Tectonic evolution of Variscan Iberia: Gondwana – Laurussia collision revisited

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Abstract

An integrated interpretation of the late Paleozoic structural and geochronological record of the Iberian Massif is presented and discussed under the perspective of a Gondwana-Laurussia collision giving way to the Variscan orogen. Compressional and extensional structures developed during the building of the Variscan orogenic crust of Iberia are linked together into major tectonic events operating at lithosphere scale. A review of the tectonometamorphic and magmatic evolution of the Iberian Massif reveals backs and forths in the overall convergence between Gondwana and Laurussia during the amalgamation of Pangea in late Paleozoic times. Stages dominated by lithosphere compression are characterized by subduction, both oceanic and continental, development of magmatic arcs, (over- and under-) thrusting of continental lithosphere, and folding. Variscan convergence resulted in the eventual transference of a large allochthonous set of peri-Gondwanan terranes, the Iberian Allochthon, onto the Gondwana mainland. The Iberian Allochthon bears the imprint of previous interaction between Gondwana and Laurussia, including their juxtaposition after the closure of the Rheic Ocean in Lower Devonian times. Stages governed by lithosphere extension are featured by the opening of two short-lived oceanic basins that dissected previous Variscan orogenic crust, first in the Lower-Middle Devonian, following the closure of the Rheic Ocean, and then in the early Carboniferous, following the emplacement of the peri-Gondwanan allochthon. An additional, major intra-orogenic extensional event in the early-middle Carboniferous dismembered the Iberian Allochthon into individual thrust stacks separated by extensional faults and domes. Lateral tectonics played an important role through the Variscan orogenesis, especially during the creation of new tectonic blocks separated by intracontinental strike-slip shear zones in the late stages of continental convergence.

1. Introduction

The Variscan orogen and its continuation into the Appalachian–Alleghanian orogen have been the object of continuous rethinking and redefinition during the last decades (Burg and Matte, 1978; Hatcher, 1978, 2002; Matte, 1986; Castro, 1987a; Martínez Catalán et al., 1997; Franke, 2000; Ribeiro et al., 2007; Faure et al., 2008; Ballèvre et al., 2009; Simancas et al., 2009; Kroner and Romer, 2013). These orogens resulted from the late Paleozoic collision of Gondwana and Laurussia, and thus represent a broad, axial suture zone right in the heart of Pangea (Bambach et al., 1980).

The Iberian Massif contains one of the most complete sections across the Variscan orogen (Fig. 1), including several high-pressure metamorphic belts (Gil Ibarguchi and Ortega Gironés, 1985; Mata and Munhá, 1986; Abalos et al., 1991b; De Jong et al., 1991; Fonseca et al., 1993; Martínez Catalán et al., 1996; Rubio Pascual et al., 2013b) and tectonic units with ophiolitic assemblages that separate tectonic slices of continental crust (Arenas et al., 1986, 2007b; Crespo-Blanc and Orozco, 1988; Fonseca and Ribeiro, 1993; Díaz García et al., 1999; Pedro et al., 2005; Pin et al., 2006; Sánchez Martínez et al., 2009, 2012; Merinero et al., 2013, 2014). In the crystalline basement of central and southern Europe, such combination has been classically considered as indicative of multiple oceanic suture zones of Variscan age (Franke, 2000, 2006, 2014; Tait et al., 2000; Matte, 2001; Von Raumer et al., 2003; Martínez Catalán et al., 2007; Simancas et al., 2009).

Until recent times, cross-sections made to characterize the structure of the whole orogen were strongly influenced by the recognition of strike-slip systems separating tectonostratigraphic domains with opposite structural vergence (Burg et al., 1981; Matte, 1986, 2001; Simancas et al., 2013). Taking that, together with the varied geographic location of ophiolitic units, geodynamic models have put forward the hypothesis of a collage of individualized continental micro-terranes variably dispersed between Gondwana and Laurussia prior to the collision of the two latter (e.g., Winchester et al., 2002). However, new geochronological data show that the ophiolitic ensembles consist of different age rocks which in some cases indicate opening of ephemeral oceanic basins as Variscan orogenic crust was built (Azor et al., 2008; Arenas et al., 2014b), thus casting doubt on the multi-terrane picture as a general pre-orogenic feature. To this figure, most of the strike-slip shear zones that have been interpreted as separating supposedly different terranes rather correspond to late structures in the evolution of the Variscan orogen (Martínez Catalán, 2011; Gutiérrez-Alonso et al., 2015). The main strike-slip systems postdate a well-established phase dominated by tangential tectonics that includes development of continental subduction systems, obduction of suture zones, and emplacement of large thrust nappes (Iglesias Ponce de Leon and Choukroune, 1980; Gates et al., 1986; Martínez Catalán, 1990; Hatcher, 2002).

In the Iberian Massif, the most influential strike-slip system in defining the structure of the Variscan orogen has been the Coimbra-Córdoba shear zone (Fig. 1) (Burg et al., 1981; Ribeiro et al., 2007). The stretching lineation associated with this sinistral shear zone is subhorizontal (Pereira et al., 2008a, 2010a, 2010b), so movements along this structure cannot explain alone the juxtaposition of eclogites, high-P granulites, and blueschists to low-grade and low-pressure rocks unless tangential tectonics or vertical extrusion existed prior to or alternating with lateral movements.

Yet, the vergence of major orogenic structures seems to change at both sides of the Coimbra-Córdoba shear zone, being to the E and NE in the block located north (Pérez-Estaún et al., 1991; Azor et al., 1994a) and to the SW in the block located south (Simancas et al., 2001). However, a careful revision of the whole tectonic evolution of each block prevents us from accepting such consensual assumption. For instance, top-to-the-E and -NE kinematics associated with tangential deformation has been also described in major structures of the block located south of the Coimbra-Córdoba shear zone (Araújo et al., 2005; Pereira et al., 2007; Rosas et al., 2008; Borrego, 2009; Ribeiro et al., 2010; Ponce et al., 2012), plus the age of structures showing opposite vergence across the Iberian Massif is different in most of the cases (see following sections).

For many years the tectonic evolution of the Iberian Massif has been presented and evaluated in a similar way to a double-blind procedure. Advances coming from its northwestern section were little considered in its southwestern part and the other way around. In the meantime, the amount of structural and geochronological data in both regions have increased noticeably, to a point that a structural correlation between these two sections of the Variscan orogen would allow a more precise integration of geological processes previously described elsewhere in the Iberian Massif.

