic component and tectonic style allows defining three domains.

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

The terrigenous Ediacaran succession of the CIZ is known as the Schist-Greywacke Complex (Teixeira, 1955) or the Alcudian (Tamain, 1971). It reflects sedimentation in a Cadomian retro-arc basin representing the transition from a Cadomian arc in the periphery of the West African Craton to Cambrian rifting marking the start of peri-Gondwanan terrane dispersion (Linnemann et al., 2008). An Ediacaran glaciation (*c*. 565 Ma) has been identified by glaciomarine deposits in the southern part of the CIZ (Linnemann et al., 2018).

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

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

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

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

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

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

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

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

3.3.2. Obejo-Valsequillo Domain (OVD)

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

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

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

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

3.4. Ossa-Morena Zone (OMZ)

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

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

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

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

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

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

3.4.1. Central Unit (CU)

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

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

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

3.4.2. Évora-Aracena Metamorphic Belt (EAMB)

The southern boundary of the OMZ is interpreted as a suture, partly occuring at the EAMB, a low-P/high-T domain with a high-P ensemble located tectonically above, the Cubito-Moura Unit (Rubio Pascual et al., 2013b). The Évora Massif in Portugal is wide (Fig. 2a), and has been interpreted as a gneiss dome where a ductile detachment with top to SE kinematics separates gneisses and schists, below, from the overlying Cubito-Moura Unit, a phyllonitic mélange including lenses of eclogite and blueschists (Araújo and Ribeiro, 1996; Dias da Silva et al., 2018) which attest for a high-P metamorphic event dated at 370-360 Ma (Moita et al., 2005). The high-P rocks are tectonic imbrications mixed with ophiolitic fragments and emplaced toward the N-NE before early Variscan recumbent folding with west and SW vergence occurred (Araújo and Ribeiro, 1996; Simancas et al., 2003). The fragments include MORB-type, *c*. 480 Ma old metabasites in greenschist and amphibolite facies and ultramafics, interpreted as remnants of the Rheic Ocean (Fonseca et al., 1999; Pedro et al.,

2010). Low-P/high-T granulite-facies metamorphism of the EAMB took place at 343-330 Ma (Dallmeyer et al., 1993; Castro et al., 1999), while magmatism associated to dome development is dated at 343-319 Ma (Dias da Silva et al., 2018). Late upright folding occurred in a left-lateral shearing environment bounded by the Southern Iberian Shear Zone (SISZ) which separates the OMZ from the SPZ (Crespo-Blanc and Orozco, 1988, 1991).

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

3.4.3. Beja-Acebuches Unit (BAU)

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

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

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

3.5. South Portuguese Zone (SPZ)

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

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

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

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

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

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

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

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

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

3.6.1. Parautochthon

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

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

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

3.6.2. Lower Allochthon

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

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

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

3.6.3. Middle Allochthon

This group includes true oceanic units and others interpreted as transitional between the extended continental margin of northern Gondwana and the Rheic oceanic crust. Ages of igneous protoliths are either Cambro-Ordovician or Lower Devonian, which leads to a subdivision in two groups, Cambro-Ordovician ophiolites and oceanic supracrustals, and Lower Devonian ophiolites. When both groups occur together, the older occupies a lower

structural position, just above the Lower Allochthon, implying that it was located closer to the extended margin of Gondwana (Martínez Catalán et al., 1996, 2002).

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

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

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

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

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

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

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

3.6.4. Upper Allochthon

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

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

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

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

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

coeval to magmatic underplating in the ensialic arc (Abati et al., 1999, 2003, 2007; Andonaegui et al., 2002). Rutiles of 390-380 Ma (Abati, 1999) and Ar-Ar ages of amphibolites and metasediments (400-370 Ma; Dallmeyer et al., 1997; Gómez Barreiro et al., 2006) date an early Variscan Barrovian metamorphic overprint associated to recumbent folds and thrusts.

In the Upper Allochthon, changes in metamorphic grade are abrupt and occur along extensional detachments (Díaz García et al., 1999b; Álvarez Valero et al., 2014). Even if their metamorphic histories differ, the lower and upper groups are considered parts of the same terrane because no suture has been found between them. Actually, relics of the Cambro-Ordovician high-T metamorphic event characteristic of the upper group has also been identified in the lower group (Fernández-Suárez et al., 2002).

3.7. Variscan intrusive magmatism in the Iberian Massif

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

3.7.1. Synkinematic granitoids

In NW Iberia, Capdevila (1969) and Capdevila and Floor (1970) identified three groups of Variscan granitoids, which were recognized in other parts of the Iberian Massif (Fig. 2a). The first and oldest group includes per- and metaluminous, biotite-rich synkinematic granodiorites, often with tonalitic facies, representing magmas with poor fluid content. They are allochthonous in relation to their country rocks and derived from partial melting of the lower crust with some contribution from the mantle (Galán et al., 1996). Vaugnerites (durbachites) have been found associated to these rocks in NW Iberia (Gil Ibarguchi, 1981; López-Moro and López-Plaza, 2004; Gallastegui, 2005). The age of this

group was established at 345-320 Ma (Serrano-Pinto et al., 1987; Fernández-Suárez et al., 2000; Gallastegui, 2005; Gloaguen, 2006; Gloaguen et al., 2006; López-Moro et al., 2017).

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

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

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

3.7.2. Postkinematic granitoids

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

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

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

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

In the southern part of the CIZ, two large postkinematic elongated plutons occur parallel to the NW-SE CIZ/OMZ boundary. In the NW, the Nisa-Alburquerque pluton, dated at 310-305 Ma, straddles the Portugal/Spain border (Fig. 2a) and consists of S-type

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

3.7.3. Granitoids of the Rhenohercynian arc

A magmatic complex straddles the OMZ/SPZ boundary. It occurs at the SW of the OMZ, mostly intruded in the EAMB (Fig. 2a), where it is known as the Beja Igneous Complex. In the east of the SPZ, it is known as the Northern Sierra de Sevilla Batholith. The complex is an association of ultramafic to felsic intrusions dated at 355-335 Ma, the same age as the BAU and partially coeval with the Volcano-Sedimentary Complex formed during the uppermost Devonian to Visean extensional episode in the SPZ. The left-lateral Southern Iberian Shear Zone (SISZ) displaced the northern part to the west relative to the southern part, and the BAU is essentially a stripe of the Beja gabbro (342 or 352 Ma; Jesus et al., 2007; Pin et al., 2008) strongly sheared by the SISZ.

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

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

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

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

3.8. Other Variscan massifs of the Iberian Peninsula

3.*8.1. La Demanda and Iberian Chains*

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

3.8.2. Pyrenees and Catalonia Coast Ranges

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

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

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

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

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

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

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

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

4. French and related massifs

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

4.1. Armorican Massif

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

4.1.1. Central Armorican Domain (CAD)

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

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

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

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

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

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

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

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

4.1.2. North Armorican Domain (NAD)

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

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

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

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

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

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

4.1.3. South Armorican Domain (SAD)

Most of the Armorican Massif SW to the Nort-sur-Erdre Fault is occupied by allochthonous units which have been grouped in three ensembles that correlate with equivalents in NW Iberia (Ballèvre et al., 2014 and references therein). The Allochthon crops out in a triangular zone between the Nort-sur-Erdre Fault and the southern branch of the South Armorican Shear Zone, and also as klippen south of the latter (Fig. 4). The original contacts between units are interpreted as thrust faults, although in many cases, including the contact with the Autochthon, they were overprinted by extensional detachments.

The Autochthon consists of Cambrian to Devonian metasediments and metavolcanics in the SE and mostly schists, granitic orthogneisses and migmatites toward the NW. In the south (Vendée littorale), the sequence includes from bottom to top migmatites and high-grade gneisses, a thick succession of Cambro-Ordovician aluminous schists with rare carbonates, Lower Ordovician porphyroids, thin quartzite layers and monotonous grey slates. Above, and separated by a thrust fault, synorogenic deposits, low-grade schists, ampelites and black cherts probably represent the Parautochthon (Ballèvre et al., 2012). Lower Ordovician felsic magmatism in the Autochthon and Parautochthon is interpreted by Pouclet et al. (2017) as related with a continental rift where magmatic underplating produced abundant crustal melting, and which can be followed from central and northern Iberia to southern France.

Above, a thrust sheet with Lower Ordovician porphyroids is still considered the Parautochthon, and occurs beneath the Lower Allochthon and the high-P Middle Allochthon represented by the Bois-de-Cené Unit (Figs. 4 and 5a). Migmatization of the lower parts of the Autochthon postdate emplacement of the allochthonous pile, and was overprinted by the southern branch of the South Armorican Shear Zone.

The Lower Allochthon is well represented in the Champtoceaux Complex, along the Vendée coastline to the south, and in poorly exposed klippen NW of the latter (Figs. 4 and 5a). The better exposures occur at the core of the antiform where the Champtoceaux Complex crops out. It consists of a thick sequence of albite-bearing micaschists derived from late Neoproterozoic greywackes, fine-grained leucocratic gneisses with eclogite lenses, garnet micaschists, and graphitic schists. Leucocratic orthogneisses of calc-alkaline and alkaline composition were dated at *c*. 495-475 Ma (Ballèvre et al., 1987, 1989, 2014). The eclogite protoliths are doleritic dikes intruding the Lower Ordovician granitoids and felsic volcanics (Godard, 1988).

Early Variscan high-P metamorphism is recorded in eclogites, orthogneisses and felsic gneisses. In the Champtoceaux Complex, an along-strike gradient shows increasing metamorphic conditions toward the west (from 1.5-2 GPa to *c*. 2.5 GPa; Godard et al., 1981; Ballèvre and Marchand, 1991; Bosse et al., 2000), indicating continental subduction toward the NW in present coordinates. High-P metamorphism has been dated at *c*. 360 Ma in a glaucophane-bearing eclogite by U-Pb and Sm-Nd methods (Paquette, 1987; Bosse et al., 2000; Paquette et al., 2017). Ar-Ar data on phengite yielded *c*. 350-340 Ma (Bosse et al., 2000; Maurel et al., 2003), probably reflecting decompression, which took place during stacking of the units and thrusting onto the Autochthon. This stage was associated with inverted metamorphism, followed by fast cooling at 340-330 Ma (Pitra et al., 2010).

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

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

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

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

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

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

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

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

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

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

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

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

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

4.2. French Massif Central (FMC)

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

A fold and thrust belt affecting Cambrian to early Carboniferous sediments and igneous rocks developed in the Montagne Noire region, which is considered the continuation of the Pyrenean Axial Zone. It consists of Cambrian to middle Carboniferous low-grade metasediments representing a passive margin. Sediments are mostly terrigenous except for a thick carbonate platform developed during the Middle Devonian to early Carboniferous. In the north, the structure consists of large, south-vergent recumbent folds and thrust faults deformed in the Axial Zone of the Montagne Noire by two large, adjacent flattened gneiss domes whose emplacement is linked with a pull-apart structure created during dextral transcurrence (Franke et al., 2011). The domes consist essentially of Cambro-Ordovician orthogneisses pervasively migmatized. Protoliths have yielded ages between 550 and 450 Ma,

and low-P/ high-T metamorphism is dated at 340-320 Ma (Cocherie et al., 2005; Faure et al., 2010b, 2014). An earlier high-P event is represented by eclogites resulting from the metamorphism of basic dikes. High-P metamorphism has been dated at *c*. 355-350 Ma, and interpreted as the result of intracontinental subduction (Faure et al., 2014). Large recumbent folds developed between 340-330 Ma in the Paleozoic series, and were later deformed by doming and flattening related to transpression at 330-315 Ma. Migmatization in the Axial Zone took place between 320-295 Ma (Poujol et al., 2017) or 315-303 Ma (Trap et al., 2017).

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

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

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

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

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

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

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

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

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

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

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

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

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

4.3. Vosges and Schwarzwald Massifs (VM, SWM)

These two massifs cropping out at the shoulders of the Rhine graben are included in the Moldanubian Zone except for their northern part, belonging to the Saxothuringian Zone (Kossmat, 1927). The following description is synthesized from Skrzypek et al. (2014) and Lardeaux et al. (2014). The Northern Vosges crop out north of the Lalaye-Lubine Shear Zone. It consists of Cambro-Ordovician to Silurian low-grade basement rocks overlain by Middle and Upper Devonian shallow-marine facies and a thick, early Carboniferous flysch. Granitoids are abundant, and include diorites and granodiorites dated at 335-325 Ma and monzogranites and granites of 330-320 Ma associated to ignimbritic volcanism.

The Central Vosges are formed by three high-grade units. The Monotonous Unit is formed by migmatitic paragneisses and some peridotite lenses, and recorded a first metamorphic event of the medium-P and T type. The Varied Unit is also formed by

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

4.5. Variscan intrusive magmatism in France

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

4.5.1. Granitoids of the Armorican Massif

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

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

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

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

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

4.5.2. Granitoids of the French Massif Central

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

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

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

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

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

4.5.3. Granitoids of other French and related massifs

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

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

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

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

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

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

5. Bohemian and related massifs

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

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

5.1. Rhenohercynian Zone (RHZ)

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

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

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

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

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

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

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

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

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

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

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

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

5.3. Saxothuringian Zone (STZ)

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

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

5.3.1. Mid German Crystalline High (MGCH)

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

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

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

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

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

5.3.2. Saxothuringian Autochthon

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

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

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

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

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

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

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

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

The Saxonian Granulitebirge is described by Rötzler and Romer (2010). Its rocks crop out in a SW-NE elongated dome with metamorphism progressively increasing toward its core, which consists of granulite-facies Cambro-Ordovician mainly bimodal igneous protoliths of calc-alkaline and tholeiitic character. These are surrounded by a detachment zone with slivers of dismembered high-grade metapelites and mafic and ultramafic rocks, and by a low-P schist cover of early Paleozoic to Upper Devonian age. The granulites recorded near isothermal decompression from metamorphic peak at 2-2.5 GPa, and post-peak recrystallization at 1.1 GPa. Peak T is dated at 341 Ma, whereas retrogradation in the base of the schist cover is dated at 334 Ma. Average rates of exhumation and cooling of the granulites have been calculated in 7-13 mm/year and 40-70 ºC/Ma. Kinematic criteria indicate top to the SE sense of shear in the detachment zone (Fig. 7a; Kroner, 1995; Franke and Stein, 2000).

The Erzgebirge forms the structurally deepest part of the Saxothuringian basement. It is formed by a complex stack of Cadomian basement and early Paleozoic sedimentary and bimodal volcanics (500-460 Ma) reflecting a passive margin environment, equivalent to that of the Thuringian Facies. The lowermost Gneiss-Amphibolite Unit includes Neoproterozoic greywackes intruded by Cadomian granitoids (570-530 Ma), and affected by medium-P/medium-T amphibolite facies metamorphism. Above, a pile of Paleozoic rocks starts with an ultrahigh-P/high-T Gneiss-Eclogite Unit, and continues upward by a high-P/high-T Gneiss-Eclogite Unit, high-P/low-T Mica schist-Eclogite Unit, medium-P/low-T Garnet Phyllite Unit and low-P/low-T Phyllite Unit (Rötzler and Plessen, 2010). Mylonitic shear zones separate the units and the ensemble from the overlying schist mantle. The Gneiss-Eclogite units also include granulites and ultramafic rocks, and micro diamonds occur in the underlying ultrahigh-P unit, pointing to pressures equivalent to at least 150 km . Závada et al. (2018) shown that the ultrahigh-P granulites and high-P granofelses preserved at the SE flank of the Erzgebirge testify a paleogradient of continental subduction originally dipping to the east, and where the deepest rocks were exhumed along a subduction channel to mid-crustal depths.

The eclogites had been exhumed yet before stacking, and later, the entire stack became overprinted by amphibolite-facies metamorphism, and was exhumed in the dome. The thrust faults responsible from the metamorphic inversion show a tectonic transport to the west or to the N-NNE (Schmädicke, 1994; Franke and Stein, 2000). The upper detachment shows a westward component of motion for the hanging wall (Krohe, 1996, Krawczyk et al., 2000).

According to Franke and Stein (2000) and references therein, ages of high-P and ultrahigh-P metamorphism are somewhat scattered, starting at *c*. 370 Ma and cooling through *c*. 350 Ma until a significant event around 340 Ma. However, specific ages for eclogite-facies metamorphism range 360-335 Ma (Schmädicke et al., 1995). The older ages come from Ar-Ar and Sm-Nd analyses, and are discussed by Rötzler and Plessen (2010), who prefer the age group clustering around 340 Ma as indicating peak of high-P metamorphism during intracontinental subduction. Modern petrochronology studies (Závada et al., 2021) show that the ultrahigh-P conditions for coesite-diamond bearing rocks were reached at 350 Ma, while the granofelses formed at around 340 Ma. The whole system was exhumed at around 338 Ma. A later group of ages around 330 Ma date retrograde penetrative deformation during exhumation.

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

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

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

The Orlica-Śnieżnik Dome represents the more oriental outcrop of the STZ Autochthon and, as the Krkonoše Massif, it is overlain by the Allochthon, here represented by the Moldanubian Zone (Figs. 6 and 7a). The Autchthon is dominated by amphibolite-grade orthogneisses, locally containing lenses of high and ultrahigh-P eclogites and granulites, alternating with bands of biotite- to staurolite-grade supracrustal rocks (Don et al., 1990; Štípská et al., 2004; 2012). The orthogneisses have been dated at 520-490 Ma (Oliver et al., 1993; Turniak et al., 2000; Kröner et al., 2001; Lange et al., 2005). The supracrustal succession consists of a monotonous series of plagioclase-rich paragneisses and a variegated formation consisting mostly of mica schists. A Saxothuringian affinity is revealed by zircon provenance studies (Mazur et al., 2012, 2015). Occurrences of ultrahigh-P rocks provide a record of continental subduction and subsequent melt-assisted extrusion of deeply subducted rocks partly driven by buoyancy forces (Chopin et al., 2012) similar to that of the Erzgebirge. The structure is interpreted as a large gneiss dome created by the expulsion of lower crustal material at the rear of the Sudetic accretionary prism (Chopin et al., 2012; Štípská et al., 2019). The dome comprises dismembered fragments of the Saxothuringian passive margin that were subducted during the Variscan collision and then exhumed in front of the rigid buttress of Brunovistulian block, according to Duretz et al. (2011) and Chopin et al. (2012).

5.3.3. Saxothuringian allochthons

Figure 7b shows a section across the SW part of the Bohemian Massif perpendicular to the western branch of the STZ. It shows the relationships between the Autochthon and the allochthons of this zone, represented here by the Münchberg Klippe, and also with the also allochthonous Teplá-Barrandian and Moldanubian. The Allochthon and Parautochthon in the western branch of the STZ is represented by the Münchberg, Wildenfels and Frankenberg klippen, whose units are separated by thrust faults (Fig. 6). Münchberg is the largest and shows the most complete succession (Klemd, 2010), which is described from bottom to top.

The lower part of the Parautochthon consists of early Carboniferous, mainly Visean proximal synorogenic flysch, often a wildflysch with blocks of probably Cambrian and also Ordovician, Silurian and Devonian blocks and olistoliths (Franke, 1989; Linneman et al., 2003; Klemd, 2010). The upper Parautochthon is represented by the distal and relatively deep Bavarian facies, including Ordovician bimodal volcanism and clastic sediments, as well as Silurian-Devonian volcanics, graphitic schists and cherts.