Here we follow on to the discovery that the Iberian Massif contains allochthonous and autochthonous tectonic units that share fundamental features of their tectonostratigraphy and can be structurally correlated across the two blocks separated by the Coimbra-Córdoba shear zone, i.e. the pile of allochthonous units recognized in NW Iberia can be correlated with that of SW Iberia (Fig. 2; Díez Fernández and Arenas, 2015). Within this framework, this article presents an integrated model showing the Variscan evolution of the Iberian Massif by piecing together the structural history and chronology of major individual events registered across different sections of its orogenic crust.

2. Synthetic cross-section and tectonostratigraphy of the Iberian Massif

The Iberian Massif combines folds and thrust faults formed during Variscan compression with open structural domes and basins, granitization and normal faults that are a product of subsequent gravitational collapse and intra-orogenic extension (Simancas et al., 2001; Martínez Catalán et al., 2007; Pereira et al., 2009). Figure 3 shows a simplified, composite cross-section of the Iberian Massif highlighting eight major tectonostratigraphic elements of the Variscan orogen (Díez Fernández and Arenas, 2015), namely: (1) Cantabrian Zone foreland, (2) Iberian Autochthon, (3) Iberian Parautochthon, (4) Basal Allochthonous Units, (5) Allochthonous Ophiolitic Units, (6) Upper Allochthonous Units, (7) Beja-Acebuches Ophiolite, and (8) South Portuguese Zone. Gathering of 4, 5, and 6 will be also referred to as Iberian Allochthon (or simply Allochthon).

A simple restoration of the cross-section shown in Figure 3 (Fig. 4a) reveals two major suture zones featured with ophiolitic units with different protolith ages. The NNE-dipping suture represented by the Beja-Acebuches Ophiolite (Bard and Moine, 1979; Munhá et al., 1986; Quesada et al., 1994) cuts the suture marked by the Allochthonous Ophiolitic Units (Arenas et al., 1986) and divides the Iberian Massif in two major blocks. The northern block comprises the Cantabrian Zone, and the Iberian Autochthon, Parautochthon and Allochthon, all of which have Gondwanan affinity

(Fernández-Suárez et al., 2003; Robardet, 2003; Martínez Catalán et al., 2004; Díez Fernández et al., 2010, 2012b; Pastor-Galán et al., 2013; Albert et al., 2015a; Pereira, 2015). In the southern block, the South Portuguese Zone is considered a composite terrane located along the southern flank of Laurussia (Avalonia-Meguma) by the Middle Devonian (Lefort et al., 1988; Matte, 1991; Quesada et al., 1994; Simancas et al., 2005; Braid et al., 2011), but close to or juxtaposed with Gondwana since (at least) the Upper Devonian (Pereira et al., 2006b).

The Iberian Massif consists of sedimentary, plutonic and metamorphic rocks whose grade ranges from very low to catazonal. The Gondwanan series consist of Precambrian and Paleozoic marine sequences alternating with volcanic and plutonic rocks of variable age and abundance. These series are deformed into a thin-skinned fold and thrust belt with easterly propagation in the Cantabrian Zone (Pérez-Estaún et al., 1988), and into a crystalline thrust sheet characterized by large recumbent folds (e.g., Mondoñedo Nappe). This tectonic stacking was followed by the development of extensional faults and doming (Lugo dome) in the more external zones of the Gondwanan Variscan hinterland (West Asturian-Leonese Zone; Martínez Catalán et al., 2003). Towards more internal zones of the Variscan orogen, deformation affecting the Iberian Autochthon produces upright, overturned, and recumbent folds (Díez Balda, 1986; Macaya et al., 1991; Díaz Azpiroz et al., 2003; Dias et al., 2010; Díez Fernández et al., 2013b), which are underlain by a complex system of granite- and migmatite-cored dome structures (Escuder Viruete et al., 1994; Díez Balda et al., 1995; González del Tánago, 1995; Pereira et al., 2009; Díaz-Alvarado et al., 2012; Díez Fernández et al., 2012c; Arango et al., 2013; Rubio Pascual et al., 2013a).

A significant part of the Iberian Autochthon occurs under a huge thrust stack (Ribeiro et al., 1964; Ries and Shackleton, 1971). The lowermost set of thrust nappes is the Iberian Parautochthon (Farias et al., 1987; Ribeiro et al., 1990), which is restricted to the sections of the Variscan orogen that were closer to the Gondwana mainland and contains alternating volcanic and sedimentary rocks with remarkable lateral variability at regional scale (Dias da Silva et al., 2014). The lower contact of the overlying Basal Allochthonous Units traces the base of a large allochthonous ensemble, the Iberian Allochthon, transported onto the Iberian Autochthon and Parautochthon. Remnants of the Iberian Allochthon are found as klippen in the complexes of Cabo Ortegal, Órdenes, Malpica-Tui, Bragança and Morais in NW Iberia (Martínez Catalán et al., 2007), whereas in the Ossa-Morena Complex of SW Iberia, the base of the allochthonous ensemble is defined by high-P metamorphic "belts" (with equivalent protolith and metamorphic ages and lithological composition) that crop out at similar structural position across the Iberian Variscides (Díez Fernández and Arenas, 2015, 2016). The Basal and Upper Allochthonous Units are separated by a set of dismembered Allochthonous Ophiolitic Units.

The Basal and Upper Allochthonous Units comprise arc-related Precambrian and Cambrian sedimentary rock sequences intruded by or alternating with arc-related and later alkaline to peralkaline Cambrian and Ordovician igneous rocks. Most of their series, therefore, bear the imprint of Ediacaran and Early Paleozoic subduction and subsequent Cambro-Ordovician rifting (Arenas et al., 1986, 2009; Ribeiro and Floor, 1987; Liñán and Quesada, 1990; Quesada, 1990; Sánchez-García et al., 2003; Pereira et al., 2006a; Díez Fernández et al., 2010, 2015; Andonaegui et al., 2012). These sequences are succeeded by passive margin strata that covers up to the lowermost

Devonian in the Upper Allochthonous Units of the Ossa-Morena Complex (Robardet and Gutiérrez Marco, 2004), but it is lacking in the rest of the allochthonous complexes.

The Allochthonous Ophiolitic Units are made of mafic and ultramafic rocks and scarce metasedimentary rocks (see compilation by Arenas and Sánchez Martínez, 2015). These units can be divided in two groups according to protolith ages and chemical signature (Arenas et al., 2007a; Sánchez Martínez et al., 2009): a Cambrian-Ordovician group related to the closure of the Iapetus/Tornquist Ocean (Bazar ophiolite; Sánchez Martínez et al., 2012) and opening of the Rheic Ocean s.l. (Vila de Cruces, Izeda-Remondes, and Internal Ossa-Morena Zone ophiolites; Pin et al., 2006; Arenas et al., 2007b; Pedro et al., 2010), and an Lower-Middle Devonian group (Careón, Purrido, Moeche, and Morais-Talhinhas ophiolites; Díaz García et al., 1999; Pin et al., 2006; Sánchez Martínez et al., 2011) representing an ephemeral oceanic basin formed in early stages of the Variscan orogenesis (Arenas et al., 2014b). When these two groups contact with each other the Devonian counterparts occur on top.