Above, a unit formed by phyllites and prasinites occurs, followed by another consisting of amphibolite represent the Middle Allochthon. They were formed around 400 Ma (Koglin et al., 2018), and metamorphosed at *c*. 365 Ma (Kreuzer et al., 1989).

The two remaining units represent the Upper Allochthon. The first consists of metasediments, orthogneisses, metagabbros and serpentinites, while the culminating unit is made of banded hornblende gneisses, amphibolites and leucocratic gneisses, and includes eclogites at the base (Klemd, 2010). The igneous protoliths have ages of 525-480 Ma, and the eclogites point to equivalence with the high-P/high-T units, with peak pressure conditions roughly dated in Münchberg at 400 Ma and bearing a record of subsequent amphibolite-facies metamorphism at 380 Ma (Franke, 2000; Koglin et al., 2018).

The structural succession of the Frankenberg klippe consists of a low-grade volcanosedimentary complex that can be ascribed to the STZ autochthon. Above, a prasinite unit would be equivalent to the phyllite and prasinite unit of Münchberg, and represent an upper crustal piece of a Devonian oceanic crust. The klippe culminates with high-P amphibolite and gneiss units (Rötzler et al., 1999; Klemd, 2010) probably representing the Upper Allochthon. In Wildenfels, only the Upper Allochthon is represented by augengneisses, gneissic micaschists, amphibolitic gneisses and amphibolite units (Klemd, 2010).

Kinematic criteria indicate final emplacement of the Münchberg klippe toward the NW, but previous kinematic criteria indicate top to the SW or WSW in augengneisses and mylonites of the Upper Allochthon (Blümel et al., 1990). Interestingly, a similar kinematics has been described in the Erbendorf-Vohenstrauβ Unit, at the base of the Moldanubian Zone (Franke, 1990; Blümel et al., 1990).

In the eastern branch of the STZ, the Allochthon occurs in the west and central Sudetes, surrounding the Krkonoše Massif, and in the Góry Sowie Massif respectively. Figure 7a shows a section partly oblique to the structure, but subparallel to it in the eastern branch of the STZ. It cuts across the Erzgebirge, Krkonoše and Orlica-Śnieżnik domes, which are equivalent (not successive) structures which delineate the Bohemian Arc. This part of the profile is intended to show the suggested mutual relationships among the different allochthons.

South and SE of the Krkonoše Massif, three units have been described. The lowermost of them crops out in just a small area SW of the pluton. It consists of very low grade Devono-Carboniferous sediments including Culm facies on top, and is considered the Parautochthon

by Mazur and Aleksandrowski (2001). Above, the South Krkonoše Unit crops out along the southern flank of the dome. It is formed by garnet-free micaschists, phyllites and marbles with a high proportion of metavolcanics, of which the metabasites show relics of blueschist facies (1.2-1.5 GPa; Majka et al., 2016; Konopásek et al., 2019), dated around 365 Ma (Maluski and Patočka, 1997; Konopásek et al., 2019).

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

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

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

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

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

5.4. Teplá-Barrandian and Moldanubian

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

Two eclogitic units occur at the base of the Teplá-Barrandian and Moldanubian zones, Mariánské Lázně and Erbendorf-Vohenstrauβ respectively (Figs. 6 and 7b). The Teplá-Barrandian Zone is considered either a separate terrane (Matte et al., 1990; Franke, 2014) or part of the Moldanubian terrane, occupying a relatively higher structural position, and separated from the Moldanubian Zone by large-scale normal and wrench faults (Pitra et al., 1999; Franke, 2000; Zulauf et al., 2002). It is possible that these faults hide the lateral transition between two parts of the same terrane with different characteristics (Schulmann et al. 2009; Lardeaux et al., 2014). In this case, the two zones would represent the upper ensemble of the Bohemian nappe pile, perhaps being equivalent to the Upper Allochthon of the French and Iberian massifs. Then, the two eclogitic units occurring at the base of the Teplá-Barrandian and Moldanubian zones (Mariánské Lázně and Erbendorf-Vohenstrauβ respectively; Figs. 6 and 7b) would represent the suture of such a terrane. In the next sections

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

5.4.1. Moldanubian Zone (MZ)

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

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

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

as a thrust sheet emplaced above the MZ, while Schulmann et al. (2005, 2009, 2014b) describe the evolution of the MZ in terms of subduction of the STZ beneath a continental terrane. The association of granulites and durbachites in the MZ has been interpreted as a Visean product of relamination, as defined by Hacker et al. (2011) in subduction zones bearing sediments or involving continental or arc crust. It implies emplacement of felsic gneisses derived from lower plate and incorporated to the upper plate, mostly at the lower crust. In the MZ, partially molten gneisses derived from the subducted continental lower crust of the STZ may have penetrated horizontally from the subduction channel to the SE relaminating the Moldanubian lower crust, part of which extruded later vertically reaching upper crustal levels (Lexa et al., 2011; Kusbach et al., 2015).

Chopin et al. (2012) have proposed flat subduction of the STZ continental crust beneath the Teplá-Barrandian rocks and extrusion of ultrahigh-P to high-P rocks in the Orlica-Śnieżnik Dome. If true, the felsic granulites with high-P relics could derive directly from this crust, and their occurrence can be explained by extrusion from the STZ crust and diapiric intrusion close to the Moldanubian Thrust in the SE, and also in the central part of the MZ (Fig. 7b; Lexa et al., 2011; Schulmann et al., 2014b; Kusbach et al., 2015).

5.4.2. Teplá-Barrandian Zone (TBZ)

Underlying the Mariánské Lázně and Erbendorf-Vohenstrauβ units, an association of low-grade metasediments, basic rocks, acid tuffs and serpentinites reaching at most epidote-amphibolite facies occurs (Kladská unit of Kachlík, 1997). It overlies the Thuringian-facies autochthonous Paleozoic and is similar to the unit with phyllites and prasinites of Münchberg.

The Mariánské Lázně Unit consists of amphibolites, often migmatitic, with minor alternations of felsic gneisses, and rare metasediments and pyroxenites. An eclogite-amphibolite zone occurs between a basal layer of serpentinite and an overlying gabbro-amphibolite zone (Collett et al., 2018). The protolith age of eclogites and amphibolites is *c*. 540 Ma old, while gabbros in the upper zone, often coronitic, have yielded 500 Ma (Timmermann et al., 2004, 2006). Lu-Hf and Sm-Nd garnet-whole rock geochronology yield *c*. 390 Ma for high-P metamorphism (*c*. 2.5 GPa, 650-750 ºC) and 375 Ma for cooling after a high-T overprint, in accordance with the previous age data (Collett et al., 2018 and references therein). Calc-alkaline gabbros intruded the Mariánské Lázně eclogites during their syncompressional and later extensional exhumation at *c*. 380 Ma (Deiller et al., submitted).

Overlying the Mariánské Lázně Unit, the Teplá Crystalline Complex is a band 10 km wide of Neoproterozoic metasediments and metabasalts where a Cambro-Ordovician low-P/high-T metamorphic event (*c*. 485 Ma) was overprinted by a Late Devonian, Barrovian event related with crustal thickening (*c*. 375 Ma), and then by low-P/high-T metamorphism related to a giant detachment along which the Maránské Lázně complex was exhumed (Peřestý et al., 2017).

The thick Neoproterozoic succession of the TBZ starts with Cadomian arc-related magmatism at the east (Davle group) which characterizes the lower part of this zone (Hajná et al., 2011), followed by late Ediacarian to Cambrian ocean floor stratigraphy and turbititditic accretionary wedge sequences of the Štěchovice group (Hajná et al., 2018; Ackermann et al., 2020). The uppermost unit is represented by a weakly metamorphosed Paleozoic section preserved at the core of the Praha or Barrandian Syncline. Metamorphic grade decreases upward from the basal Mariánské Lázně and Erbendorf-Vohenstrauβ units through a thick Neoproterozoic succession until the core of the Praha Syncline (Peřestý et al., 2017).

Above, a medium to low-grade Neoproterozoic succession is followed by the Paleozoic of the Praha Syncline. This sedimentary pile starts with an Ordovician silicicalstic succession (485-443 Ma) followed by early Silurian graptolitic shales which contiuously passes to late Silurian and Lower Devonian (Lochkovian and Pragian) limestones. The sequence terminates with siliciclastic deposition of the Givetian Srbsko Formation, which contains detrital zircons dated at 382 Ma (Vacek et al., 2019). Similarities between this sequence and that of the Iberian Autochthon had been claimed based mostly in stratigraphic and faunal correlations (Verneuil and Barrande, 1885; Delgado, 1908; Gutiérrez-Marco et al., 1999b). However, a study of Ordovician trilobites carried out by Ballèvre (2016) shows that a very limited number of species are common between the Barrandian and the Armorican and Iberian Autochthon, which points to the TBZ as a terrane separated from mainland Gondwana at the Cambro-Ordovician transition.

The first event affecting the TBZ produced upright folding and development of a NE-SW trending vertical fabric associated to crustal thickening at around 380 Ma (Hajná et al., 2012; Peřestý et al., 2017; 2020). The folding of the Praha syncline had to occur after the Givetian hiatus. Its age can correlate with formation of strong vertical cleavage affecting the eastern margin of the TBZ (Jílové Zone) which developed coeval with the syntectonic intrusion of arc-derived magmas of the Central Bohemian Pluton (Žák et al., 2005). Therefore

the shortening of the eastern TBZ and folding of Praha synforms can be indirectly dated at 360-350 Ma (Hajná et al., 2012; Vacek and Žák, 2017).

5.5. Variscan intrusive magmatism in the Bohemian Massif

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

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

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

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

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

6. Morocco

The European Variscan Belt continues in NW-Africa south of Iberia by the Rif in northern Morocco, the Moroccan Meseta and its Algerian continuation to the east, and the Anti-Atlas. South of the Maghrebides, an Alpine belt which incorporates Variscan relics mainly in the Rif, the Moroccan Meseta represents the little or no Alpinized continuation of the European Variscan internides while the Anti-Atlas basically represents a foreland of the belt. The two latter domains are separated by the South Meseta Zone, a dextral transpressive fault system trending WSW-ENE, running more or less parallel to the later Alpine High Atlas (Michard et al., 2010a), and bounded to the north by the South Meseta Fault (SMF; Figs. 1 and 8) or Atlas Paleozoic Transform Zone (Hoepffner et al., 2006), and to the south by the South Meseta Front (SMFr; Michard et al., 2010a, b; Cerrina Feroni et al., 2010; Ouanaimi et al., 2016, 2018). South of the South Meseta Front, all basement units outcropping in inliers belong to the northern margin of the West African Craton, which was here only reactivated by thick-skinned tectonics during the Variscan event (Burkhard et al., 2006; Robert-Charrue and Burkhard, 2008; Baidder et al., 2016). The South Atlas Fault (SAF) is an Alpine structure bounding the High Atlas. The segments of the Variscan belt exposed in southernmost

Morocco and the northwestern corner of Mauritania (Zemmour area; Souttoufides of Villeneuve et al., 2015) and those defining the Mauritanides beyond the Reguibat Shield in western Mauritania (Villeneuve, 2005) are not included in this description.

6.1. Rif Domain

Pieces of the Variscan Belt are incorporated into the Alpine belt describing the Alboran Arc, with the Betic Range in the north (Spain) and the Maghrebides (Rif in Morocco) in the south (Do Couto et al., 2016). In the Betics, the sequences vary from little and no-metamorphic Ordovician to early Carboniferous sediments in the Maláguide Complex (Martín-Algarra et al., 2004) to metasediments difficult to separate from Mesozoic sediments with Alpine metamorphic imprint in the Alpujárride and Nevado-Filábride complexes (Alonso-Chaves et al., 2004; Puga et al., 2004). The deposition of the clastic protolith of the latter complex was recently determined using detrital zircon ages ranging between 349-282 Ma (Santamaría-López and Sanz de Galdeano, 2018). Based on Carboniferous conodonts found in the lowest Nevado-Filábride Unit, Rodríguez-Cañero et al. (2018) correlate that Variscan basement with the Cantabrian Zone, while occurrences in the Alpujárride and Maláguide complexes, integrated in the Alborán Domain, are placed in the northern margin of a Paleotethys Ocean.

High-grade Variscan metamorphic rocks are present in both Betic and Rif cordilleras in southern Spain and Morocco. The metamorphic core of this previously thickened Cenozoic orogen was dismembered during Neogene extension forming several discontinuous outcrops along the Alboran Arc (Michard et al., 1997; Do Couto et al., 2016; Gimeno-Vives et al., 2020). The polymetamorphic Variscan outcrops are found in the Internal Zone of the Alboran Arc, namely in the Alpujáride-Sebtide complex, which form the metamorphic envelope of the Ronda in the Betic Cordilerra and the Ceuta and Beni Bousera peridotite massifs in the Rif (*e.g.* Michard et al., 1997, Ruiz-Cruz and Sanz de Galdeano, 2013). The metamorphic pattern of these complexes is characterized by downward increase of metamorphic gradient with structurally deepest granulites, migmatites and higher gneisses and schists (Rossetti et al., 2010, Acosta-Vigil et al., 2014). These units reveal a polymetamorphic record of both Variscan and Alpine orogenic cycles for which the determination of age and pressure-temperature conditions remains problematic (Michard et al., 1997; Massone, 2014). However, it becomes increasingly evident that the granulite-migmatite unit of the Moroccan Sebtide

complex surrounding Beni Bousera peridotite preserves remnants of high-P/high-T late Carboniferous-Permian metamorphism reaching *c*. 1.6 GPa at 800 ºC (Bouybaouene et al., 1998). These granulites and granite sills representing products of deep crustal anatexis were dated at 300-290 Ma using U-Pb zircon method (Rossetti et al., 2010, 2020). Similar peak pressure conditions of 1.2-1.4 GPa at 850 ºC followed by reequlibration at 0.5-0.6 Gpa and 800-850 ºC were determined in the anatectic Jubrique envelope of the Ronda massifs (Barich et al., 2014). Here, the age of anatexis was determined by U-Pb zircon geochronology to 280-290 Ma (Acosta-Vigil et al., 2014) confirming EMP monazite age *c*. 287 Ma from the same unit (Massonne et al., 2014). These data corroborate the previously obtained metamorphic 300 Ma ages from amphibolite facies schists (Zeck and Williams, 2000; Zeck and Whitehouse, 2002) and *c*. 283 Ma age of late Variscan pegmatite magmatism (Sánchez-Navas et al., 2014) from the metamorphic core of the central Betic Cordillera. Similar *c*. 281 Ma granite magmatism was also documented by U-Pb zircon geochronology of granite pebbles from a Tertiary conglomerate lying on a Ghomaride nappe of the eastern Rif (Olivier and Paquette, 2018).

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

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

6.2. Meseta Domain

Although the main part of the Meseta is covered by Meso- and Cenozoic sediments, several massifs allow the separation of various domains (Fig. 8), namely the Eastern and Western Meseta, the later including the Coastal and Sehoul blocks (Piqué and Michard, 1989). Except for the Coastal Block, these domains are characterized by a deformation stronger than in the Anti-Atlas, associated metamorphism, pre- and late orogenic magmatism and weakly deformed continental late Carboniferous and Permian deposits. The paleogeographic significance of these domains, limited by extensive shear zones (Michard et al., 2008; Accoto et al., 2020), however remains debatable, and only the Sehoul Block have a clearly distinctive signature (Tahiri et al., 2010; Pérez-Cáceres et al., 2017). The domain of strong Variscan deformation also includes most of the basement of the Atlas Mountains included within the South Meseta Zone, up to the South Meseta Front. It will be here described collectively with the Meseta Domain, yet this domain has Anti-Atlas affinities.

The preorogenic sequences can be compared with those of the Anti-Atlas. The lower Paleozoic is similar although the Cambrian carbonates are thinner and thick Cambrian clastics were deposited in graben structures fed by local sources and associated with tholeiitic volcanism, which accounts for early to late Cambrian extension (Pouclet et al., 2008; Letsch et al., 2018). The graben system might have had a southward prolongation toward the Anti-Atlas (Piqué, 2003). The Ordovician is comparatively thinner, with a marked distalisation toward the NE across the Western Meseta (Baudin et al., 2003). Lower Ordovician red beds comparable to those of the Armorican Massif and southern CIZ are reported in Central Morocco and the Eastern Meseta, but are absent in the Anti-Atlas and nearly so in most of the Coastal Block (Ouanaimi et al., 2016). The Hirnantian glacial episode is widely recorded throughout the Meseta, with the ice sheet having probably reached the Rehamna massif (Le Heron et al., 2007), and an ice-distal turbiditic succession being recognized in the Tazeka Massif (Ghienne et al., 2018). Extension continued during the early Paleozoic, marked by intraplate volcanism in the Lower Ordovician (El Hadi et al., 2014) and the Silurian (El Kamel et al., 1998). Carbonate platforms developed in the Devonian. A more pronounced tectonic activity was recorded during the Lower-Middle Devonian around the Variscan Western Meseta Shear Zone (WMSZ; Fig. 8), where coarse continental conglomerates filled restricted fault-bounded basins (Michard et al., 2008).

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

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

The Western Meseta (WM; Michard et al., 2010a) is divided into a western, almost undeformed part named the Coastal block, and an eastern part characterized by three principal Variscan massifs (Central, Rehamna, and Jebilet). This domain is separated from the Eastern Meseta by the Middle Meseta Fault Zone (MMFZ) and from the Coastal Block by the Western Meseta Shear Zone (WMSZ). The structural and Raman spectroscopy thermometry study of Lahfid et al. (2019) suggests that in the easternmost part of Western Meseta, the latest Devonian to early Visean folding event, reminiscence of the above-described tectonic evolution in the Eastern Meseta, was preceded by a high-T extensional Late Devonian event. The late Variscan evolution of the Western Meseta is debated, but NW-verging, late Carboniferous-Permian folds and thrust faults prevailed (Michard et al., 2010a). Chopin et al. (2014) proposed that a south- to SSW-verging shortening event occurred in the latest Carboniferous-earliest Permian, having no genetic connection with extensive Visean basins. This event is bracketed by Delchini et al. (2018) between 325-310 Ma, while a second deformation phase producing transpressional westward thrusting took place at 310-280 Ma in relation with the WMSZ. This scenario is more compatible with the late Carboniferous nappe tectonics recognized in the northern Anti-Atlas (Cerrina Feroni et al., 2010), as well as with the Anti-Atlas tectonic evolution (Baidder et al., 2016). Later, an Early Permian orthogonal convergence occurred, and a post-tectonic pluton was emplaced at *c*. 275 Ma.

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

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

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

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

6.3. Anti-Atlas

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

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

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

7. Correlation of the Variscan massifs

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

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

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

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

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

7.1. Correlations in the Autochthon

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

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

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

Concerning the Meseta, Michard et al. (2008) argued against the possibility that it became separated from Gondwana and disconnected from the Anti-Atlas during the opening of the Paleotethys, as proposed by Stampfli and Borel (2002). Using stratigraphic, faunal and paleomagnetic data, they support restricted displacements relative to Africa, at least for the main part of the Meseta.