The Beja-Acebuches Ophiolite is a rather continuous band of amphibolites, minor ultramafic rocks, mylonitic gabbros and a sheeted dike complex (Bard, 1977; Bard and Moine, 1979; Fonseca and Ribeiro, 1993; Quesada et al., 1994) that have been interpreted as a relict oceanic crust formed in a rifting context s.l. (Dupuy et al., 1979), either in a back-arc or intra-arc setting (Quesada et al., 1994), or in a mid-ocean ridge (Castro et al., 1996). Protoliths of this oceanic crust have been dated at Viséan, so it formed long after the onset of Variscan deformation (Azor et al., 2008).

The northern domains of the South Portuguese Zone contain an ensemble of metasedimentary rocks and lenses of metabasites (Pulo do Lobo Unit; Carvalho et al., 1976; Oliveira, 1990; Fonseca, 2005). The Pulo do Lobo Unit has been considered as an accretionary prism related to a north-dipping subduction zone (Eden and Andrews, 1990; Silva et al., 1990; Braid et al., 2010) or as a dismembered ophiolitic ensemble s.l. (Fonseca and Ribeiro, 1993) connected to the closure of the Beja-Acebuches oceanic basin (Quesada et al., 1994). Alternatively, other authors suggest it may represent an accretionary prism over a south-dipping subduction zone developed prior to the opening of the Beja-Acebuches oceanic basin (Azor et al., 2008). The general structure of this domain is an upright antiform (Silva et al., 1990; Martínez Poza et al., 2012; Pérez-Cáceres et al., 2015) and the age of its series could be as old as Silurian in the lower part of the stratigraphy (Braid et al., 2011), and have been dated at Middle-Upper Devonian in the intermediate part (Pereira et al., 2008b), and early Carboniferous in the overlying, discordant series (Santa Iria basin, Pereira et al., 2008b; Braid et al., 2011). The Iberian Pyrite Belt is located to the south of the Pulo do Lobo Unit and contains Upper Devonian sedimentary rocks and Upper Devonian to early Carboniferous volcanic strata covered by younger Carboniferous turbiditic series (Oliveira, 1990; Leistel et al., 1997; Pereira et al., 2012a). Its regional structure is defined by a southerly propagation of a fold and thrust belt (Silva et al., 1990).

3. Record of Variscan events

Table 1 is a synopsis that presents a summary of the main tectonic events that characterize the principal geotectonic zones of the Iberian Massif. It offers a general view of the time-based correlation of contrasted geological processes presented in this work (descriptions and citations in the text below). For comparison with isotopic ages,

ages obtained from fossil record have been converted into absolute ages following the IUGS International Chronostratigraphic Chart 2013 (Cohen et al., 2013).

3.1. Lower Devonian (Lochkovian-Emsian)

3.1.1. Upper Allochthonous Units of NW Iberia

The first phase of Variscan deformation in NW Iberia corresponds to a high-P/high-T tectonometamorphic event that is traceable across the lowermost structural levels of the Upper Allochthonous Units of Cabo Ortegal (Vogel, 1967; Gil Ibarguchi et al., 1990; Puelles et al., 2005; Albert et al., 2012), Órdenes (Arenas and Martínez Catalán, 2002; Gómez Barreiro, 2007), Bragança and Morais complexes (Marques et al., 1996). Altogether these structural levels define a high-P/high-T metamorphic belt characterized by variably-retrogressed eclogitic foliation formed in the course of dextral, west-directed continental subduction (Martínez Catalán et al., 1997; Ábalos et al., 2003). The high-P rocks occur below a thick series of siliciclastic rocks affected by Variscan intermediate-P metamorphism (Castiñeiras, 2005). Yet, the uppermost structural levels of this series still preserve the imprint of a previous Cambrian deformation (Díaz García et al., 2010). The high-P/high-T metamorphism has been dated at ca. 410-390 Ma (Santos Zalduegui et al., 1996; Ordóñez Casado et al., 2001; Fernández-Suárez et al., 2002, 2007). According to provenance analysis of their sedimentary sequences, both the upper and lower plate to this continental subduction zone represents a piece of Gondwanan continental crust located in the periphery of the West African Craton (Fernández-Suárez et al., 2003; Albert et al., 2015a, 2015b). This metamorphic belt occurs together with a set of ultramafic rocks considered as one of the world-class examples of heterogeneous upper mantle (Girardeau et al., 1989; Girardeau and Ibarguchi, 1991).

3.1.2. Upper Allochthonous Units of SW Iberia

The Precambrian to Lower Devonian series of the Upper Allochthonous Units of the Ossa-Morena Complex are deformed into a SW-verging train of recumbent folds (e.g., Monesterio Anticline) with associated axial planar foliation developed under Barrovian, low-T conditions (Vauchez, 1974, 1976; Chacón, 1979; Chacón et al., 1983; Apalategui et al., 1990; Quesada, 1990; Expósito Ramos, 2005). These folds are cut by south-directed thrusts (e.g., Monesterio Thrust; Eguíluz, 1987) and are covered by a syn-orogenic basin (Terena flysch), whose base rests discordant over the folds (Expósito et al., 2002). Some authors consider this flysch yet coeval with the SW-verging folds (Rocha et al., 2009; Araújo et al., 2013).

The base of the Terena flysch has been dated at Lower Devonian (Lochkovian; Piçarra, 1997; Pereira et al., 1998, 1999; Piçarra et al., 1998; Robardet and Gutiérrez Marco, 2004; Rocha et al., 2010). The youngest series that is bent into the recumbent folds contains intraformational conglomerates indicating basin instability and syn-orogenic deformation (Giese et al., 1994). Limestones in that series have been dated with fossils at Lochkovian-Emsian (Perdigão et al., 1982; Oliveira et al., 1991, 1992; Robardet et al., 1998), so the age of the SW-verging folds can be further constrained between ca. 419 Ma and ca. 393 Ma, and deformation preceding or coeval with fold nucleation probably started in the Lochkovian. Other age constrains exist for fold formation, such as K/Ar ages obtained from biotite associated with the axial planar foliation (385±11 Ma; Galindo et al., 1986, 1987) and the rejuvenation of Precambrian (protolith?) ages in amphibolites and metasedimentary rocks at ca. 400-390 Ma (Dallmeyer and Quesada, 1992). The later thrusts are covered by lowermost

Carboniferous sedimentary rocks, so these south-directed faults have been considered part of a continuous deformational process following fold amplification (Expósito et al., 2002).

3.2. Lower-Middle Devonian (Emsian-Eifelian-Givetian)

3.2.1. Upper Allochthonous Units of SW Iberia

Located to the north and near the Pulo do Lobo Unit, there are few outcrops of Emsian-Eifelian reef limestones with interbedded mafic calc-alkaline volcanic rocks that may represent a magmatic arc associated with the closure of the Rheic Ocean (Silva et al., 2011). Middle Devonian sedimentary rocks are very scarce or absent in the southern part of the Upper Allochthonous Units of SW Iberia, i.e. exposures south of the Coimbra-Córdoba Shear Zone (Robardet and Gutiérrez Marco, 2004).

Conversely, Givetian sedimentary rocks might exist in the block located north of the Coimbra-Córdoba shear zone, in the so-called Obejo-Valsequillo Domain (Sánchez Cela and Gabaldón, 1977a; Pérez Lorente, 1979; Rodríguez and Soto, 1979; Pardo and García Alcalde, 1996; García-López et al., 1999). A recent revision of the stratigraphy of the Obejo-Valsequillo Domain has not confirmed the occurrence of Middle Devonian sedimentary rocks (Matas et al., 2015a). Therefore, either the absence or scarcity of sediments of that age seems a common feature among the Upper Allochthonous Units of SW Iberia. The widespread Middle Devonian sedimentary gap has been explained by significant crustal uplift during Devonian tectonic activity (Giese et al., 1994).

3.2.2. Allochthonous Ophiolitic Units of NW Iberia

The protoliths of the Devonian Allochthonous Ophiolitic Units account for a short period of lithosphere extension and ocean-floor generation (Díaz García et al., 1999). Preliminary geochemical data led to place such extension in an intra-oceanic subduction setting (Sánchez Martínez et al., 2007), or as connected to incipient collisional processes (Pin et al., 2002). The isotopic signature of zircon in these ophiolites favors a setting that involves continental crust in their formation (Sánchez Martínez et al., 2011). A revision of all the geochemical, isotopic, and regional data available for these ophiolites in Iberia, considered a short-lived oceanic basin opened within a continental realm at ca. 400-395 Ma as their most likely setting (Arenas et al., 2014b). In any case, lithosphere extension occurred in a domain located between the paleogeographic realms of the Upper and Basal Allochthonous Units, both of which represent sections of Gondwana.

3.2.3. Iberian Autochthon and Cantabrian Zone

A series of Lower-Middle Devonian events distributed across the Iberian Massif have been integrated into an extensional setting affecting a large tract of the Gondwana margin at ca. 395 Ma (Gutiérrez-Alonso et al., 2008). These include alkaline volcanism in the Iberian Autochthon (Gutiérrez-Alonso et al., 2008) and Cantabrian Zone (Loeschke, 1983), increased subsidence in the Cantabrian Zone (Veselovski, 2004), and a sedimentary gap in the Iberian Autochthon (Puschmann, 1967).

3.3. Upper Devonian (Frasnian-Fammenian)

3.3.1. Upper Allochthonous Units of NW Iberia

The high-P/high-T record of the Upper Allochthonous Units is variably retrogressed as a consequence of a multi-stage Upper Devonian exhumation process

(Gómez Barreiro et al., 2007). Post-eclogitic contractional shearing was developed under amphibolite facies conditions and produced a widespread mylonitic foliation (Castiñeiras, 2005; Gómez Barreiro, 2007). The high-P/high-T units are separated from overlying intermediate-P units by a set of extensional detachments (Martínez Catalán et al., 2002; Castiñeiras, 2005), dated at ca. 375-371 Ma (Dallmeyer et al., 1997; Gómez Barreiro et al., 2006). Both, the regional mylonitic foliation and the extensional detachments, are affected by east-verging recumbent folds at a regional scale, both in the Órdenes Complex (Martínez Catalán et al., 2002; Gómez Barreiro, 2007; González Cuadra, 2007) and in the Cabo Ortegal Complex (Marcos et al., 1984; Ábalos et al., 2003; Albert et al., 2012). Therefore the nucleation of these recumbent folds occurred later than ca. 371 Ma. Besides the age of intervening extensional structures, the exhumation of the high-P and high-T units to amphibolite facies conditions has been estimated at ca. 380 Ma (Van Calsteren et al., 1979; Peucat et al., 1990; Dallmeyer et al., 1991; Santos Zalduegui et al., 1996; Valverde Vaquero and Fernández, 1996; Gómez Barreiro et al., 2006), whereas subsequent retrogression to greenschist facies conditions is dated at ca. 360-350 Ma (Peucat et al., 1990; Dallmeyer et al., 1997).

3.3.2. Upper Allochthonous Units of SW Iberia

The southern part of the metasedimentary series exposed in the Obejo-Valsequillo Domain are located in the hanging wall of a NE-directed thrust (Espiel thrust) and are deformed into a NE-verging train of recumbent folds (Martínez Poyatos et al., 2001; Martínez Poyatos, 2002). The youngest rocks affected by these folds are Devonian. Some authors propose the existence of Middle Devonian series (Givetian; Sánchez Cela and Gabaldón, 1977a; Pérez Lorente, 1979; Rodríguez and Soto, 1979; Pardo and García Alcalde, 1996), whereas other authors have also found Upper

Devonian strata (Frasnian-Famennian; Febrel and Saenz de Santa María, 1964; Herranz, 1985; Racheboeuf et al., 1986). On the contrary, Matas et al. (2015a) restrict the Upper Devonian series to the footwall of the Espiel thrust, and identify a stratigraphic hiatus spanning from the Middle to the Upper Devonian in the series affected by the NE-verging folds. These folds are covered by a discordant Carboniferous succession (Culm facies), whose base is dated at Tournaisian (ca. 359-347 Ma; Sánchez Cela and Gabaldón, 1977b; Garrote and Broutin, 1979; García Alcalde et al., 1984; Rodríguez et al., 1990). Amphibolites occurring in the Precambrian series and affected by these folds show rejuvenation of a Precambrian metamorphic imprint mainly during the Upper Devonian (Dallmeyer and Quesada, 1992). Consequently, the age of recumbent folding is mostly Upper Devonian-early Carboniferous (ca. 383-347 Ma).

3.3.3. Allochthonous Ophiolitic Units

The age of imbrication of the Cambrian-Ordovician and Devonian ophiolitic ensemble exposed in NW Iberia has been constrained, by means of $^{40}Ar/^{39}Ar$ dating of their low- to medium-T metamorphic fabrics (greenschist and amphibolite facies), to a range that extends from ca. 391 Ma to ca. 364 Ma (Peucat et al., 1990; Dallmeyer et al., 1997). The metamorphic grade decreases progressively down structure, as the ages of metamorphism become younger (Arenas et al., 2007a). Kinematic indicators for accretion consistently indicate top-to-the-Cantabrian foreland, i.e. subduction under the Upper Allochthonous Units (Arenas et al., 2007b; Gómez Barreiro et al., 2010b).