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

while in the CIZ it does crop out only in the OVD. The southern CIZ, including the OVD, is characterized by more terrigenous and proximal Middle Ordovician facies, as the NAD (Robardet et al., 1990). So, although a NAD/CIZ correlation can be envisaged, a more precise link should be established only with the southern part of the CIZ (Figs. 1 to 5).

There is not a clear equivalent to the OMZ in the Armorican Massif. A possible candidate is the eclogite-bearing gneissic unit in the Léon, at the northwestern end of the Armorican promontory. Balé and Brun, (1986) interpreted it as derived from the SAD and so, representing a detached piece of the Allochthon. But its structural position is the same as that of the Central Unit in the OMZ: at the contact with the OVD in Iberia and with the NAD in Brittany. Rolet et al. (1986) consider the possibility of underthrusting of the Léon under the rest of the CAD at the Upper Devonian. We will assume this idea and also follow Schulz et al. (2007) regarding the eclogite-bearing unit in the Léon as a potential link between the OMZ and the STZ of the Bohemian Massif (Fig. 1).

The continuation of the OMZ may be found further to the east, in the relatively small outcrops of the Vosges and Schwarzwald to the north of the Lalaye-Lubine and Baden-Baden shear zones. There, low- to medium grade Cambro-Ordovician to Silurian deposits are overlain by marine Middle and Upper Devonian marine facies and early Carboniferous flysch ascribed to the STZ of the Bohemian Massif, which is considered equivalent to the OMZ (Pereira et al., 2006a; Linnemann et al., 2008).

In the SAD, the Autochthon can be compared with the Paleozoic of the northern part of the CIZ by the occurrence of Cambro-Ordovician monotonous schists and porphyroids, the possible existence of the Armorican Quartzite, the Middle Ordovician gray slates and the Silurian black cherts. The latter, as well as some synorogenic deposits, could form part of the Parautochthon, which includes metavolcanic Ordovician porphyroids, as in the Cabo Ortegal and the Portuguese complexes.

It is worth noting that while the Autochthon in the SAD correlates with that on the northern CIZ east and south of the Iberian Allochthon, the CAD is rather comparable with the central part of the CIZ. This correlation, as well as that of some allochthonous units, drove Ballèvre et al. (1992) to propose the existence of a suture between the CAD and SAD which would continue by NW and central Iberia. But in NW Iberia, the suture represented by the Middle Allochthon, is rootless, only preserved in klippen, and does not continue to the SE of the CIZ (Fig. 2a). This is one of the problems we try to solve in this contribution.

In the FMC, the Montagne Noire (MN) can be correlated with the Pyrenees and continue, in the other side of the Bay of Biscay, with the branch of the CIZ limiting to the east and south the NW Iberian Allochthon (GTMZ) and forming also its relative autochthon. The Paleozoic sequence in the Pyrenees and MN shows differences with the CIZ and CAD as, for instance, the lack of the characteristic Armorican Quartzite, and younger or more lasting early Paleozoic felsic magmatism, which is here reported up to the Upper Ordovician. While the MN represents the only outcrop of the Autochthon in the FMC, the occurrence of the Parautochthon in several tectonic windows suggest that the massif is partially or totally underlain by an autochthonous crust comparable to that of the CIZ (Figs. 4 and 5b, c).

In Sardinia, the Autochthon of Iglesiente represents a less-deformed domain of the northern Gondwana foreland. The presence of the Upper Ordovician Sardic unconformity, as in the Pyrenees (Casas, 2010) confirms that it forms part of the northern Gondwana Mediterranean branch. This, as younging of Ordovician magmatism only reflect lateral changes along the northern Gondwana paleomargin, whose western continuation through the Pyrenees and the Bay of Biscay reaches the part of the CIZ underthrust beneath the GTMZ. There are however differences with the Iberian Parautochthon, where no calc-alkaline Middle-Upper Ordovician arc is clearly represented.

7.1.3. Links among Bohemian and western Variscan autochthons (STZ, OMZ, CIZ, RHZ, SPZ, MSZ)

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

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

faunas in eastern Germany compare with those of France, Iberia and Morocco (Elicki, 2007), with links pointing to the CIZ and the Montagne Noire. Upper Ordovician glacigenics exist in the STZ and OMZ, as well as in the other autochthonous zones of Iberia (CIZ, WALZ and CZ) and in the Central and Northern domains of the Armorican Massif.

Subsidence patterns calculated for the Barrandian, the STZ and the eastern part of the CIZ in Iberia are comparable, less regarding the OMZ, but they are especially striking for the Cambrian, and start to diverge from the Lower Ordovician, when separation of the future Upper Allochthon took place (von Raumer and Stampfli, 2008). Even if the patterns are considered similar during the rest of the Paleozoic, this can be attributed to symmetrical evolutions of the continental margins of Gondwana on one side and the pulled apart terrane on the other.

With these data, a broad link can be established among the Autochthon of Iberia, France and Bohemia, namely the STZ, the CAD and NAD of western France and the Iberian OMZ and CIZ. A more accurate correlation can be attempted using detrital zircon age spectra. Linnemann et al. (2008) have shown that spectra are similar for Ediacaran, early Cambrian and Lower Ordovician metasediments of the STZ and SW Iberia (OMZ and OVD). The spectra in the CIZ, WALZ and CZ are also similar except for a significant Mesoproterozoic population centered at 1000 Ma (Gutiérrez-Alonso et al., 2003; Martínez Catalán et al., 2004b), which is absent or nearly so in the OMZ, STZ and NAD (Linnemann et al., 2008; Dabard et al., 2021). The source for this population is the East African-Arabian Zircon Province (Gómez Barreiro et al., 2007; Bea et al., 2010), fed by the eastern Gondwana super-fan system (Meinhold et al., 2013; Stephan et al., 2019). The abundance of this population in the CIZ, CAD, WALZ and CZ is related to the paleoposition of this part of the Iberian Massif close to present northern Africa, whereas its scarcity or absence indicates a more westward position of the OMZ and quite probably, of the STZ, both of which facing a more western segment of Northern Africa in agreement with the faunal record (Gutiérrez-Marco et al., 2002, 2016). Also it is worth noticing that STZ, NAD and WM share pieces of an Eburnean (*c*. 2.05 Ga) basement most likely derived from the West African Craton (Pereira et al., 2015b). This is in line with a relatively western derivation of the STZ of Bohemia, a character that might have been acquired since the Cadomian (Pérez-Cáceres et al., 2017), and amplified during the Variscan orogen.

Another affinity between STZ and OMZ is the existence of intrusives related with the closure of the Rhenohercynian Ocean. In the STZ they were emplaced in the MGCH at 360-

325 Ma, while equivalent rocks intruded at 355-335 Ma mostly in the SW of the OMZ and eastern SPZ (Figs. 2, 4 and 6). Also supporting the correlation, the eclogites in the Odenwald could have their counterpart in those of the Cubito-Moura Unit, in the southern OMZ. Eclogite-facies metamorphism is probably older than 363-350 Ma in the Odenwald (Scherer et al., 2002) and is dated at 371±17 Ma in the southern OMZ (Moita et al., 2005).

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

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

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

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

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

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

7.2. Correlations in the Allochthon

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

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

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

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

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

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

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

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

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

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

The best correlation can be established between the Upper Gneiss Unit and the Upper Allochthon: same lithologies, igneous protolith ages (Cambro-Ordovician) and metamorphic evolution. Parts of the Leptyno-Amphibolitic Complex share with the catazonal Iberian Upper Allochthon the Lower Devonian age of high-P metamorphism. No ultrahigh-P metamorphism has been described in NW Iberia, but at least 1.9-2.0 GPa were calculated for eclogites and ultramafics of the Cabo Ortegal Complex (Mendia, 2000; Mendia and Gil Ibarguchi, 2005).

Above the Leptyno-Amphibolitic Complex, but still in the Upper Gneiss Unit, migmatitic paragneisses and orthogneisses devoid of high-P relics may represent the high-grade upper units of the Upper Allochthon. The uppermost part of the nappe pile in the FMC, consisting of low- to medium grade Cambrian greywackes, quartzites and rhyolites, have also an equivalent in the Upper Allochthon of the Órdenes and Morais complexes.

The units of the Central Vosges and the Schwarzwald gneiss complexes have been compared with the Leptyno-Amphibolitic Complex of the FMC, particularly the part between the Varied and Granulite units (Pin and Vielzeuf, 1988; Fabriès and Latouche, 1988; Lardeaux et al., 2014). So, these units might be considered part of an Upper Allochthon similar to that of NW Iberia and to the Teplá-Barrandian Zone of the Bohemian Massif. However, the young age of high-P metamorphism in the Schwarzwald (*c*. 340 Ma; Kalt et al., 1994; Marschall et al., 2003), the fact that granulites possibly represent vertical extrusions at mid-crustal levels, and the ultra-potassic magmatism dated at 345-335 Ma rather suggest correlation with the Moldanubian Zone of the Bohemian Massif (Schulmann et al., 2002; Skrzypek et al., 2014).

Nevertheless, the Variscan evolution can still be correlated with that of the FMC because: i) the ophiolites of the Ligne des Klippes in the VM are equivalent to the Brévenne Unit, a back-arc related with the subduction of the Saxothuringian Ocean (Faure et al., 1997; Schulmann and Gayer, 2000; Lardeaux et al., 2014); ii) the overlying Visean flysch and the Devono-Carboniferous sediments and volcanics of the metasedimentary belt separating the two gneiss complexes in the SWM can be equivalent to the Tuffs Anthracifères, related to the subduction of the Rhenohercynian Ocean, whose influence was probably responsible for the subsequent voluminous granitic plutonism. Affinities of the VM and SWM with both the FMC and the Bohemian Massif are not surprising given their position. The differences and the possible causes have been discussed by Lardeaux et al. (2014), and will be dealt with below in the correlation with the Bohemian Massif.

The massifs of the southern Variscan Belt occur in continuation of, and their units can be correlated with the southern FMC. The internal zones are represented in the MTM by the Central and Eastern domains which together seem to represent the Upper Allochthon, although they might also include a dismembered suture according to Schneider et al. (2014). In the Western Domain, one of the thrust sheets bears a rift-related succession with alkaline rocks possibly representing the Lower Allochthon, although no high-P has been described. In any case, the Middle Allochthon is missing or hidden in the Leptyno-Amphibolitic Complex, as in the FMC. The Central Domain presently underlies an inverted early Carboniferous synorogenic flysch which is in turn overlain by a Paleozoic sequence including Silurian black shales (Bronner and Bellot, 2000). The original stacking order was the opposite, with the Eastern Domain above and the Western Domain below. So, the Western Domain may represent the Parautochthon and/or the Autochthon.

In Corsica, most metamorphic outcrops represent a Leptyno-Amphibolitic Complex similar to that of the FMC, with eclogites and high-P/high-T granulites, which can be assimilated to the Upper Allochthon. The uppermost Allochthon crops out in northern Corsica as a low-grade Neoproterozoic and Paleozoic sequence that Rossi et al. (2009) interpreted as different from that of the Autochthon in SW Sardinia and the Montagne Noire, and deposited in a peri-Gondwanan microcontinent. It may find its equivalents in the uppermost Allochthon of NE Iberia (Órdenes and Morais complexes), the SAD (Mauges and Roc-Cervelle units) and the FMC (Thiviers-Payzac Unit).

In Sardinia, the Internal Zone represents the high-P/high-T part of the Upper Allochthon, whereas the Internal Nappes of the Nappe Zone may correlate with the Lower Allochthon on the basic of its eclogite-facies event in absence of high-P granulites. From a structural point of view, the External Nappes constitute a Parautochthon, which is coherent with their metamorphic Paleozoic sequence representing in this case only the preorogenic part of the succession.

In the External Massifs of the Alps, the metamorphic evolution of the third domain described by Ballèvre et al. (2018), characterized by the high-P eclogitic/granulitic event, may be compared to that of the FMC, VM and SWM (von Raumer and Neubauer, 1993). Consequently, it is correlated with the Upper Allochthon even if not all available ages of the high-P event yield the Lower Devonian age obtained in these and other Variscan massifs. The Chamrousse Ophiolite may represent the same ocean whose relicts have been preserved in the Middle Allochthon of NW Iberia and in the Audierne Bay Complex of the Armorican Massif.

The ascription of the remaining domains is more problematic. The association of a Devono-Carboniferous bimodal suite with conglomerates underlying the ophiolite (second domain), suggest a correlation with the Morvan Arc of the FMC. Perhaps also with the migmatitic gneisses, early Carboniferous amphibolite and gabbro boudins in the southern SWM and/or with the Devono-Carboniferous sediments and volcanics of the metasedimentary belt separating the two gneiss complexes in the SWM. Finally, the undated low- to medium-grade schists of the first domain could represent the upper units of the Upper Allochthon (the option adopted in Fig. 4) because they are considered part of the polymetamorphic group by von Raumer et al. (1993), but might also represent the Parautochthon or even the Autochthon.

7.2.2. Links among Bohemian and western Variscan allochthons (STZ, TBZ, MZ, VM, SWM, FMC, SAD, GTMZ)

The correlation described among the three groups of allochthonous units of the GTMZ and the SAD (Ballèvre et al., 2014) can be extended to the STZ allochthons based on the lithological association, structural sequence and ages. All the Iberian and Armorican groups are well represented in the STZ, although not always the three allochthons occur in one single massif.

In the western branch of the STZ, the Münchberg massif exhibits the most complete succession. Its nappe stack recalls that of Cabo Ortegal and Bragança complexes. There, as in Münchberg, the Parautochthon consists of a synorogenic flysch with an early Carboniferous maximum depositional age, overlain by a relatively distal preorogenic Paleozoic sequence (Dias da Silva et al. 2014a; Martínez Catalán et al., 2008, 2016). In Münchberg, the Lower Allochthon is not present while in Cabo Ortegal it is only represented by a small unit. The only identified oceanic units witnessing the Middle Allochthon are Devonian phyllites, prasinites and amphibolites equivalent to the Lower Devonian ophiolites of NW Iberia. Finally, the Upper Allochthon is represented mainly by the high-P/high-T units, with only a small unit of the intermediate-P group in Cabo Ortegal, and no equivalent in the western STZ.

The three allochthons are also well represented along the eastern branch of the STZ, from the lower part of the nappe pile south of the Krkonoše Massif to the upper part in the Góry Sowie Massif. A small low-grade unit partially synorogenic represents the Parautochthon, while the South Krkonoše Unit depicts the lithological association and Late Devonian high-P and relatively low-T metamorphism of the Lower Allochthon in Iberia and

the Armorican Massif. To the east, the Kaczawa Complex represents the Middle Allochthon, with the rocks of oceanic affinity being of Cambro-Ordovician, Silurian and Devonian age, of which, the first and the later are also identified in NW Iberia (Martínez Catalán et al., 2020). Upper Devonian-early Carboniferous flysch with chaotic assemblages and tectonic mélanges in the Kaczawa Complex may have an equivalent in the Cabo Ortegal mélange, formed in an Upper Devonian subduction zone (Martínez et al., 2012; Arenas and Sánchez Martínez, 2015). A basaltic unit of probable Late Devonian age (Seston et al., 2000), may represent a series of young ophiolites existing in the French Massif Central (Brévenne; Génis; Pin and Paquette, 2002; Faure et al., 2009) and the Vosges (Ligne des Klippes; Skrzypek et al., 2012; Lardeaux et al., 2014), but not identified in NW Iberia. Further east, the Góry Sowie Massif represents the high-P/highT part of the Upper Allochthon, with identical protolith ages, metamorphic evolution and structural position as the high-P/high-T units of the GTMZ in NW Iberia and the Münchberg Massif in Germany (Martínez Catalán et al., 2020; Tabaud et al., 2020).

In the STZ allochthons of Münchberg, Wildenfels, Frankenberg, Krkonoše and Góry Sowie massifs, Upper Allochthon units showing an intermediate-P Cambro-Ordovician event are absent. In NW Iberia, the Armorican Massif and French Massif Central, these units occupy the uppermost position of the nappe pile, so that their absence here can be explained by erosion. Moreover, this absence is compensated by the dominant occurrence of low-P Cambro-Ordovician metamorphism in the TBZ and MZ, which together represent the Upper Allochthon in the Bohemian Massif. Only in the Orlica-Śnieżnik Dome the STZ autochthon is surrounded by rocks of the MZ which in the eastern part bear the imprint of Cambro-Ordovician medium-P/high-T metamorphism (Staré Město Unit; Štípská et al., 2001).

In the west of the TBZ, the basal association of low-grade metasediments, basic rocks, acid tuffs and serpentinites under the Mariánské Lázně and Erbendorf-Vohenstrauβ units can be correlated with the phyllites and prasinites of Münchberg, and represent the Middle Allochthon. For the rest of the TBZ, a correlation may be established with the Upper Allochthon of the GTMZ (NW Iberia), SAD and FMC.

The Mariánské Lázně, Erbendorf-Vohenstrauβ units and the Góry Sowie Massif underwent a high-P/high-T metamorphic event during the Lower Devonian, followed by Middle Devonian exhumation in amphibolite facies. The same happened in the lower units of the Upper Allochthon in the GTMA (Cabo Ortegal, Órdenes and Bragança complexes), SAD (Essarts and Audierne Bay units) and the FMC (Leptyno-Amphibolitic Complex of the Upper

Gneiss Unit). Above, the Teplá Crystalline Complex recorded a history similar to that of the upper group of the Upper Allochthon in the Órdenes and Morais complexes, with the difference that the Cambro-Ordovician metamorphic event (495-485 Ma) is of low-P in Teplá and of intermediate-P in NW Iberia. The Essarts Unit in the SAD did also record an Ordovician high-T event before Devonian eclogitization. In NW Iberia and the TBZ, Devonian shortening under Barrovian P/T conditions was followed by a low-P event connected with extension (Gómez Barreiro et al., 2006; Peřestý et al., 2017). The metamorphic contrast between the structurally upper and lower groups of the Upper Allochthon (Cambro-Ordovician low or intermediate-P above, Lower Devonian high-P/high-T below) is quite strong in all cases, accentuated by extensional shear zones and faults.

A thick Neoproterozoic succession with Cadomian arc-related and Cambro-Ordovician rift-related magmatism characterizes the lower part of the TBZ (Dörr et al. 2002; Drost et al. 2004) whose uppermost unit is a rather complete and weakly metamorphosed Paleozoic succession preserved at the core of the Praha or Barrandian Syncline. The metamorphic grade decreases upward from the basal Mariánské Lázně and Erbendorf-Vohenstrauβ units through the Teplá Crystalline Complex (Peřestý et al., 2017) and a thick Neoproterozoic succession, until the core of the Praha Syncline.

The Neoproterozoic sequence of the TBZ is compared with that of the Mauges Unit in the SAD, while in the GTMZ, the age of the high-grade metasediments is unconstrained. For the Paleozoic succession, several species of Middle Ordovician trilobites of the uppermost Allochthon in the SAD are also found in the TBZ, while only one is found in the Iberian Autochthon (Ballèvre, 2016), thus supporting separation from mainland Gondwana at Cambro-Ordovician times. In the FMC, the Paleozoic is represented by the Cambrian Thiviers-Payzac Unit. In the GTMZ, the Paleozoic succession is terrigenous and monotonous in the uppermost units of the Órdenes and Morais complexes, and its age is constrained to the middle-upper Cambrian by detrital zircons and intrusive magmatism (Fernández-Suárez et al., 2003; Fuenlabrada et al., 2010; Arenas et al., 2016). The latter, dated at 510-490 Ma, is bimodal, with arc affinities in the upper units (Andonaegui et al., 2002, 2012) and extension-related in the lower units of the Upper Allochthon (Galán and Marcos, 1997).