3.3.4. Basal Allochthonous Units of NW Iberia

The first deformation event recorded in the Basal Allochthonous Units relates to continental subduction (Gil Ibarguchi and Ortega Gironés, 1985; Martínez Catalán et al., 1996). This tectonic event developed a penetrative fabric whose remnants occur in weakly or non-retrogressed lenses of high-P/low-intermediate-T rocks and as mineral trails within porphyroblasts grown during decompression (Díez Fernández and Martínez Catalán, 2012). Kinematic criteria associated with fabrics developed under high-P/low-intermediate-T metamorphic conditions show a consistent top-to-the-northeast shear-sense, which indicates dextral oblique subduction to the west in present-day coordinates (Díez Fernández et al., 2012a). Deformation took place under metamorphic conditions ranging from blueschist to eclogite facies (Munhá et al., 1984; Arenas et al., 1995, 1997; Gil Ibarguchi, 1995; Rubio Pascual et al., 2002; Rodríguez et al., 2003; López-Carmona et al., 2010, 2014), and has been consistently dated at ca. 380-370 Ma (Van Calsteren et al., 1979; Santos Zalduegui et al., 1995; Rodríguez et al., 2003; Abati et al., 2010).

Subsequent exhumation of these units was controlled by crustal-scale ductile thrusting (Fervenza thrust), which was accompanied by recumbent folding and attenuation of overlying lithosphere (tectonic and erosional). Early exhumation produced a mylonitic foliation throughout the high-P metamorphic belt. The kinematics of this event is consistent with east-directed tectonic transport (Díez Fernández et al., 2011). Deformation took place under high-P conditions, although in a clear decompressive path (Rodríguez et al., 2003; Díez Fernández et al., 2011). Dating of tectonic fabrics and partial melting formed in the course of subsequent exhumation under lower pressure conditions (amphibolite/greenschist facies) has yielded ages in the range ca. 360-346 Ma (Abati and Dunning, 2002; Rodríguez et al., 2003; López-Carmona et al., 2014). Therefore, the initial exhumation from peak-pressure conditions up to the lower crust occurred in the Famennian (ca. 370-360 Ma).

3.3.5. Basal Allochthonous Units of SW Iberia

In SW Iberia, the Basal Allochthonous Units crop out in two domains (Fig. 2). A northern domain made of continental crust, referred to as the Central Unit (Azor et al., 1994b), experienced high-P/low-to high-T metamorphism (Mata and Munhá, 1986; Eguiluz et al., 1990; Abalos et al., 1991b; Pereira and Apraiz, 2006; Pereira et al., 2010a). Most of this record is strongly retrogressed into amphibolite/greenschist facies rocks, within which only small retrogressed high-P granulite and eclogite pods and mineral relicts testifies for their subduction-related history. As a consequence, no kinematic criteria have been provided so far for the development of the high-P fabrics. This way, the current NE-dipping character of the post-eclogitic foliation has been taken as the sole criteria for a NE subduction polarity (Azor et al., 1994b).

The individual ages obtained for the high-P metamorphism in the Central Unit show some variation but complementary results. First attempts to date this deformation yielded very imprecise Silurian-Devonian ages that called for a Variscan event (427±45 Ma; Schäfer et al., 1991). Subsequent surveys performed in eclogite boudins provided a minimum age of ca. 370-360 Ma for the high-P metamorphism in the Central Unit (Quesada and Dallmeyer, 1994). Finally, a more specific, yet imprecise age of ca. 380-350 Ma was obtained for this metamorphic event by means of U-Pb zircon dating (Ordóñez Casado, 1998). On the other hand, dating of mylonites formed during post-peak-P metamorphism yielded ages of ca. 355 Ma (García Casquero et al., 1988). If we consider all these data, the timing of continental subduction experienced by these rocks should be restricted to the Upper Devonian, probably ranging between ca. 380-360 Ma.

The Basal Allochthonous Units that are located in the southernmost domain of the Ossa-Morena Complex were recently referred to as Cubito-Moura Unit, which, similarly to the rest of Basal Allochthonous Units, consists of two juxtaposed sequences (Díez Fernández and Arenas, 2015). The upper sequence of this unit is known under the name of Cubito-Moura Schists (Fonseca et al., 1999). The lower sequence includes a series of metasedimentary rocks and orthogneisses referred to as Fuenteheridos group (Rubio Pascual et al., 2013b), or as the Serie Negra succession and Igneous-felsic dominated-sedimentary complex (only sections affected by high-P metamorphism; Chichorro, 2006; Chichorro et al., 2008; Rosas et al., 2008). Altogether these sequences represent a piece of continental crust subjected to high-P/low-intermediate-T metamorphism, as typified by evidence (blueschist and eclogite boudins and weakly-retrogressed mineral assemblages) of a first tectonic event (De Jong et al., 1991; Fonseca et al., 1993, 1999, 2004; Pedro, 1996; Leal et al., 1997; Moita, 1997; Booth-Rea et al., 2006; Ponce et al., 2012; Rubio Pascual et al., 2013b).

Strong retrogression affected the high-P rocks of the Cubito-Moura Unit during exhumation and no kinematic criteria extracted from direct observation of high-P fabrics have been reported so far for the subduction process. Yet, both NE and SW subduction polarities have been proposed (Fonseca et al., 1999; Díaz Azpiroz et al., 2004; Ribeiro et al., 2007; Rosas et al., 2008; Pin et al., 2008; Simancas et al., 2009; Rubio Pascual et al., 2013b). The age of high-P metamorphism obtained for the lower sequence of this unit is ca. 371 Ma (Moita et al., 2005), which is in agreement with an age of ca. 358 Ma obtained for post-peak-P metamorphism (Rosas et al., 2008).

Preservation of the structural record associated with the initial stages of exhumation is very rare in the Cubito-Moura Unit. One example corresponds to the development of boudinaged-folds affecting marbles and amphibolites (Rosas et al., 2002) under a top-to-the-SSW shear sense regime (Rosas et al., 2008). This particular shearing has been inferred older than a cooling age of ca. 358 Ma (Rosas et al., 2008), and consequently it should have been developed during the Famennian (ca. 371-358 Ma).