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

et al., 2014). However, correlation is straightforward with the southern small massifs: the Leptyno-Amphibolitic Complex is recognized in the Central and Eastern domains of the Maures, Corsica and the Internal Zone of Sardinia. That correlation can be extended at least to the third domain described in the External Massifs of the Alps by Ballèvre et al. (2018), characterized by a polymetamorphic evolution including eclogites or high-P granulites. Very interesting is the uppermost allochthonous unit of northern Corsica, with Neoproterozoic terrigenous sediments, interbedded metabasalts and a rather complete Paleozoic sequence, with a possible equivalent in the Mauges Unit of the SAD and perhaps also in the upper parts of the TBZ.

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

Due to differences in the evolution between the Massif Central and Bohemian massifs, Lardeaux et al. (2014) proposed a division of the Moldanubian Zone between western and eastern parts, establishing as a possible limit the Bray Fault. That leaves the VM and SWM in the eastern Moldanubian, even when the authors interpret the northernmost FMC and the Southern Vosges as part of the same unit, as suggested by Jung and Roques (1952). However, similarities of the Central Vosges and the Schwarzwald gneiss complexes with the Moldanubian exist, namely the relatively young age of high-P metamorphism, the possible vertical extrusions of granulites to mid-crustal levels and the early-middle Carboniferous ultra-potassic magmatism (Kalt et al., 1994; Schulmann et al., 2002; Marschall et al., 2003; Skrzypek et al., 2014).

It seems that the VM and SWM share characteristics and events with both the Bohemian and French Central massifs, in agreement with their relative geographic positions. Taken together, their tectonic, metamorphic and magmatic evolution seems to change gradually from east to west. For the Variscan magmatism, the change is abrupt between the eastern and western parts of the FMC, an additional argument against the role of the Bray Fault. These changes are related with the subduction of the Saxothuringian and Rhenohercynian zones under the Allochthon which to the west (SAD and GTMZ) apparently did not occur.

7. 3. Mid-Variscan suture and Mid-Variscan Ocean

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

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

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

The common feature of the suture represented by the Middle Allochthon is that it exists in all the large massifs strictly representing the Variscan Belt, *i.e*. the large central and western European massifs. And the suture occurs roughly in the middle of the belt. Mid-Variscan suture would then be a better term, the disappeared sea could be referred to as the

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

8. Restoring late Variscan deformation

8.1. Scope and revision of age data

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

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

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

For these reasons, restoring wrenching deformation to the scale of the belt will be necessarily approximate, retrodeformation can only be partial and the successive steps might not represent the situation at a given time interval. The approach followed is to restore the main strike-slip shear zones and faults in steps marked by cross-cutting relationships to the map scale. The younger structures are restored first, and lately, the late Variscan internal deformation of a large part of the Variscan Belt is restored considering it as a broad dextral ductile shear zone based on the relationship of ductile shear zones with late folds (Iglesias Ponce de León and Choukroune, 1980; Gumiaux, 2003; Gumiaux et al., 2004) and the fact that they are always present in the Allochthon and the surrounding Autochthon. The restoration of arcuate geometries (oroclines) is very imprecise, but it is one of the goals of this contribution, and also one that yields the more interesting results.

Age data, although not always precise, provide a time frame to constrain Variscan wrench-type deformation. In the Massif Central, Armorican and Maures-Tanneron Massifs, plutonism related with Variscan convergence and collision started at the Middle-Upper Devonian and extended until the early Permian. Much of that magmatism was synkinematic with the activity of transcurrent shear zones, the oldest of which are dated at *c*. 370 Ma in the South Armorican Domain and the Massif Central (Rolin et al., 2009). These authors link the 370-315 Ma wrenching activity with indentation of continental microplates during Varican collisional tectonics. Shear zones coeval with the emplacement of old plutons were sealed by younger ones, and were frequently reactivated later. Rolin et al. (2009) distinguished two families, the older of which is mostly dextral, striking N100º-130º E, and formed during the Famennian-early Visean (370-340 Ma). The younger is middle Visean-Namurian in age (340-315 Ma) and includes dextral shear zones striking N150º-160º E and N130º E, and conjugate sinistral shear zones striking N0º-90º E.

The older family of Rolin et al. (2009) affects the allochthonous units and locally its relative autochthon, and can be related to the emplacement of the Mid-Variscan Allochthon, which in NW Iberia ended at 345-340 Ma (Dallmeyer et al., 1997). The younger family is represented in the other the Variscan massifs, but mostly clustering at the end of the middle Visean-Namurian time interval or even later. Ductile deformation in the Elbe Shear Zone was dated by Hofmann et al. (2009) at 337-323 Ma (considering the maximum time span allowed by the error bars). The old age is based on the sigmoidal shape of a deformed granitic pluton, while the young age corresponds to an undeformed granite close to the EFZ and cross-cutting previous faults and cleavage. A wider time interval of 350-315 Ma is provided by Žák et al.

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

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

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

Transpression in the Pyrenees was probably associated to the NPF, one of a group of large NW-SE dextral faults and associated shear zones which also includes the SMF, SIF, BrF and EFZ (Fig. 1). They are oblique to the general SW-NE trend of the belt and overprint all previous structures. Consequently, they are thought to represent the latest significant movements and must be restored before the folds and shear zones they cut across. Some of these faults cannot be studied, because they are hidden or were reactivated as plate boundaries later, as the SIF between Iberia and Africa, or the NPF. Others, like the Elbe Fault Zone and

Bray Fault have been dated, but their age contradicts the structural evidence of being the latest significant Variscan structures.

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

8.2. Step by step restoration

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

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

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

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

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

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

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

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

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

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

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

folding may be that of Figure 11f if the later, generalized ductile dextral shear had a value γ = 1.5, and more acute if that value was lower.

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

9. Plate tectonics and Variscan evolution

9.1. Sources and approach

The Pangea configuration proposed by Wegener (1929) for the late Carboniferous, commonly known as Pangea A is generally accepted before the opening of the Atlantic Ocean. However, inconsistency between previous paleomagnetic poles of Laurentia and Gondwana led Irving (1977) to propose an alternative Pangea B model where Laurentia occupied a more western position, implying large dextral shearing to explain the transition from models B to A. Relative dextral displacement between Laurussia and Gondwana continued in principle from the late Carboniferous to the Permian and perhaps up to the Lower Triassic (Muttoni et al., 2009; Gallo et al., 2017).

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

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

Matthews_etal_GPC_2016_ContinentalPolygons.gpmlz

Matthews_etal_GPC_2016_Coastlines.gpmlz

Matthews_etal_GPC_2016_410-0Ma_GK07.rot. The latter allows reconstruction from 420 Ma. For the Cambro-Ordovician maps (500-470 Ma), the reconstruction tree Master2010_II.rot and animations 500.A97 and 470.A97 of GMap2015 were used (http://www.iggl.no/resources.html, accessed 5/5/2021).

The choice of Domeier and Torsvik (2014) reconstruction is because it is mostly based on paleomagnetism, although this method does not allow establishing paleolongitudes. The authors acknowledge that paleomagnetic poles are not abundant for the time interval of their reconstruction, and they also use eruptions in Large Igneous Provinces and kimberlite emplacements in relation with a mantle reference frame for provisionally establishing paleolongitudes, as well as additional geological and paleontological data. Moreover, the reconstruction of Domeier and Torsvik (2014) counts on the work done by Mathews et al. (2016) in the files ready to use for GPlates. We prefer this reconstruction also because it fits better our knowledge about the Variscan evolution in comparison with, for instance, one of the most popular alternatives, that of Stampfli et al. (2013).

Domeier and Torsvik (2014) stated that a number of Gondwana paleomagnetic data reveal a relatively stationary position about the South Pole near Angola during the Middle and Upper Devonian. Meanwhile, Devonian paleomagnetic data for the main Variscan massifs, although scarce, indicate that they remained at constant latitude. Then, to explain early Carboniferous collision with Baltica, they move them thousands of kilometers form northern Africa to Laurentia during the Devonian and turn them back to northern Africa during the Carboniferous, with the Iberian Massif returning to practically to the same position but after having experienced a counterclockwise rotation of 90º. And all that happened without changes in the shape of the massifs, which behaved as rigid elements.

According to stratigraphic similarities between the Autochthon of the Iberian Massif and Morocco, the existence of a suture between them had not been considered (Michard et al., 2008, 2010a). However, high-P metamorphism and new ages of high-T metamorphism in the Rif Domain and Betics point to a convergent plate boundary being active there toward the end of the Variscan cycle. Even so, correlation among zones of both sides suggests that relative movements did not modified essentially the relative position of Iberia and Africa. Based on

that correlation and in the evidence that strong deformation characterizes the Variscan Massifs, we will keep the model of Domeier and Torsvik (2014) for the major plates, inserting our own model regarding more specifically the evolution of the Variscan Belt.

9.2. Reconstruction

The main continental masses and previously described significant elements of the Variscan Belt are shown in Figure 12 in orthographic projection for different ages. For the two Cambro-Ordovician reconstructions, a projection centered in the South Pole has been used, while for the tothers, covering the 400-290 Ma time interval in 10 stages, a fixed geographic grid shows the absolute and relative motions of the elements ("features" in GPlates' slang). Features unrelated to Variscan evolution, such as northern and eastern territories (Siberia, China, etc.) are omitted. For a better understanding of the Variscan evolution, Figure 13 shows cross-section snapshots of continents, terranes and plate boundaries corresponding to the time frames of Figure 12. The sections vary for each frame in that figure to show different aspects of the tectonic evolution until 320 Ma.

9.2.1. Terrane individualization

The reconstruction starts at 500 Ma, when future Avalonia was still part of Gondwana and a wide passive continental margin developed to the north and NW of present Africa. A piece of the outermost margin of NW Gondwana was being pulled apart by slab roll back of a subduction zone which created a magmatic arc in the side facing the Iapetus Ocean. This was probably the same responsible for the Avalonia volcanic arc (Pickering and Smith, 1995, 1997), which continued further to the south along more than 2000 km in present South America forming the Famatinia-Puna Ordovician arc (Ducea et al., 2017). The detached piece formed later the Mid-Variscan Upper Allochthon. It was a microcontinent whose individualization may have started at 520-510 Ma and where magmatic and metamorphic peaks were reached around 500 Ma.

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

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

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

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

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

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

9.2.2. Early Variscan convergence

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

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

At 400 Ma, the oceanic lithosphere between Gondwana and the future Upper Allochthon is that of the Mid-Variscan Ocean preserved in the Middle Allochthon. It was partly the lithosphere of the Ordovician Rheic Ocean and partly new, created by the relative movements between Gondwana and the microcontinent during the Silurian and Lower Devonian. Details are unknown, but suprasubduction signatures have been found in some oceanic rocks of these ages, suggesting that intra-oceanic subduction zones and perhaps incipient volcanic arcs developed close to the microcontinent (Díaz García et al., 1999a; Pin et al., 2002, 2006; Sánchez Martínez et al., 2007b, 2011; Arenas et al., 2014).

The first early Variscan events were recorded in the Upper Allochthon during the Lower Devonian: high-P/high-T metamorphism reaching eclogite facies and dated at (415?) 400-380 Ma is identified in the lower group of all Variscan Massifs where the Upper Allochthon occurs (Schäfer et al., 1993b; Santos Zalduegui et al., 1996; Franke, 2000; Lardeaux et al., 2001; Ordóñez Casado et al., 2001; Kryza and Fanning, 2007; Godard, 2009; Koglin et al., 2018; Collett et al., 2018; Martínez Catalán et al., 2020). This event reflects subduction of the Upper Allochthon lower group under the units presently forming the upper group (Martínez Catalán et al., 2009). If this subduction zone was the same as that proposed by Domeier and Torsvik (2014), its polarity should be the opposite, (*i.e.*, to the NW; Fig. 13) because the part of the microcontinent subducted at *c*. 400 Ma was the rifted margin of the microcontinent, facing Gondwana, while the upper plate was the active Cambro-Ordovician margin, facing the Rheic Ocean (Fig. 12, 400 Ma).

That early subduction initiated the building of an accretionary wedge by incorporation of new elements. The Mid-Variscan Ocean was closed by oceanic subduction, underthrusting and imbrication by 380 Ma. This was followed by continental subduction of the extended external part of the northern Gondwana margin. Exhumation of the previously subducted part

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

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

The reconstruction is coherent with the evolution of SW Iberia at the contact between the OMZ and SPZ. There, an ophiolite dated at 480 Ma represents the original lithosphere of the Rheic Ocean (Fonseca et al., 1999; Pedro et al., 2010). But it is included in a high-P mélange forming part of an orogenic wedge and representing the closure of an ocean by SW-directed subduction at *c*. 400 Ma (Rubio Pascual et al., 2013b). An arc and back-arc pair, another high-P unit and a NE-verging accretionary wedge developed later and complex continent/ocean interactions, including an extensional event, occurred before a thin-skinned thrust belt developed in the South Portuguese Zone between 340-310 Ma (Simancas et al., 2004). Note that because the *c*. 90º late-Variscan counterclockwise rotation of SW Iberia, present SW-directed subduction and NE-verging thrusting correspond, before 340 Ma, to

NW-directed and SE-verging respectively, fitting the reconstruction of Fig. 12.The opening of the Rhenohercynian or Lizard-Giessen Ocean occurred in the latest Silurian-Lower Devonian (425-392 Ma; Zeh and Will, 2010; 405-390 Ma; Clark et al., 1998; Woodcock et al., 2007), causing the closure of the Rheic Ocean by the Upper Devonian-early Carboniferous. Later, the closure of the Rhenohercynian Ocean is indicated by eclogites of 350 Ma, and the collision and underthrusting of Avalonia started probably around *c*. 345 Ma in the South Portuguese Zone (Simancas, 2004), the Rhenohercynian Zone (Oncken et al., 2000; Zeh and Will, 2010), and the Moravo-Silesian Zone (Tomek et al., 2019).

9.2.3. Variscan collision

The collision of the accretionary wedge with Gondwana emplaced the Mid-Variscan Allochthon and was accompanied by the incorporation of the Parautochthon and of generalized compressional deformation. Recumbent folds formed in NW Iberia, and the first folds in the Central Iberian Zone and the Moroccan Eastern Meseta started developing in the earliest Carboniferous. The Saxothuringian Zone was covered by the Allochthon and partly subducted beneath. But there is no evidence that the same occurred to the west, in the Ossa-Morena Zone or in Morocco, where compression was probably related with collision of the Silurian-Devonian arc of Avalonian derivation. This is supported by zircon age spectra in Upper Devonian synorogenic flysch in the southern Ossa-Morena Zone and the northern part of Eastern Meseta, which points to an Avalonian source (Pereira et al., 2012; Accotto et al., 2020, 2021). That source may have been a piece of the Silurian-Devonian arc obducted at *c*. 360 Ma not far from the flysch occurrences (Fig. 12), while arc collision may explain the recumbent folds in the central part of the Ossa-Morena Zone and the subduction of the Cubito-Moura Unit in the south and of the Central Unit in the north. Both the latter occurred at the Upper Devonian (380-360 Ma) and are coeval with high-P metamorphism of the eclogitic unit in the Pays de Léon and in the Mid German Crystalline High.

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

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

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

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

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

9.2.4. Late Variscan deformation

The final emplacement of the Iberian Allochthon has been dated at 345-340 Ma, including thrusting onto the Parautochthon and of the whole ensemble onto the Autochthon (Dallmeyer et al., 1997; Martínez Catalán et al., 2007, 2009). This can be related to the closure of the Rhenohercynian Ocean and final collision with Laurussia. The continental subduction of the Saxothuringian Zone is dated between 370-340 Ma (Franke and Stein, 2000; Žáčková et al., 2010; Konopásek et al., 2019), although the peak of high-P metamorphism occurred at 360-350 Ma (Schmädicke et al., 1995; Rötzler and Plessen, 2010; Závada et al., 2021). Continuation of convergence until 340 Ma suggests a relationship with Laurussian push, which may also be the cause of propagation of early Variscan folds and thrust faults toward the foreland, and of the initial curvature of the fold axes that ultimately led to the Central Iberian Arc and possibly the Eastern Meseta Arc (Figs. 1, 2 and 8), creating a configuration similar to that of Figure 11f. The reconstruction at 350-330 Ma indicates that most of the northern boundary of the Mid-Variscan Allochthon came into contact with Avalonia and supports the contention that the Brunovistulian block derives from, or formed part of that ribbon continent.

Subsequent relative dextral displacement of Laurussia and Gondwana was responsible for the late Variscan deformation restored in Figure 11, whose successive steps are approximately reflected in Figure 12 between 330 and 290 Ma. They show that the suture related with the emplacement of the Mid-Variscan Allochthon derives from an ocean situated to the NW, but after strong dextral shearing involving rotational deformation, the root of the suture was moved to a N-S attitude in the east of the Variscan Belt.

Placing the root in the center of the Bohemian Massif makes sense, because: i) this is the only place where part of the Moho may be that of the Allochthon (Fig. 7b); ii) no Middle or Lower Allochthon units have been described (Fig. 6). The latter is what can be expected in the root zone of a large thrust fault, while in the periphery, detached fragments of its relative autochthon can be commonly found as imbricates. The probable continuation of the root zone to the south is an area where the Variscan basement is mostly covered, and probably includes a terrane or terranes different from that of the Mid-Variscan Upper Allochthon.

In that interpretation, the Mediterranean province of the Variscan Belt does not represent an individual terrane, but simply the eastern part of the extended northern Gondwana platform, which recorded a somewhat different evolution (younger Ordovician

magmatism, Sardic unconformity, along margin facies changes). The transition from the Iberian and central Europe province to the Mediterranean province occurs between the two flanks of the Ibero-Armorican Arc (IAA; Fig. 1). But the reconstruction shows that it occurs also between the two sides of the more or less acute angle of the tongue-shaped Mid-Variscan Allochthon (Fig. 11e). The microcontinent whose collision formed the Mid-Variscan Allochthon impinged the northern Gondwana platform at a place where some lateral variations existed previously, but changes are progressive and occurred along hundreds of km: the separation from the Central Iberian Zone under the Cabo Ortegal Complex in NW Iberia and the western Pyrenees was at least of 500 km, and from there to the Montagne Noire another 300 km must be added.

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

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

The space between the Mediterranean province and the Anatolides and Taurides was occupied, at least partially, by Gondwanan continental crust: the Alpine terranes of Figure 12, which were later incorporated to the Alpine chains. These include for instance Adria, Calabria, the Dinaride platform, Greece, and the Peloponnese. According to Matthews et al. (2016), they stayed tied to Gondwana until de Alpine collision, while Stampfli et al. (2013) include them in the continuation of the Variscan Belt to the east, separated at the Middle

Devonian by the opening of the Paleotethys and attached to the south of Laurasia during the Upper Carboniferous.