3.4. Early Carboniferous (Tournaisian-Viséan)

3.4.1. Upper Allochthonous and Ophiolitic Units of NW Iberia

A set of out-of-sequence-thrusts carried the Upper Allochthonous and Ophiolitic Units of NW Iberia approximately to their current position in the thrust pile (see extended description in Martínez Catalán et al., 2002). Out-of-sequence thrusting proceeded under low-T conditions (greenschist facies) and generated mylonites and ultramylonites with an associated top-to-the-southeast sense of motion. This thrust system moved the Upper Allochthonous Units over the Ophiolitic Allochthonous Units and affected thrust sheets of the previously (under)stacked ophiolites (Díaz García et al., 1999; Arenas et al., 2007b). The emplacement of the Upper Allochthonous Units generated duplexes with both hinterland-dipping and foreland-dipping horses, duplexes of antiformal stack type, as well as isolated horses made of both the Upper Allochthon and Ophiolitic Allochthonous Units.

The out-of-sequence thrusts cut and partially rework extensional shear zones dated at ca. 375-371 Ma. They also cut the regional fabrics of the Upper Allochthonous Units associated with their retrogression to greenschist facies conditions dated at ca.

360-350 Ma. They might also affect some of the uppermost Basal Allochthonous Units, such as the Agualada Unit. The age of post-peak-P partial melting in that unit is ca. 346 Ma (Abati and Dunning, 2002), so the out-of-sequence thrusting is likely younger. This thrust system also cuts the Lalín-Forcarei Thrust (age estimated at ca. 340 Ma; Dallmeyer et al., 1997), which is structurally below and overprinted by an event of amphibolite facies metamorphism that is absent in the out-of-sequence thrusts. The out-of-sequence thrusts must be older than the extensional detachments of Pico Sacro and Bembibre-Ceán (Díez Fernández et al., 2012c), which cut all previous thrust faults and are bracketed between ca. 323-314 Ma (Martínez Catalán et al., 2002) and dated at ca. 337 Ma (López-Carmona et al., 2014), respectively. Fine-grained white micas extracted from a greenschist facies ultramylonitic gneiss located in an out-of-sequence thrust contact gave an $^{40}Ar/^{39}Ar$ cooling age of ca. 325 Ma. Considering all the previous data the most probable age for the major event of out-of-sequence thrusting is Viséan (ca. 346-337 Ma). However, a thrust-related phyllonite located at the base of the Cabo Ortegal Complex yielded an age of ca. 316 Ma (Dallmeyer et al., 1997), what certainly opens the possibility of later episodes of out-of-sequence thrusting in NW Iberia (Martínez Catalán et al., 2002).

3.4.2. Upper Allochthonous Units of SW Iberia

One of the most salient structures affecting the Iberian Allochthon exposed in the Obejo-Valsequillo Domain is a thrust fault that transports to the NE a NE-verging train of recumbent folds of Upper Devonian age (see section 3.3.2) onto Culm facies sedimentary rocks, the Espiel Thrust (Figs. 2 and 3; Martínez Poyatos et al., 2001; Martínez Poyatos, 2002). Some of the youngest series of its footwall may be upper Viséan or even early Serpukhovian (~340-330 Ma; Ortuño, 1971; Sánchez Cela and

Gabaldón, 1977b; Garrote and Broutin, 1979; Matas et al., 2015a), what places a reference age for the onset of thrusting. On the other hand, a regional analysis carried out in the syn-orogenic series, that occur both in the hanging wall and footwall of this fault, revealed that thrusting may have been active from the Viséan up to the upper Bashkirian (~340-315 Ma; Martínez Poyatos et al., 1998; Matas et al., 2014).

In the first description of the Espiel Thrust, Apalategui and Pérez-Lorente (1983) documented the existence of peridotites, amphibolites, mylonitic gneisses and phyllonites in close relation to some sections of its fault zone. These authors also suggested a correlation between those gneisses and the lithological ensemble that today constitutes the Basal Allochthonous Units bounding the Obejo-Valsequillo Domain to the south (the so-called Central Unit). The presence of peridotites (even if scarce) together with gneisses that have been considered as markers of a suture zone elsewhere, represent a singularity within the Upper Allochthonous Units of SW Iberia. Such singularity can be solved by considering the Espiel Thrust as a series of out-of-sequence structures that cut, bounded and transported upwards an underlying and previously structured suture zone consisting of high-P gneisses and ophiolitic rocks (peridotites and amphibolites).

NE of the Espiel Thrust, a group of reverse faults with top-to-the-east tectonic transport makes a tectonic imbricate and duplexes of continental crust within the Upper Allochthonous Units (Zalamea de la Serena imbricates; Castro, 1987b). Due to their similar geometry and kinematics, these faults have been considered as genetically related to the main thrust system that transported the NE-verging recumbent folds (Martínez Poyatos et al., 2001).

3.4.3. Basal Allochthonous Units of NW Iberia

The mylonitic shear zones developed during the Famennian (Fervenza Thrust; see section 3.3.4), together with the rest of the lithostratigraphy of the Basal Allochthonous Units, are affected by a regional-scale train of recumbent folds (Díez Fernández et al., 2011). The regional foliation in the Basal Allochthonous Units is axial planar to these folds (Díez Fernández, 2011) and developed under greenschist to amphibolite facies conditions (Rodríguez et al., 2003). The asymmetry depicted by this train and kinematic criteria observed in the axial planar foliation indicates top-to-the-Cantabrian Zone sense of shear (Díez Fernández and Martínez Catalán, 2012), which is in agreement with quartz crystal preferred orientation fabrics (Llana-Fúnez, 2002; Gómez Barreiro et al., 2010a; Fernández et al., 2011). Dating of this regional foliation has yielded Tournaisian ages in the range ca. 360-350 Ma (Rodríguez et al., 2003).

To the external parts of the hinterland, the regional foliation of the Basal Allochthonous Units is bent into a large recumbent anticline and minor associated folds (Carrio Anticline). The Lalín-Forcarei Thrust runs along the reverse limb of that major anticline, and produced ductile mylonitic deformation throughout the base of the Iberian Allochthon. Kinematic criteria show an eastward movement for this thrust, which accounts for the emplacement of the Iberian Allochthon onto the Iberian Parautochthon (Martínez Catalán et al., 1996). Isotopic ages for this nappe-fold structure are lacking, although fold nucleation must be younger than ca. 360-350 Ma (age of the folded foliation at a regional scale). A reference age for the Lalín-Forcarei Thrust is ca. 340 Ma, which is the timing of pervasive ductile deformation recorded by the Iberian

Parautochthon in response to the overriding of the Iberian Allochthon s.l. (Dallmeyer et al., 1997).

The uppermost part of the Basal Allochthonous Units of NW Iberia experienced high-T conditions and partial melting during post-eclogitic decompression (Arenas et al., 1997). Post-peak-P partial melting has been dated at ca. 346-341 Ma (Abati and Dunning, 2002).

3.4.4. Basal Allochthonous Units of SW Iberia

Ductile shearing associated with the exhumation of the Basal Allochthonous Units exposed in the Central Unit generated a penetrative foliation under amphibolite to greenschist facies conditions (Abalos et al., 1991a, 1991b), reaching the granulite facies in some sections (Pereira et al., 2010a). Although no major structures have been recognized so far within this unit, recumbent folding has been considered as a process related to the formation of the regional planar fabric (Azor, 1994). The kinematics shown by this fabric is top-to-the-NW, an orogen-parallel trend that may indicate a strong lateral shear component during and/or after exhumation (Azor et al., 1994b; Pereira et al., 2008a, 2010a).

The timing of exhumation show some variation in the Central Unit, but a Tournaisian-Viséan age can be taken as the most likely, if the whole set of geochronological data available is taken in consideration. Dating of biotite in gneisses yielded ages of ca. 335-330 Ma (Blatrix and Burg, 1981; Pereira et al., 2012b), in agreement with subsequent results obtained in muscovite (ca. 340-333 Ma; Dallmeyer and Quesada, 1992; Pereira et al., 2012b). The maximum age for the early (and warmer) stages of exhumation can be vaguely constrained by the oldest ages obtained from amphibole in mafic rocks (ca. 370 Ma; Dallmeyer and Quesada, 1992), although dating of the regional fabric by Rb-Sr in mica provided an age of ca. 355 Ma (García Casquero et al., 1988). Additionally, some sections of the Central Unit experienced high-T conditions and eventual partial melting in the course of their exhumation to the mid-lower crust, a process that has been dated at ca. 340 Ma (Ordóñez Casado, 1998; Pereira et al., 2010a).

The Basal Allochthonous Units that occur to the south of the Ossa-Morena Complex (Cubito-Moura Unit) exhibit a widespread foliation developed under amphibolite to greenschist facies conditions (Pedro, 1996; Moita, 1997; Fonseca et al., 1999; Booth-Rea et al., 2006). No major structures associated with this fabric have been identified so far within this unit, although the generation of large recumbent folds has been tentatively proposed (Araújo and Ribeiro, 1995). In those sections where this foliation is best preserved, the analysis of kinematic criteria has provided a consistent top-to-the-northeast tectonic transport (Araújo et al., 2005; Rosas et al., 2008), whereas a top-to-the-east sense of shear has been deduced for those domains affected by subsequent tangential deformation (kinematics inferred after unfolding tectonic fabrics; Ponce et al., 2012). In both cases the tectonic transport includes a significant component directed to the foreland located in the Cantabrian Zone. The maximum age of this fabric is ca. 358 Ma (Rosas et al., 2008), whereas the intrusion of Variscan magmas in the region at ca. 350-340 Ma (Azor et al., 2008; Pin et al., 2008) could be taken as a reliable minimum age for this deformation.

3.4.5. Iberian Parautochthon and Autochthon of NW and Central Iberia

The onset of Variscan deformation in the Iberian Parautochthon is characterized by overturned to recumbent folds with axial planar foliation. These folds are overprinted by a pervasive flat-lying foliation. These two initial deformation events developed under prograde intermediate-P, Barrovian-type metamorphism and have been related to the juxtaposition of the Iberian Allochthon over the Iberian Parautochthon and then onto the Iberian Autochthon (Marquínez García, 1984; Farias et al., 1987; Ribeiro et al., 1990; Dias da Silva, 2014). Some sections of the Parautochthon were buried following a pressure gradient higher than classical Barrovian (Rubio Pascual et al., 2015). There are not isotopic age constrains for the earlier folds but the subsequent subhorizontal shearing has been dated at ca. 340 Ma (Dallmeyer et al., 1997).

The initial Variscan record of the Iberian Autochthon is rather similar. The first deformation produced a series of overturned to recumbent folds (Díez Balda, 1986; Macaya et al., 1991; Díez Fernández et al., 2013b), the vergence of which shows a radial pattern in relation to later oroclinal bends such as the Central Iberian arc (tectonic transport towards the Cantabrian Zone; Martínez Catalán et al., 2014). Early folding was accompanied by the development of a penetrative foliation under relatively low-grade metamorphic conditions. Subsequent subhorizontal shearing generated penetrative cleavages and phyllonites close to thrust faults. The basal contact of the Iberian Parautochthon has been interpreted as the most important of these thrusts (Ribeiro et al., 1990), and cuts previous folds (e.g., Marcos and Farias, 1999; Díez Montes, 2007). Similar thrusts exist within the Iberian Autochthon, but they seem to merge into a sole fault connected with the basal thrust of the Iberian Parautochthon (González Clavijo and Martínez Catalán, 2002; Dias da Silva, 2014). Dating of the axial planar foliation of the first folds yielded an age of ca. 354-347 Ma (Rubio Pascual et al., 2013a), which is in agreement with the broad age range previously obtained for the onset of Variscan deformation in this domain (ca. 370-342 Ma; Bea et al., 2009). Subsequent thrusting is responsible for the establishment of Viséan syn-orogenic basins at the advancing front of the Iberian Allochthon (Martínez Catalán et al., 2016), and considered coeval with subhorizontal ductile shearing in the Iberian Parautochthon, dated at ca. 340 Ma (Dallmeyer et al., 1997).

The juxtaposition of the Iberian Allochthon and Parautochthon over the Autochthon produced minimum pressures of up to ~0.9 GPa in several domains of the latter (Escuder Viruete et al., 2000; Rubio Pascual et al., 2013a), reaching up to ~1.1-1.4 GPa in some cases (Barbero and Villaseca, 2000; Rubio Pascual et al., 2015). Crustal thickening in other parts seems much less pronounced (typically in the chlorite-biotite zone), thus suggesting an inhomogeneous thrust stack.

3.4.6. South Iberian Autochthon

The Iberian Autochthon of SW Iberia occurs in two domains (Fig. 2). Both, the northeastern and southwestern occurrences, experienced Barrovian-type metamorphism during the first phases of Variscan deformation, with peak-pressures of about 0.6-0.7 GPa (González del Tánago and Arenas, 1991; Díaz Azpiroz et al., 2006). Although the type and geometry of coeval structures is not well-established for the northeastern domain, large-scale recumbent folding has been proposed for the early stages of tectonic evolution in the southwestern domain (Díaz Azpiroz et al., 2003). As to the timing of deformation, geochronological data suggest a rejuvenation of Neoproterozoic ages that might have started at ca. 390 Ma in the northeastern domain, with a remarkable maximum at ca. 360-351 Ma (Dallmeyer and Quesada, 1992). The age range of ca. 341-

337 Ma obtained for tectonic fabrics from the southwestern domain of the Iberian Autochthon (Dallmeyer et al., 1993) provides additional support for a Tournaisian-Viséean age as the onset of main Variscan deformation in the Autochthon.