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

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

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

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

Pereira et al. (2015a, 2017b) interpreted late Variscan (315-280 Ma) calc-alkaline granitoids of Iberia as related to subduction of the Paleotethys linked to the development of the IAA, and unrelated to the Variscan collision. However, the same granitoids have been interpreted as resulting from partial melting of the lower crust with variable mantelic heat and magmatic input and also variable contribution of orogenic crustal thickening (Castiñeiras et al., 2008; Carracedo et al., 2009; Orejana et al., 2012; Villaseca et al., 2017). Even assuming a crustal origin with variable mantle contribution, Gutiérrez-Alonso et al. (2011) defended delamination of thickened mantle lithosphere during the closure of the IAA orocline as the source of heat necessary for massive granite production. And Alcock et al. (2015) calculate that heat production due to crustal thickening could had contributed significantly to heat imput for NW Iberian granite production, without discarding an additional source due to asthenospheric upwelling.

Our model does not back the Paleotethys oceanic subduction hypothesis of Pereira et al. (2015a, 2017b) for late Variscan granitoids, because south of the Iberian Peninsula, the small Paleotethys closed at 320 Ma, and in Matthews et al. (2016) reconstruction, no oceanic crust existed south of Iberia between 325-280 Ma. This does not mean that the model by Pereira et al. (2015a, 2017b) should be discarded, because the reconstruction may be incorrect. Meanwhile, we find that our model of the Variscan Belt implies a late Variscan

narrowing and associated crustal thickening which rather supports the alternative of crustal thickening and melting with variable amounts of mantle involvement.

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

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

10. Conclusions

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

end of the early Variscan stage. Based on that picture, the Mid-Variscan Allochthon is explained by overthrusting of a microcontinent, a peri-Gondwanan terrane detached from a region close to NW Africa and collided with more eastern parts of northern Gondwana platform.

What is preserved of the microcontinent presently forms the Upper Allochthon, but its collision involved the closure of an intervening Mid-Variscan Ocean whose lithosphere partly belongs to the initial (Cambro-Ordovician) Rheic Ocean and partly was created later, during the Silurian and Devonian. Some of these oceanic lithospheres were incorporated to an accretionary prism during closure of the Mid-Variscan Ocean, and were later overthrust onto the northern Gondwana platform, presently forming the Middle Allochthon. Outer parts of that continental platform were also incorporated to the accretionary prism and later emplaced to more internal parts of it, forming the Lower Allochthon and the Parautochthon.

The available knowledge of pre-Variscan and early Variscan evolution and the results of late Variscan retrodeformation have been integrated in a calibrated plate-tectonics reconstruction covering the late Cambrian to early Permian, which is shown with a focus on the Variscan deformation. The resulting model explains the non-cylindrical geometry by: i) the small size of the Upper Allochthon microcontinent compared with the extent of the deformed northern Gondwana and Avalonia plates and ii) large intracontinental dextral shearing caused by the relative eastward displacement of Laurussia relative to Gondwana. The latter also explains the rotation of the Mid-Variscan Allochthon root zone from an original position to the NW of the belt to its proposed present location in the central and southern parts of the Bohemian Massif.

Southward translation and moderate clockwise rotation of Laurussia relative to Gondwana explain indentation leading to formation of the Ibero-Armorican Arc and a hidden arc between Iberia and Morocco. The open Massif Central Arc may be a side effect of the Ibero-Armorican Arc. However, the Central Iberian Arc is primarily related to the indentation of the Mid-Variscan Allochthon, although it was strongly amplified by late-Variscan dextral shearing. The Bohemian Arc is also related to dextral shearing, resolved there in the Elbe Shear Zone and the Moldanubian Thrust.

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Figure captions

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

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

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

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

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

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

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

Fig. 5. Cross sections representative of the structure of the French Variscan massifs. **(a)** Armorican section A-A"', including the Mid-Variscan suture in the southern part. **(b)** Western Massif Central section B-B'. **(c)** Eastern Massif Central section C-C". The sections are built from field geology, based on the Geological map of France to scale 1:1 000 000 (BRGM, 1996), and on geological sections by Ledru et al. (1994b), Faure et al. (2009), Ballèvre et al. (2014), and Lardeaux et al. (2014). Section A-A"'follows the trace of seismic reflection profiles ARMOR 2 South and 2 North (Bitri et al., 2003, 2010), AR2, AR1 and SWAT 10 (Bitri et al., 2001), whose interpretations have been included. For the Central Armorican Domain, reflections from the Brittany Wide Angle Refection Profile (BWARF; Matte and Hirn, 1988) separated c. 100 km to the east, have been projected into the section. Whrench faults are continued at depth avoiding to cross low-dipping reflections. Colors corresponding to zones or groups are superposed to frames or grey shades indicating lithologies or units. **Armorican Domains (AD):** CAD, Central; NAD, North; SAD, South. **Shear zones (SZ)** and **faults (F):** CF, Combrailles; GCF, Granville-Cancale; NASZ, North Armorican; NEF, Nort-sur-Erdre; OF, Ouzilly; SASZ, South Armorican (N and S, northern and southern branches); SLF, South Limousin. **Other:** FMC, French Massif Central; MN, Montagne Noire. Upper Allochthon: IP, Intermediate-P; HP-HT, High-P/High-T. See Figure 4 for location of the sections.

Fig. 6. Geological map of the Bohemian Massif. Based on the Geological maps of Germany to scale 1:1 000 000 (BGR, 1993), and the Czech Republic to scale 1:500 000 (Czech

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

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

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

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

Fig. 9. Correlation of zones, groups and units and of early Variscan and Variscan orogenic magmatism along the Variscan Belt. Only the main massifs are included, while small ones can generally be assimilated to nearby large ones. Colors and abbreviations are the same used in maps and sections.

Fig. 10. Timing of early Variscan and Variscan events of the large massifs. The time interval spans 425-275 Ma. Small massifs are assumed to have similar evolutions as the massifs to which they are related in the text, and many of them are included in the figure. Bar colors and abbreviations are the same used in the maps, sections and text. The events related to subduction of the Rheic and Rhenohercynian oceans are given separated from those considered related with the building of the Mid-Variscan accretionary complex forming the Allochthon, their emplacement on the northern Gondwana Autochthon and the subsequent evolution of the ensemble.

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

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

Fig. 12. Evolution of the Variscan Belt in the context of Paleozoic plate dynamics. The main continental masses and the elements forming the Variscan Belt are shown in orthographic projection for different ages. Northern and eastern territories (Siberia, China, Indonesia) are omitted. For the late Cambrian and Lower Ordovician, a projection centered in the South Pole has been used. The geographic grid is drawn every 15º, with a thicker line for the Equator, and is fixed for the 500-470 and 400-290 Ma intervals. This figure was built using animations 500.A97 and 470.A97 of GMap2015, the reconstruction geometries of Matthews et al. (2016) for GPlates, and the partial retrodeformation carried out in this work (Fig. 11). Blue lines indicate the location of sections in Figure 13 (dotted parts not shown). Note that the Mid-Variscan Allochthon represents the future Upper Allochthon until 400 Ma. Between 390 and 360 Ma it progressively incorporated the Middle and Lower Allochthons and by 340 Ma also the Parautochthon.

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

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

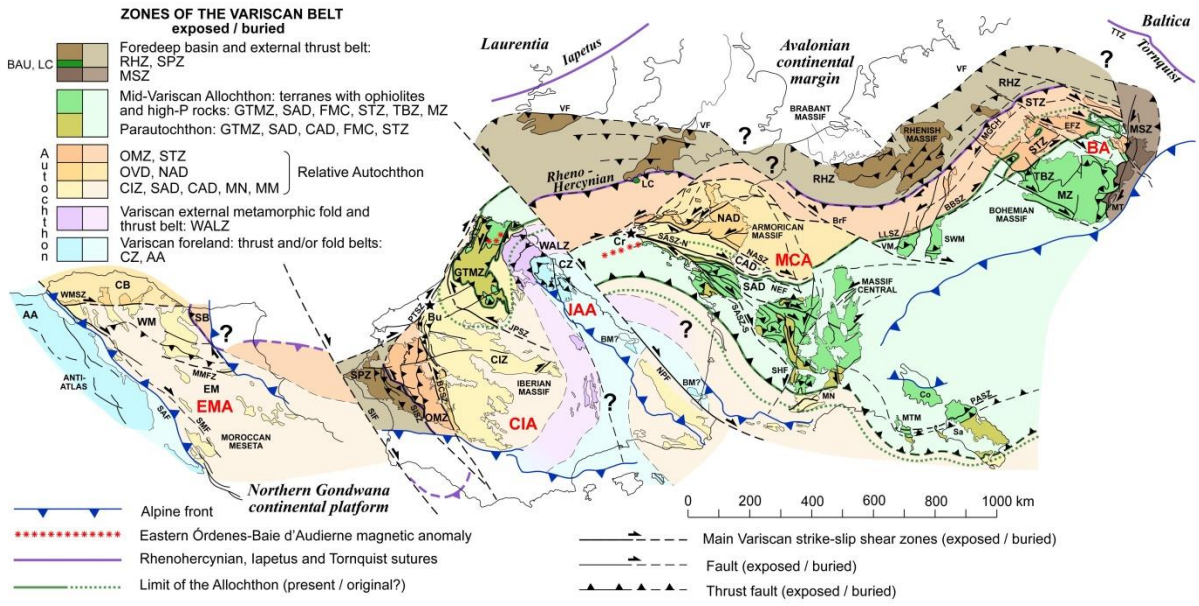
ZONES OF THE VARISCAN BELT
exposed / buried
BAU, LC
Foredeep basin and external thrust belt:
RHZ, SPZ
MSZ
Mid-Variscan Allochthon: terranes with ophiolites and high-P rocks: GTMZ, SAD, FMC, STZ, TBZ, MZ
Parautochthon: GTMZ, SAD, CAD, FMC, STZ
Autochthon
OMZ, STZ
OVD, NAD
CIZ, SAD, CAD, MN, MM
Relative Autochthon
Variscan external metamorphic fold and thrust belt: WALZ
Variscan foreland: thrust and/or fold belts: CZ, AA
Laurentia
Iapetus
Avalonian continental margin
Baltica
Tornquist
Rheno-Hercynian
EMA
MCA
JAA
CIA
Northern Gondwana continental platform
MOROCCAN MESETA
IBERIAN MASSIF
ARMORICAN MASSIF
BOHEMIAN MASSIF
MASSIF CENTRAL
RHENISH MASSIF
BRABANT MASSIF
0 200 400 600 800 1000 km
Alpine front
Eastern Ördenes-Baie d'Audierne magnetic anomaly
Rhenohercynian, Iapetus and Tornquist sutures
Limit of the Allochthon (present / original?)
Main Variscan strike-slip shear zones (exposed / buried)
Fault (exposed / buried)
Thrust fault (exposed / buried)

Fig. 1

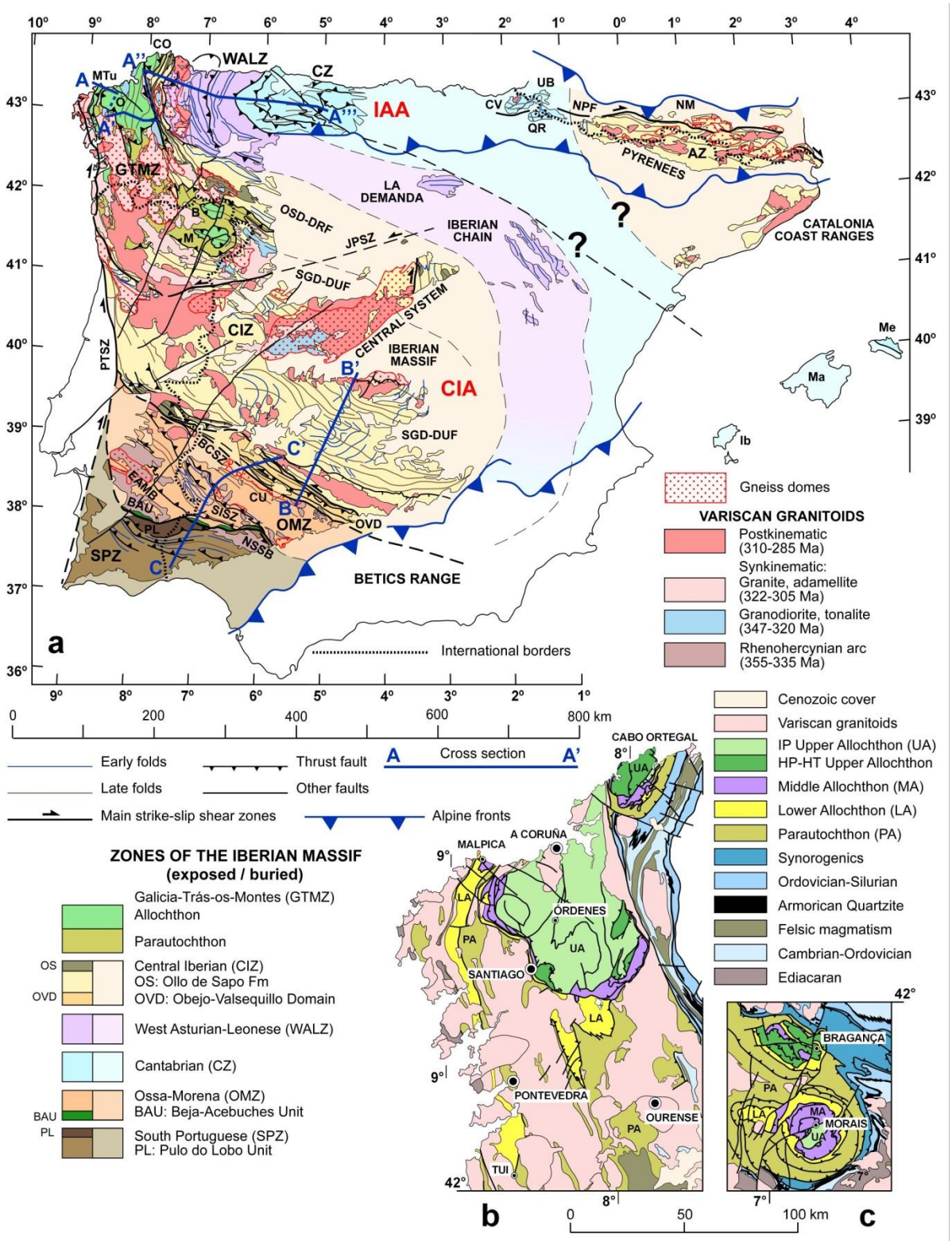
CO
WALZ
CZ
A
A''
A'
A'''
MTu
IAA
UB
CV
QR
NPF
NM
PYRENEES
AZ
GTMZ
LA DEMANDA
OSD-DRF
JPSZ
IBERIAN CHAIN
CATALONIA COAST RANGES
SGD-DUF
CENTRAL SYSTEM
CIZ
PTSZ
IBERIAN MASSIF
CIA
B'
C'
BCSZ
EAMB
BAU
PL
SISZ
CU
B
OMZ
OVD
NSSB
SPZ
C
BETICS RANGE
Me
Ma
Ib
a
International borders
Gneiss domes
VARISCAN GRANITOIDS
Postkinematic (310-285 Ma)
Synkinematic: Granite, adamellite (322-305 Ma)
Granodiorite, tonalite (347-320 Ma)
Rhenohercynian arc (355-335 Ma)
0 200 400 600 800 km
Early folds
Late folds
Main strike-slip shear zones
Thrust fault
Other faults
Alpine fronts
A Cross section A'
ZONES OF THE IBERIAN MASSIF
(exposed / buried)
Galicia-Trás-os-Montes (GTMZ)
Allochthon
Parautochthon
OS
OVD
Central Iberian (CIZ)
OS: Ollo de Sapo Fm
OVD: Obejo-Valsequillo Domain
West Asturian-Leonese (WALZ)
Cantabrian (CZ)
Ossa-Morena (OMZ)
BAU: Beja-Acebuches Unit
BAU
PL
South Portuguese (SPZ)
PL: Pulo do Lobo Unit
Cenozoic cover
Variscan granitoids
IP Upper Allochthon (UA)
HP-HT Upper Allochthon
Middle Allochthon (MA)
Lower Allochthon (LA)
Parautochthon (PA)
Synorogenics
Ordovician-Silurian
Armorican Quartzite
Felsic magmatism
Cambrian-Ordovician
Ediacaran
CABO ORTEGAL
A CORUÑA
MALPICA
ORDENES
SANTIAGO
PONTEVEDRA
OURENSE
TUI
BRAGANÇA
MORAIS
b
c
0 50 100 km

Fig. 2

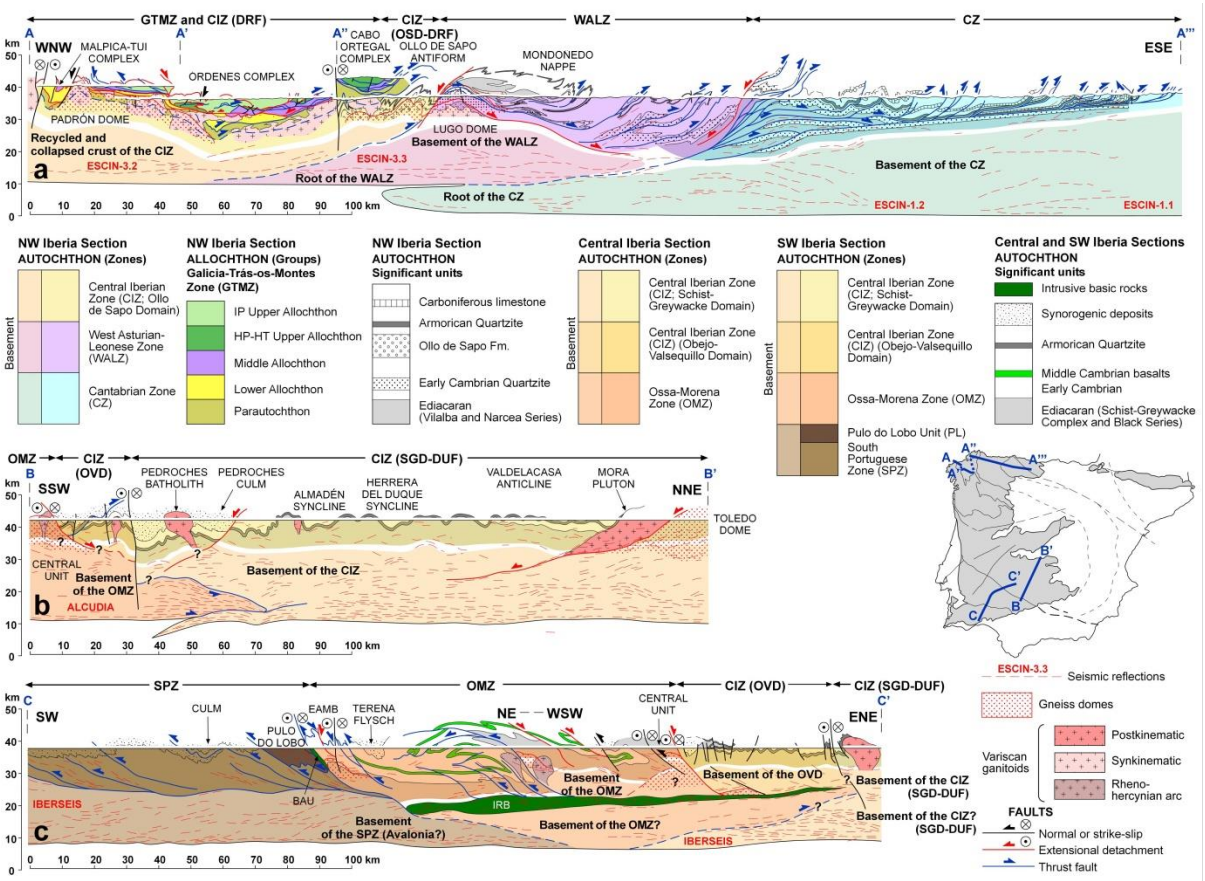
GTMZ and CIZ (DRF)
CIZ
WALZ
CZ
WNW
ESE
MALPICA-TUI COMPLEX
ORDENES COMPLEX
CABO ORTEGAL COMPLEX
(OSD-DRF)
OLLO DE SAPO ANTIFORM
MONDONEDO NAPPE
PADRÓN DOME
LUGO DOME
Recycled and collapsed crust of the CIZ
ESCIN-3.2
ESCIN-3.3
Basement of the WALZ
Root of the WALZ
Basement of the CZ
Root of the CZ
ESCIN-1.2
ESCIN-1.1
a
NW Iberia Section
AUTOCHTHON (Zones)
Central Iberian Zone (CIZ; Ollo de Sapo Domain)
West Asturian-Leonese Zone (WALZ)
Cantabrian Zone (CZ)
NW Iberia Section
ALLOCHTHON (Groups)
Galicia-Trás-os-Montes Zone (GTMZ)
IP Upper Allochthon
HP-HT Upper Allochthon
Middle Allochthon
Lower Allochthon
Parautochthon
NW Iberia Section
AUTOCHTHON
Significant units
Carboniferous limestone
Armorican Quartzite
Ollo de Sapo Fm.
Early Cambrian Quartzite
Ediacaran (Vilalba and Narcea Series)
Central Iberia Section
AUTOCHTHON (Zones)
Central Iberian Zone (CIZ; Schist-Greywacke Domain)
Central Iberian Zone (CIZ) (Obejo-Valsequillo Domain)
Ossa-Morena Zone (OMZ)
SW Iberia Section
AUTOCHTHON (Zones)
Central Iberian Zone (CIZ; Schist-Greywacke Domain)
Central Iberian Zone (CIZ) (Obejo-Valsequillo Domain)
Ossa-Morena Zone (OMZ)
Pulo do Lobo Unit (PL)
South Portuguese Zone (SPZ)
Central and SW Iberia Sections
AUTOCHTHON
Significant units
Intrusive basic rocks
Synorogenic deposits
Armorican Quartzite
Middle Cambrian basalts
Early Cambrian
Ediacaran (Schist-Greywacke Complex and Black Series)
Basement
OMZ
CIZ (OVD)
CIZ (SGD-DUF)
SSW
NNE
PEDROCHES BATHOLITH
PEDROCHES CULM
ALMADÉN SYNCLINE
HERRERA DEL DUQUE SYNCLINE
VALDELACASA ANTICLINE
MORA PLUTON
TOLEDO DOME
CENTRAL UNIT
Basement of the OMZ
ALCUDIA
Basement of the CIZ
b
SPZ
OMZ
CIZ (OVD)
CIZ (SGD-DUF)
SW
NE
WSW
ENE
CULM
PULO DO LOBO
EAMB
TERENA FLYSCH
CENTRAL UNIT
IBERSEIS
BAU
IRB
Basement of the OMZ
Basement of the OVD
Basement of the CIZ (SGD-DUF)
Basement of the SPZ (Avalonia?)
Basement of the OMZ?
Basement of the CIZ? (SGD-DUF)
c
ESCIN-3.3
Seismic reflections
Gneiss domes
Variscan ganitoids
Postkinematic
Synkinematic
Rheno-hercynian arc
FAULTS
Normal or strike-slip
Extensional detachment
Thrust fault

Fig. 3

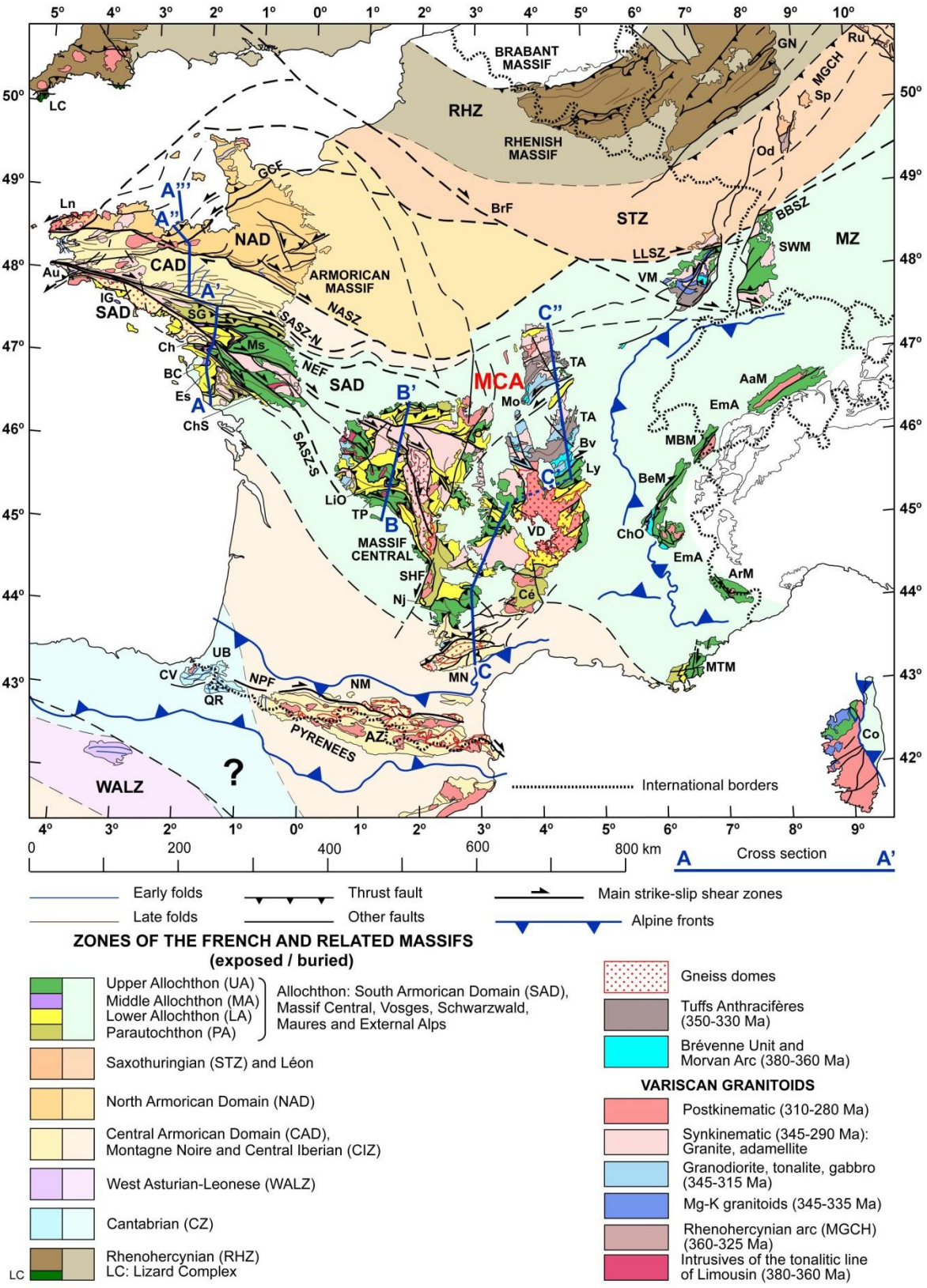
BRABANT MASSIF
RHENISH MASSIF
RHZ
STZ
MZ
NAD
CAD
SAD
ARMORICAN MASSIF
MASSIF CENTRAL
MCA
WALZ
PYRENEES
AZ
GN
Ru
MGCH
Sp
Od
BBSZ
SWM
LLSZ
VM
GGF
BrF
NASZ
SASZ-N
SASZ-S
NEF
SHF
NPF
NM
LC
Ln
Au
IG
SG
Ch
Ms
BC
Es
ChS
LiO
TP
VD
Nj
Cé
MN
TA
Mo
Bv
Ly
AaM
EmA
MBM
BeM
ChO
ArM
MTM
Co
UB
CV
QR
A'''
A''
A'
A
B'
B
C''
C'
C
?
4° 3° 2° 1° 0° 1° 2° 3° 4° 5° 6° 7° 8° 9°
50° 49° 48° 47° 46° 45° 44° 43° 42°
0 200 400 600 800 km
International borders
A Cross section A'
Early folds
Late folds
Thrust fault
Other faults
Main strike-slip shear zones
Alpine fronts
ZONES OF THE FRENCH AND RELATED MASSIFS
(exposed / buried)
Upper Allochthon (UA)
Middle Allochthon (MA)
Lower Allochthon (LA)
Parautochthon (PA)
Allochthon: South Armorican Domain (SAD), Massif Central, Vosges, Schwarzwald, Maures and External Alps
Saxothuringian (STZ) and Léon
North Armorican Domain (NAD)
Central Armorican Domain (CAD), Montagne Noire and Central Iberian (CIZ)
West Asturian-Leonese (WALZ)
Cantabrian (CZ)
Rhenohercynian (RHZ)
LC: Lizard Complex
Gneiss domes
Tuffs Anthracifères (350-330 Ma)
Brévenne Unit and Morvan Arc (380-360 Ma)
VARISCAN GRANITOIDS
Postkinematic (310-280 Ma)
Synkinematic (345-290 Ma): Granite, adamellite
Granodiorite, tonalite, gabbro (345-315 Ma)
Mg-K granitoids (345-335 Ma)
Rhenohercynian arc (MGCH) (360-325 Ma)
Intrusives of the tonalitic line of Limousin (380-360 Ma)

Fig. 4

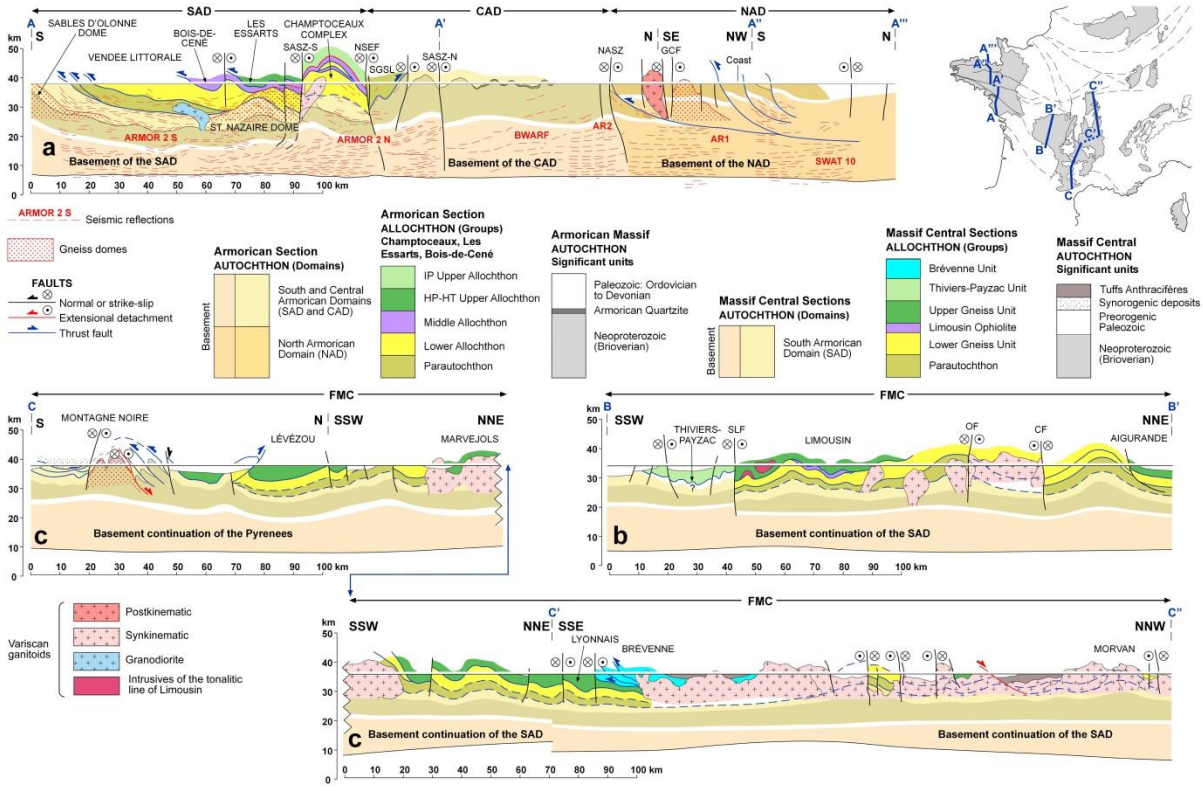
SAD
CAD
NAD
SABLES D'OLONNE DOME
VENDEE LITTORALE
BOIS-DE-CENE
LES ESSARTS
CHAMPTOCEAUX COMPLEX
SASZ-S
NSEF
SGSL
SASZ-N
NASZ
GCF
Coast
ST. NAZAIRE DOME
ARMOR 2 S
ARMOR 2 N
BWARF
AR2
AR1
SWAT 10
Basement of the SAD
Basement of the CAD
Basement of the NAD
a
Seismic reflections
Gneiss domes
FAULTS
Normal or strike-slip
Extensional detachment
Thrust fault
Armorican Section
AUTOCHTHON (Domains)
Basement
South and Central Armorican Domains (SAD and CAD)
North Armorican Domain (NAD)
Armorican Section
ALLOCHTHON (Groups)
Champtoceaux, Les Essarts, Bois-de-Cené
IP Upper Allochthon
HP-HT Upper Allochthon
Middle Allochthon
Lower Allochthon
Parautochthon
Armorican Massif
AUTOCHTHON
Significant units
Paleozoic: Ordovician to Devonian
Armorican Quartzite
Neoproterozoic (Brioverian)
Massif Central Sections
AUTOCHTHON (Domains)
South Armorican Domain (SAD)
Massif Central Sections
ALLOCHTHON (Groups)
Brévenne Unit
Thiviers-Payzac Unit
Upper Gneiss Unit
Limousin Ophiolite
Lower Gneiss Unit
Parautochthon
Massif Central
AUTOCHTHON
Significant units
Tuffs Anthracifères
Synorogenic deposits
Preorogenic Paleozoic
Neoproterozoic (Brioverian)
FMC
MONTAGNE NOIRE
LÉVÉZOU
MARVEJOLS
Basement continuation of the Pyrenees
c
THIVIERS-PAYZAC
SLF
LIMOUSIN
OF
CF
AIGURANDE
Basement continuation of the SAD
b
LYONNAIS
BRÉVENNE
MORVAN
Basement continuation of the SAD
Basement continuation of the SAD
Variscan ganitoids
Postkinematic
Synkinematic
Granodiorite
Intrusives of the tonalitic line of Limousin

Fig. 5

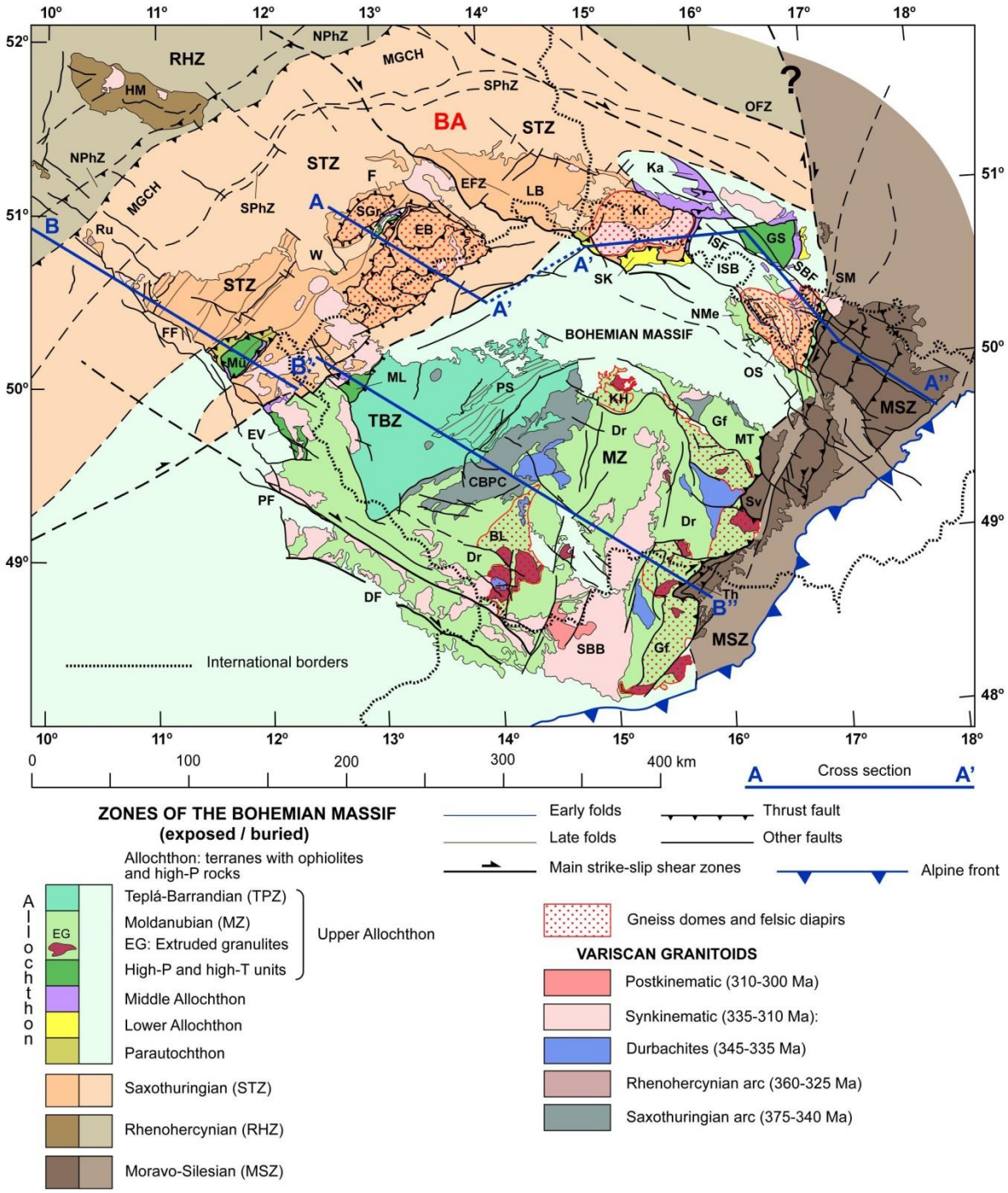

Fig. 6

6929

6930

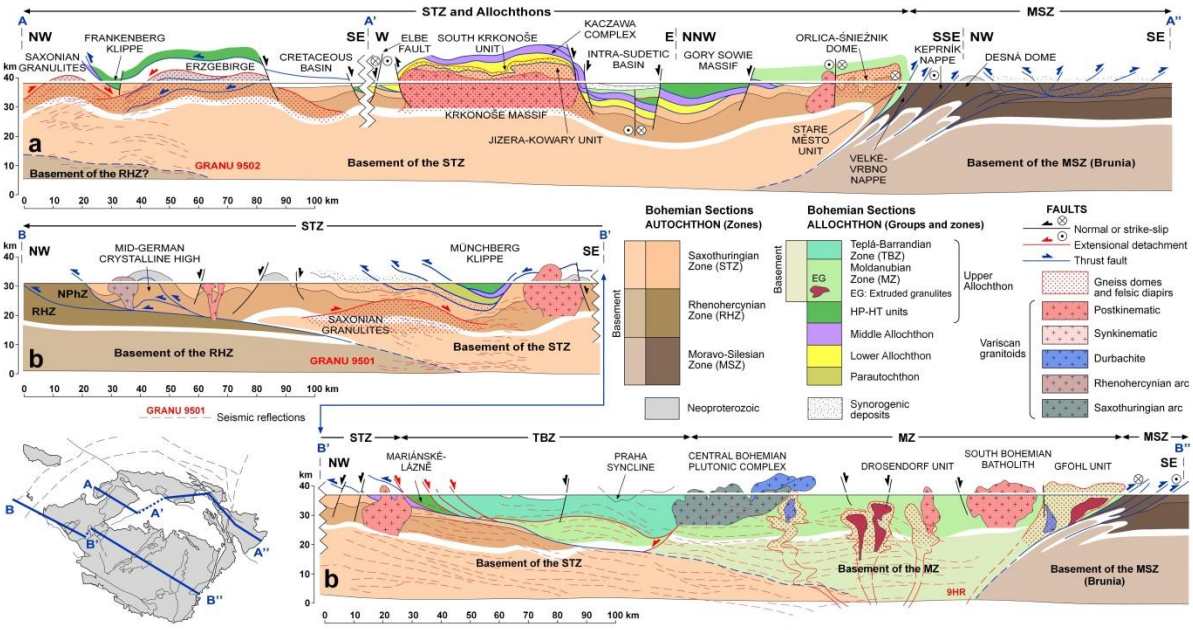
STZ and Allochthons
MSZ
NW
FRANKENBERG KLIPPE
SAXONIAN GRANULITES
ERZGEBIRGE
CRETACEOUS BASIN
SE
W
ELBE FAULT
SOUTH KRKONOŠE UNIT
KACZAWA COMPLEX
E
NNW
INTRA-SUDETIC BASIN
GORY SOWIE MASSIF
ORLICA-ŚNIEŻNIK DOME
SSE
NW
KEPRNÍK NAPPE
DESNÁ DOME
SE
KRKONOŠE MASSIF
JIZERA-KOWARY UNIT
STARÉ MĚSTO UNIT
VELKÉ-VRBNO NAPPE
Basement of the RHZ?
GRANU 9502
Basement of the STZ
Basement of the MSZ (Brunia)
a
km
STZ
NW
MID-GERMAN CRYSTALLINE HIGH
MÜNCHBERG KLIPPE
SE
NPhZ
RHZ
SAXONIAN GRANULITES
Basement of the RHZ
GRANU 9501
Basement of the STZ
b
GRANU 9501
Seismic reflections
Bohemian Sections
AUTOCHTHON (Zones)
Saxothuringian Zone (STZ)
Rhenohercynian Zone (RHZ)
Moravo-Silesian Zone (MSZ)
Neoproterozoic
Basement
Bohemian Sections
ALLOCHTHON (Groups and zones)
Teplá-Barrandian Zone (TBZ)
Moldanubian Zone (MZ)
EG: Extruded granulites
HP-HT units
Middle Allochthon
Lower Allochthon
Parautochthon
Synorogenic deposits
Upper Allochthon
Variscan granitoids
FAULTS
Normal or strike-slip
Extensional detachment
Thrust fault
Gneiss domes and felsic diapirs
Postkinematic
Synkinematic
Durbachite
Rhenohercynian arc
Saxothuringian arc
STZ
TBZ
MZ
MSZ
NW
MARIÁNSKÉ-LÁZNĚ
PRAHA SYNCLINE
CENTRAL BOHEMIAN PLUTONIC COMPLEX
DROSENDORF UNIT
SOUTH BOHEMIAN BATHOLITH
GFOHL UNIT
SE
Basement of the STZ
Basement of the MZ
9HR
Basement of the MSZ (Brunia)
b

Fig. 7

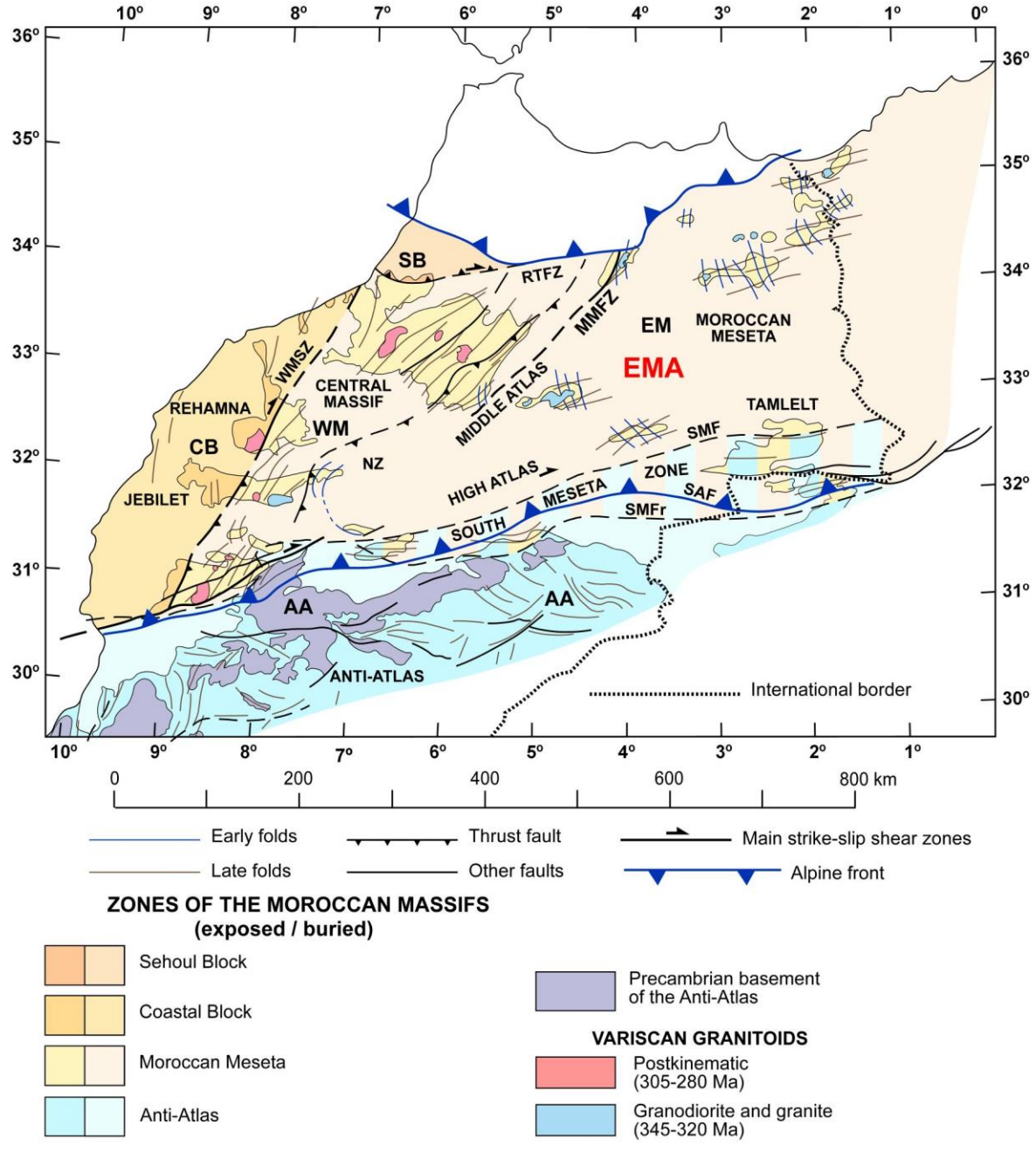
10°
9°
8°
7°
6°
5°
4°
3°
2°
1°
0°
36°
35°
34°
33°
32°
31°
30°
SB
RTFZ
MMFZ
EM
MOROCCAN MESETA
EMA
WMSZ
CENTRAL MASSIF
REHAMNA
CB
WM
NZ
MIDDLE ATLAS
TAMLELT
SMF
HIGH ATLAS
MESETA
ZONE
SAF
SMFr
SOUTH
JEBILET
AA
AA
ANTI-ATLAS
International border
0
200
400
600
800 km
Early folds
Late folds
Thrust fault
Other faults
Main strike-slip shear zones
Alpine front
ZONES OF THE MOROCCAN MASSIFS
(exposed / buried)
Sehoul Block
Coastal Block
Moroccan Meseta
Anti-Atlas
Precambrian basement of the Anti-Atlas
VARISCAN GRANITOIDS
Postkinematic (305-280 Ma)
Granodiorite and granite (345-320 Ma)

Fig. 8

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

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

Fig. 9

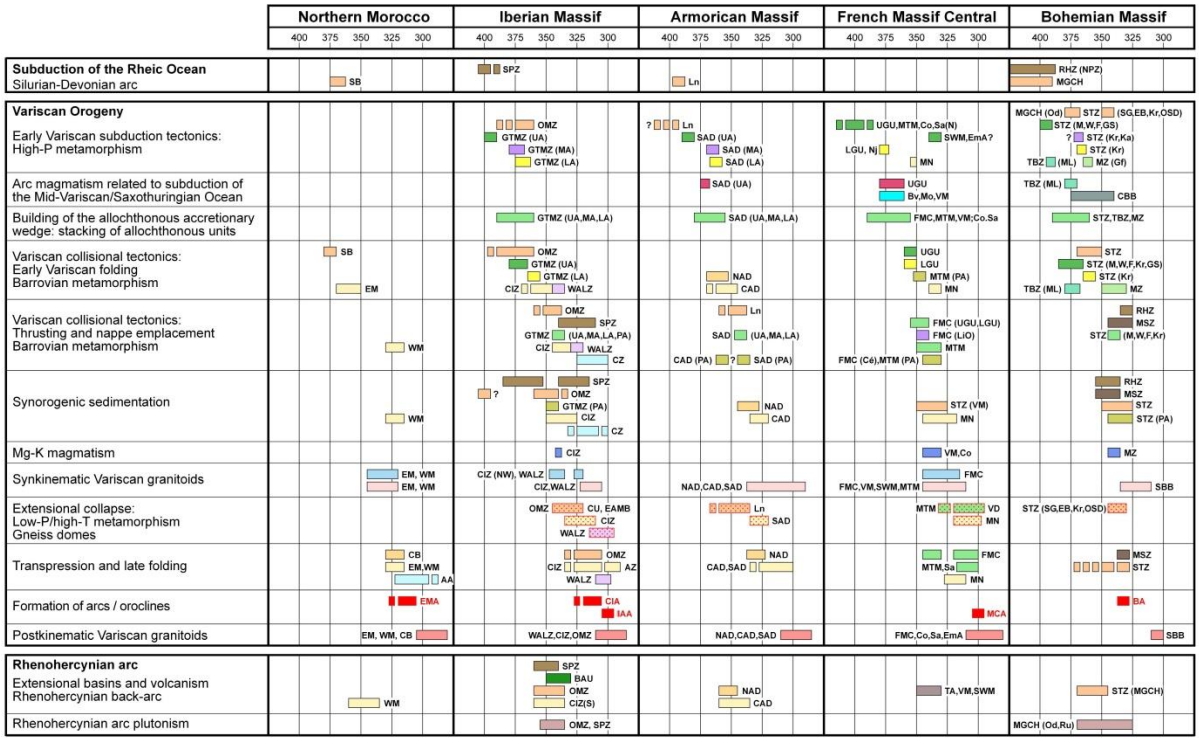
Northern Morocco
Iberian Massif
Armorican Massif
French Massif Central
Bohemian Massif
400 375 350 325 300
Subduction of the Rheic Ocean
Silurian-Devonian arc
SB
SPZ
Ln
RHZ (NPZ)
MGCH
Variscan Orogeny
Early Variscan subduction tectonics:
High-P metamorphism
OMZ
GTMZ (UA)
GTMZ (MA)
GTMZ (LA)
?
Ln
SAD (UA)
SAD (MA)
SAD (LA)
UGU,MTM,Co,Sa(N)
SWM,EmA?
LGU, Nj
MN
MGCH (Od)
STZ
(SG,EB,Kr,OSD)
STZ (M,W,F,GS)
?
STZ (Kr,Ka)
STZ (Kr)
TBZ
(ML)
MZ (Gf)
Arc magmatism related to subduction of
the Mid-Variscan/Saxothuringian Ocean
SAD (UA)
UGU
Bv,Mo,VM
TBZ (ML)
CBB
Building of the allochthonous accretionary
wedge: stacking of allochthonous units
GTMZ (UA,MA,LA)
SAD (UA,MA,LA)
FMC,MTM,VM;Co.Sa
STZ,TBZ,MZ
Variscan collisional tectonics:
Early Variscan folding
Barrovian metamorphism
SB
EM
OMZ
GTMZ (UA)
GTMZ (LA)
CIZ
WALZ
NAD
CAD
UGU
LGU
MTM (PA)
MN
STZ
STZ (M,W,F,Kr,GS)
STZ (Kr)
TBZ (ML)
MZ
Variscan collisional tectonics:
Thrusting and nappe emplacement
Barrovian metamorphism
WM
OMZ
SPZ
GTMZ
(UA,MA,LA,PA)
CIZ
WALZ
CZ
Ln
SAD
(UA,MA,LA)
CAD (PA)
?
SAD (PA)
FMC (UGU,LGU)
FMC (LiO)
MTM
FMC (Cé),MTM (PA)
RHZ
MSZ
STZ
(M,W,F,Kr)
Synorogenic sedimentation
WM
SPZ
?
OMZ
GTMZ (PA)
CIZ
CZ
NAD
CAD
STZ (VM)
MN
RHZ
MSZ
STZ
STZ (PA)
Mg-K magmatism
CIZ
VM,Co
MZ
Synkinematic Variscan granitoids
EM, WM
EM, WM
CIZ (NW), WALZ
CIZ,WALZ
NAD,CAD,SAD
FMC
FMC,VM,SWM,MTM
SBB
Extensional collapse:
Low-P/high-T metamorphism
Gneiss domes
OMZ
CU, EAMB
CIZ
WALZ
Ln
SAD
MTM
VD
MN
STZ (SG,EB,Kr,OSD)
Transpression and late folding
CB
EM,WM
AA
OMZ
CIZ
AZ
WALZ
NAD
CAD,SAD
FMC
MTM,Sa
MN
MSZ
STZ
Formation of arcs / oroclines
EMA
CIA
IAA
MCA
BA
Postkinematic Variscan granitoids
EM, WM, CB
WALZ,CIZ,OMZ
NAD,CAD,SAD
FMC,Co,Sa,EmA
SBB
Rhenohercynian arc
Extensional basins and volcanism
Rhenohercynian back-arc
WM
SPZ
BAU
OMZ
CIZ(S)
NAD
CAD
TA,VM,SWM
STZ (MGCH)
Rhenohercynian arc plutonism
OMZ, SPZ
MGCH (Od,Ru)

Fig. 10

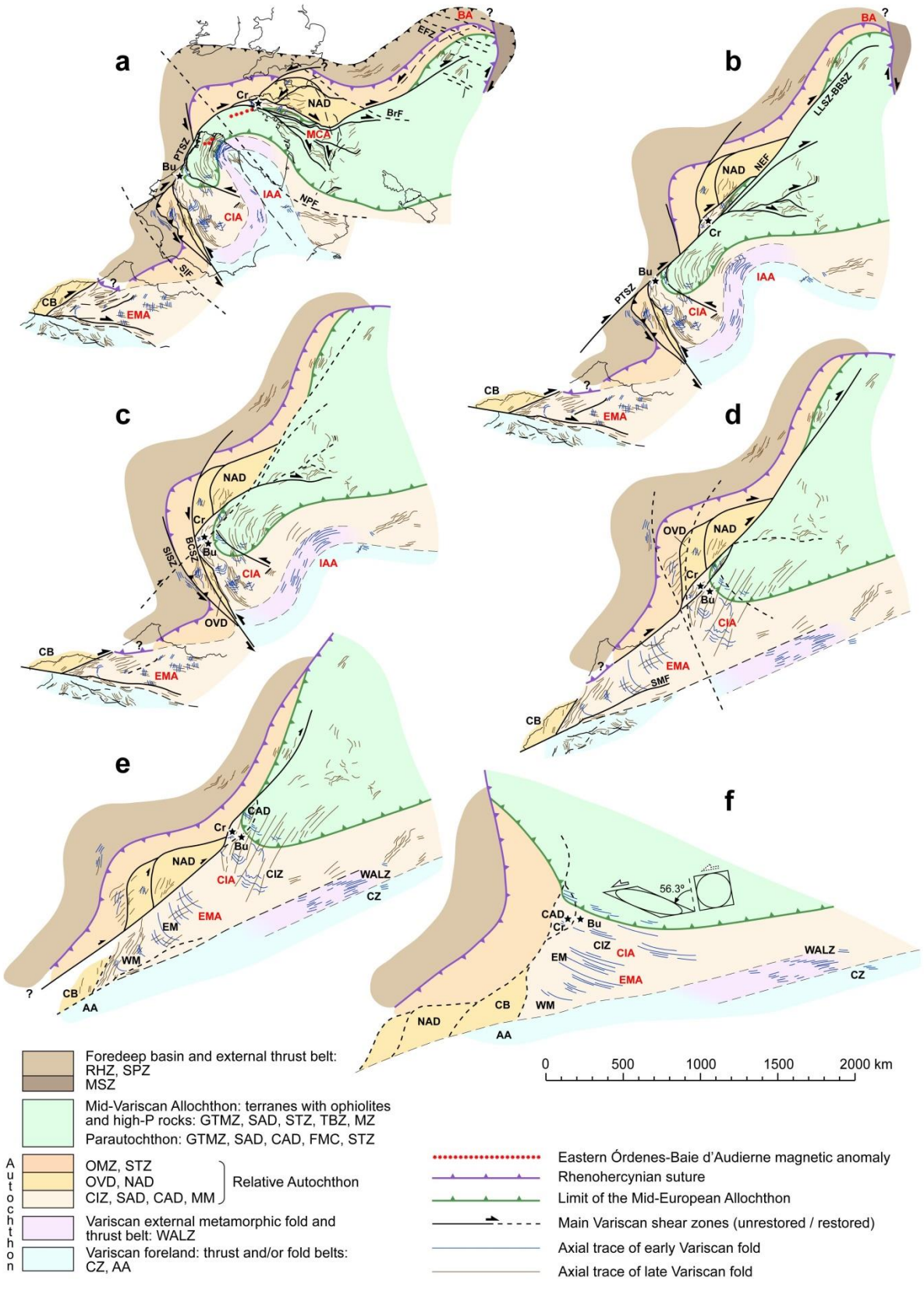
a
b
c
d
e
f
BA
EFZ
Cr
NAD
BrF
MCA
PTSZ
Bu
IAA
NPE
CIA
SIF
CB
EMA
LLSZ-BBSZ
NEF
SISZ
BCSZ
OVD
SMF
CAD
CIZ
WALZ
CZ
EM
WM
AA
56.3°
0
500
1000
1500
2000 km
Foredeep basin and external thrust belt: RHZ, SPZ MSZ
Mid-Variscan Allochthon: terranes with ophiolites and high-P rocks: GTMZ, SAD, STZ, TBZ, MZ
Parautochthon: GTMZ, SAD, CAD, FMC, STZ
Autochthon
OMZ, STZ
OVD, NAD
CIZ, SAD, CAD, MM
Relative Autochthon
Variscan external metamorphic fold and thrust belt: WALZ
Variscan foreland: thrust and/or fold belts: CZ, AA
Eastern Órdenes-Baie d'Audierne magnetic anomaly
Rhenohercynian suture
Limit of the Mid-European Allochthon
Main Variscan shear zones (unrestored / restored)
Axial trace of early Variscan fold
Axial trace of late Variscan fold

Fig. 11

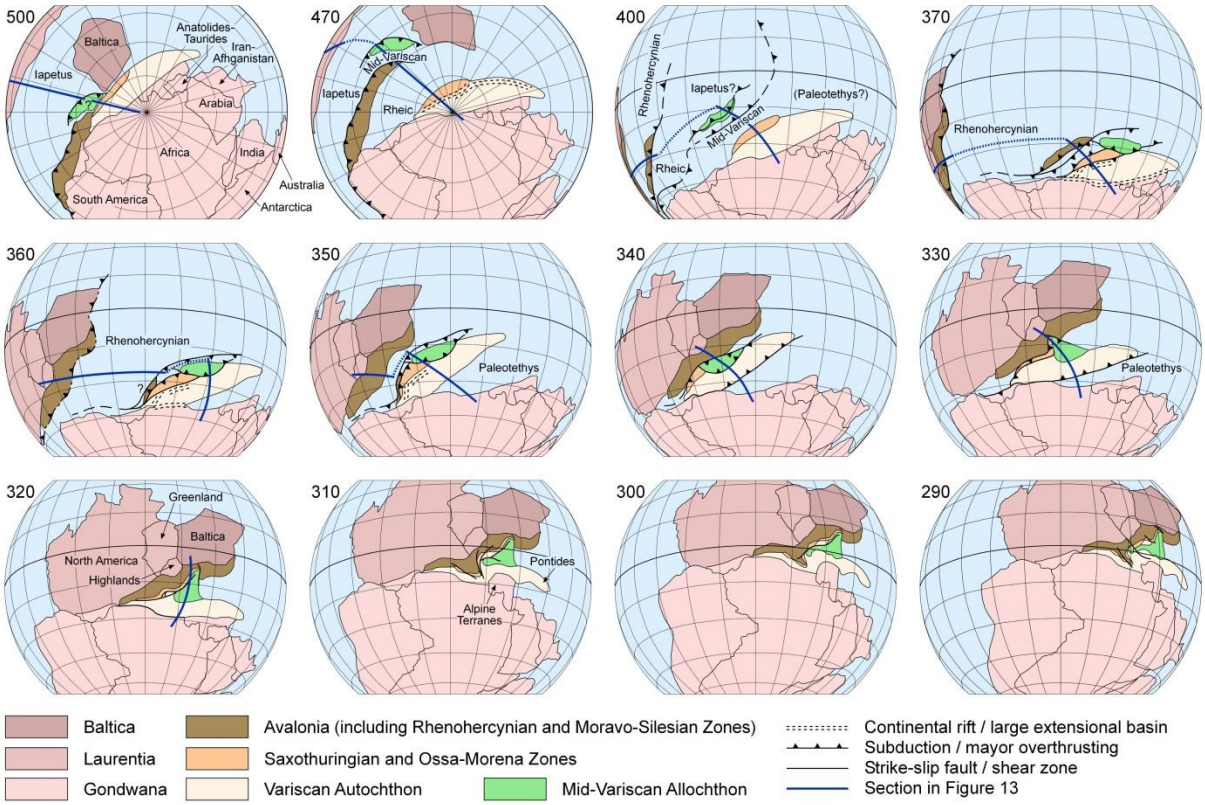
500
Baltica
Anatolides-
Taurides
Iran-
Afhganistan
Iapetus
Arabia
Africa
India
Australia
South America
Antarctica
470
Mid-Variscan
Iapetus
Rheic
400
Rhenohercynian
Iapetus?
(Paleotethys?)
Mid-Variscan
Rheic
370
Rhenohercynian
360
Rhenohercynian
350
Paleotethys
340
330
Paleotethys
320
Greenland
Baltica
North America
Highlands
310
Pontides
Alpine
Terranes
300
290
Baltica
Laurentia
Gondwana
Avalonia (including Rhenohercynian and Moravo-Silesian Zones)
Saxothuringian and Ossa-Morena Zones
Variscan Autochthon
Mid-Variscan Allochthon
Continental rift / large extensional basin
Subduction / mayor overthrusting
Strike-slip fault / shear zone
Section in Figure 13

Fig. 12

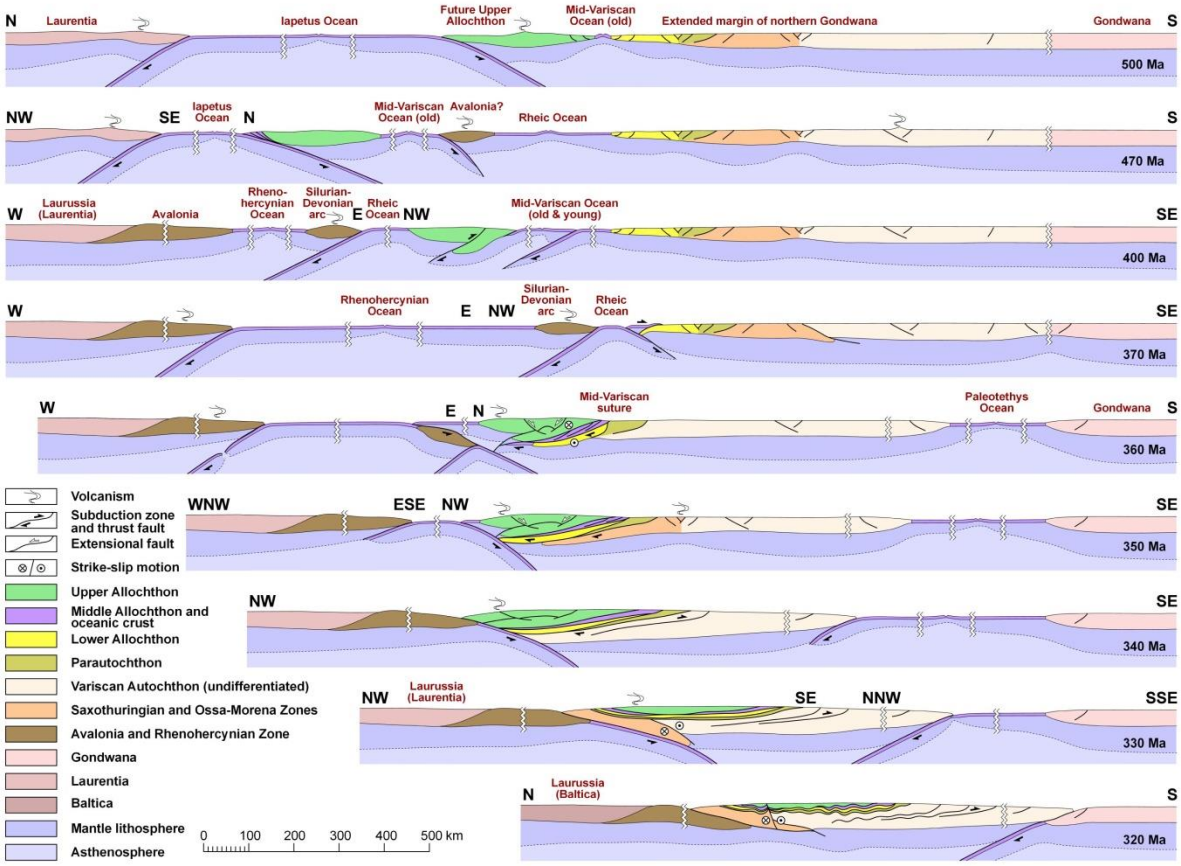
N
Laurentia
Iapetus Ocean
Future Upper Allochthon
Mid-Variscan Ocean (old)
Extended margin of northern Gondwana
Gondwana
S
500 Ma
NW
SE
Iapetus Ocean
N
Mid-Variscan Ocean (old)
Avalonia?
Rheic Ocean
S
470 Ma
W
Laurussia (Laurentia)
Avalonia
Rheno-hercynian Ocean
Silurian-Devonian arc
E
Rheic Ocean
NW
Mid-Variscan Ocean (old & young)
SE
400 Ma
W
Rhenohercynian Ocean
E
NW
Silurian-Devonian arc
Rheic Ocean
SE
370 Ma
W
E
N
Mid-Variscan suture
Paleotethys Ocean
Gondwana
S
360 Ma
Volcanism
Subduction zone and thrust fault
Extensional fault
Strike-slip motion
Upper Allochthon
Middle Allochthon and oceanic crust
Lower Allochthon
Parautochthon
Variscan Autochthon (undifferentiated)
Saxothuringian and Ossa-Morena Zones
Avalonia and Rhenohercynian Zone
Gondwana
Laurentia
Baltica
Mantle lithosphere
Asthenosphere
0
100
200
300
400
500 km
WNW
ESE
NW
SE
350 Ma
NW
SE
340 Ma
NW
Laurussia (Laurentia)
SE
NNW
SSE
330 Ma
N
Laurussia (Baltica)
S
320 Ma

Fig. 13

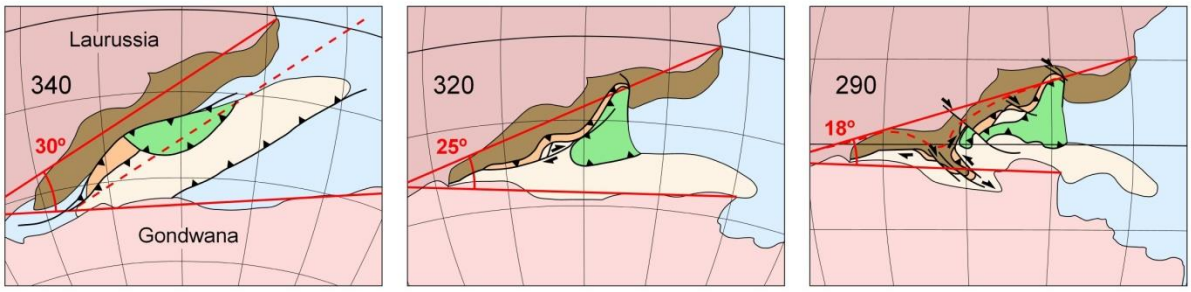

Fig. 14

Abbreviations

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

AA	Anti-Atlas (Morocco)
AaM	Aar Massif (Alps)
ArM	Argentera Massif (Alps)
Au	Audierne Bay Complex (Armorican Massif)
AZ	Axial Zone (Pyrenees)
B	Bragança (NW Iberia allochthonous complex)
BA	Bohemian Arc
BAU	Beja-Acebuches Unit (Iberian Massif)
BBSZ	Baden-Baden Shear Zone (Schwarzwald Massif)
BC	Bois-de-Cené Unit (Armorican Massif)
BCSZ	Badajoz-Córdoba Shear Zone (Iberian Massif)
BeM	Belledonne Massif (Alps)
BL	Blanský les Unit (Bohemian Massif)
BM	Basque Massifs
BrF	Bray Fault
Bu	Buçaco (Portugal)
Bv	Brévenne Unit (French Massif Central)
CAD	Central Armorican Domain
CB	Coastal Block (Morocco)
CBPC	Central Bohemian Plutonic Complex
Cé	Cévennes (French Massif Central)
CF	Combrailles Fault (French Massif Central)
Ch	Champtoceaux Complex (Armorican Massif)
ChO	Chamrousse Ophiolite (Alps)
ChS	Chantonnay Syncline (Armorican Massif)
CIA	Central Iberian Arc
CIZ	Central Iberian Zone
Co	Corsica
CO	Cabo Ortegal (NW Iberia allochthonous complex)
Cr	Crozon (Armorican Massif)
CU	Central Unit (Iberian Massif)
CV	Cinco Villas (Basque Massif, Pyrenees)
CZ	Cantabrian Zone (Iberian Massif)
DF	Danube Fault (Bohemian Massif)
Dr	Drosendorf Unit (Bohemian Massif)
DRF	Domain of Recumbent Folds (Iberian Massif)
DUF	Domain of Upright Folds (Iberian Massif)
EAMB	Évora-Aracena Metamorphic Belt (Iberian Massif)
EB	Erzgebirge Massif (Bohemian Massif)
EFZ	Elbe Fault Zone (Bohemian Massif)
EM	Eastern Meseta (Morocco)
EmA	External Massifs of the Alps
EMA	Eastern Meseta Arc (Morocco)
Es	Essarts Unit (Armorican Massif)
EV	Erbendorf-Vohenstrau□ Unit (Bohemian Massif)
F	Frankenberg Klippe (Bohemian Massif)
FF	Franconian Fault (Bohemian Massif)
FMC	French Massif Central
GCF	Granville-Cancale Fault (Armorican Massif)
Gf	Gföhl Unit (Bohemian Massif)
GN	Giessen Nappe (Rhenish Massif)
GS	Góry Sowie Massif (Bohemian Massif)
GTMZ	Galicia-Trás-os-Montes Zone (Iberian Massif)
HM	Harz Mountains (Germany)
HP-HT	High-P/High-T Upper Allochthon (European massifs)
IAA	Ibero-Armorican Arc
Ib	Ibiza (Balearic Islands)
IG	Ile-de-Groix (Armorican Massif)
IP	Intermediate-P Upper Allochthon (European massifs)
IRB	Iberian Reflective Body
ISB	Intra-Sudetic Basin (Bohemian Massif)
ISF	Intra-Sudetic Fault (Bohemian Massif)
JPSZ	Juzbado-Penalva Shear Zone (Iberian Massif)
Ka	Kaczawa Complex (Bohemian Massif)
KH	Kutna Hora Unit (Bohemian Massif)
Kr	Krkonoše Massif (Bohemian Massif)
LA	Lower Allochthon (European massifs)
LB	Lausitz Block (Bohemian Massif)
LC	Lizard Complex (England)
LiO	Limousin ophiolite (French Massif Central)
LLSZ	Lalaye-Lubine Shear Zone (Vosges Massif)
Ln	Léon (Armorican Massif)
Ly	Lyonnais (French Massif Central)
M	Morais (NW Iberia allochthonous complex)
Ma	Majorca (Balearic Islands)
MA	Middle Allochthon (European massifs)

MBM	Mont Blanc Massif (Alps)
MCA	Massif Central Arc
Me	Menorca (Balearic Islands)
MECS	Maures-Estérel-Corsica-Sardinia block
MGCH	Mid-German Crystalline High
ML	Mariánské Lázně Unit (Bohemian Massif)
MM	Moroccan Meseta
MMFZ	Middle Meseta Fault Zone (Morocco)
MN	Montagne Noire (French Massif Central)
Mo	Morvan Arc (French Massif Central)
Ms	Mauges Unit (Armorican Massif)
MSZ	Moravo-Silesian Zone (Bohemian Massif)
MT	Moldanubian Thrust (Bohemian Massif)
MTM	Maures-Tanneron Massif (Southern France)
MTu	Malpica-Tui (NW Iberia allochthonous complex)
Mü	Münchberg Klippe (Bohemian Massif)
MZ	Moldanubian Zone (Bohemian Massif)
NAD	North Armorican Domain
NASZ	North Armorican Shear Zone
NEF	Nort-sur-Erdre Fault (Armorican Massif)
Nj	Najac Massif (French Massif Central)
NM	Northern Massifs (Pyrenees)
NMe	Nové Město Unit (Bohemian Massif)
NPF	North Pyrenean Fault
NPhZ	Northern Phyllite Zone (Germany)
NSSB	Northern Sierra de Sevilla Batholith (Iberian Massif)
NZ	Nappe Zone (Morocco)
O	Órdenes (NW Iberia allochthonous complex)
Od	Odenwald Massif (Germany)
OF	Ouzilly Fault (French Massif Central)
OFZ	Odra Fault Zone (Bohemian Massif)
OMZ	Ossa-Morena Zone (Iberian Massif)
OS	Orlica-Śnieżnik Dome (Bohemian Massif)
OSD	Ollo de Sapo Domain (Iberian Massif)
OVD	Obejo-Valsequillo Domain (Iberian Massif)
PA	Parautochthon (NW Iberian Massif)
PASZ	Posada-Asinara Shear Zone (Sardinia)
PF	Pfahl Fault (Bohemian Massif)
PL	Pulo do Lobo Unit (Iberian Massif)
PS	Praha Syncline (Bohemian Massif)
PTSZ	Porto-Tomar Shear Zone (Iberian Massif)
QR	Quinto Real (Basque Massif, Pyrenees)
RHZ	Rhenohercynian Zone
RTFZ	Rabat-Tiflet Fault Zone (Morocco)
Ru	Ruhla Massif (Germany)
Sa	Sardinia
SAD	South Armorican Domain
SAF	South Atlas Fault (Morocco)
SASZ-N	South Armorican Shear Zone, Northern branch
SASZ-S	South Armorican Shear Zone, Southern branch
SB	Sehoul Block (Morocco)
SBB	South Bohemian Batholith
SBF	Sudetic Boundary Fault (Bohemian Massif)
SG	St. Georges-sur-Loire Unit (Armorican Massif)
SGD	Schist-Greywacke Domain (Iberian Massif)
SGr	Saxonian Granulitebirge (Bohemian Massif)
SHF	Sillon Houiller (French Massif Central)
SIF	South Iberia Fault
SISZ	Southern Iberian Shear Zone
SK	South Krkonoše Unit (Bohemian Massif)
SLF	South Limousin Fault (French Massif Central)
SM	Staré Město Unit (Bohemian Massif)
SMF	South Meseta Fault (Morocco)
SMFr	South Meseta Front (Morocco)
Sp	Spessart Massif (Germany)
SPhZ	Southern Phyllite Zone (Germany)
SPZ	South Portuguese Zone
STZ	Saxothuringian Zone
Sv	Svratka window (Bohemian Massif)
SWM	Schwarzwald Massif (Germany)
TA	Tuffs Anthracifères (French Massif Central)
TBZ	Teplá-Barrandian Zone (Bohemian Massif)
Th	Thaya window (Bohemian Massif)
TP	Thiviers-Payzac Unit (French Massif Central)
TTZ	Teisseyre-Tornquist Zone (Baltica)
UA	Upper Allochthon (European massifs)
UB	Ursuya-Baygura (Basque Massif, Pyrenees)
VD	Velay Dome (French Massif Central)
VF	Variscan Front
VM	Vosges Massif (France)
W	Wildenfels Klippe (Bohemian Massif)
WALZ	West Asturian-Leonese Zone (Iberian Massif)
WM	Western Meseta (Morocco)
WMSZ	Western Meseta Shear Zone (Morocco)