3.5. Early-late Carboniferous transition (Viséan-Bashkirian)

3.5.1. NW and Central Iberian Allochthon, Parautochthon, and Autochthon

The record of early Barrovian-type metamorphism in the Autochthon and Parautochthon of NW and Central Iberia is overprinted by a flat-flying foliation that mantles and dominates the internal structure of granite- and migmatite-cored domes (Martínez Catalán et al., 2014). Usually bounded by extensional detachments, these domes alternate with large, open upright synforms and structural basins that have preserved the allochthonous complexes at their cores (Fig. 3). This extensional overprinting did not affect the Iberian Allochthon homogeneously. The upper structural levels, occupied by the Upper and Ophiolitic Allochthonous Units, escaped widespread ductile deformation and were only involved into discrete detachment faults cutting across previous tectonic boundaries (Martínez Catalán et al., 2002). Near the domes, some sections of the Basal Allochthonous Units were strongly affected by extensional deformation, including pervasive ductile shearing, faulting, and heat transfer from underlying granitic massifs (Gómez Barreiro et al., 2010a; Díez Fernández et al., 2012c).

Deformation took place under mid- to low-P conditions, and produced a low-grade crenulation cleavage in the lower parts of the suprastructure that turns quickly into a schistosity and a high-grade gneissic and migmatitic banding in the infrastructure. The extensional detachments separating these two structural levels of the crust are characterized by low-grade phyllonites and mylonites (Escuder Viruete et al., 1994, 1998; Díez Balda et al., 1995). Extensional ductile flow gave way to shearing and stretching of syn-kinematic granitoids (Díaz-Alvarado et al., 2012) as well as a profound distortion of previous structures via flattening (Díez Fernández et al., 2013b), sometimes resulting in large-scale isoclinal folding (Arango et al., 2013). Crustal attenuation varies depending on previous thickness, reaching decompression of the infrastructure up to the andalusite stability field in many cases. The kinematics of extensional flow reveals the non-coaxial character of deformation, which stretched the previous orogenic crust following divergent vectors, usually normal to those that governed previous crustal thickening in the hinterland (Díez Fernández et al., 2012c). Dating of migmatization and colder ductile deformation in different domes yields a Viséan-Bashkirian age of ca. 350-317 Ma (Escuder Viruete et al., 1998; Montero et al., 2004; Bea et al., 2006; Castiñeiras et al., 2008; Valle Aguado et al., 2008; Rubio Pascual et al., 2013a; López-Carmona et al., 2014), a range that confers a long-lasting nature to this lithosphere extension event. Despite that broad range, most of the domes seem to have experienced maximum thermal activity at ca. 340-320 Ma, whereas granite production abounds later (around 315-305 Ma; Valle-Aguado et al., 2005; Gutiérrez-Alonso et al., 2011; Martínez Catalán et al., 2014).

3.5.2. SW Iberian Allochthon and Autochthon

Southern Iberia is the site of Variscan metaluminous, alkaline to calc-alkaline magmatism, which is represented by gabbro-granodiorite suites including granites, granodiorites, diorites, tonalites, gabbros, gabbros with gabbroic and noritic cumulates, and mantle-derived ores (Capdevila et al., 1973; Aparicio et al., 1977). Some of the intrusives occurring just north of the Beja-Acebuches Ophiolite are characterized by a boninitic signature (Castro et al., 1996), an imprint that may be related to subduction, similarly to what is proposed for the calc-alkaline volcanism occurring near that region (Toca da Moura-Cabrela basin, Santos et al., 1990; Oliveira et al., 1991; Quesada et al., 1994; Onézime et al., 2003). The geochemical and isotopic characteristics of some mafic to intermediate (calc-alkaline) intrusives located farther north have been explained as a combination of high heat flow and contamination by pelitic crustal material of a mafic magma chamber formed in the mid-crust (Casquet et al., 2001; Salman, 2002; Tornos et al., 2005), or as the variable combination of mantle-derived and crustal-derived magmas (Romeo et al., 2006; Moita et al., 2009; Pereira et al., 2015). The age of all this earlier magmatic event spans a range between ca. 350-335 Ma (Dallmeyer et al., 1993, 1995; Casquet et al., 2001; Romeo et al., 2006; Jesus et al., 2007; Pin et al., 2008; Pereira et al., 2009, 2015; Cambeses et al., 2015) and is coeval with volcanism, contributing to Viséan syn-orogenic sedimentary basins (Armendariz et al., 2008; Pereira et al., 2012b; Oliveira et al., 2013).

The middle and lower structural levels of the tectonic pile in SW Iberia (Autochthon and Basal Allochthonous Units) were overprinted by high-T/low-P metamorphism and pervasive ductile deformation (González del Tánago, 1995; Díaz Azpiroz et al., 2002; Pereira et al., 2007). Peak metamorphic conditions reached the granulite facies along the southern border of the Iberian Autochthon, which constitutes a salient thermal anomaly in this region (Díaz Azpiroz et al., 2004). The general structure of these high-grade domains corresponds to granite- and migmatite-cored domes flanked by low-angle normal faults (González del Tánago, 1995; Pereira et al., 2009), where the Iberian Allochthon is mostly restricted to their hanging wall. Ductile deformation is widespread in the infrastructure, and was responsible for the generation of crenulation cleavages and mylonitic foliation.

The high-T/low-P deformation affected syn-orogenic granitoids (peraluminous granodiorite-granite suite) and previous folds, and was parted into subhorizontal and strike-slip shear zones parallel or subparallel to the structural trend of the orogen (Díaz Azpiroz et al., 2004; Pereira et al., 2009, 2012b, 2013). On the other hand, the Iberian Allochthon displays a number of low-angle normal faults that cut previous faults and folds (Azor et al., 1994b; Expósito et al., 2002). The age of these low-angle faults can be constrained to a range of ~345-315 Ma, as they are closely related to syn-orogenic deposits of uppermost Tournaisian to Viséan age but are affected by later upright folds during the Moscovian (Gabaldón and Quesada, 1986; Giese et al., 1994; Martínez Poyatos, 2002; Expósito Ramos, 2005). Dating of later syn-orogenic granitoids and migmatization in the lower structural levels of the Autochthon provides a similar time interval for the high-T event, which appears to show maximum thermal activity at ca. 335-325 Ma (Castro et al., 1999; Díaz Azpiroz et al., 2002; Pereira et al., 2009; Lima et al., 2011, 2013; Moita et al., 2015), somewhat younger than the earlier Variscan magmatism.

A special case is made for the Puente Génave-Castelo de Vide Detachment (see detailed description by Martín Parra et al., 2006), a low-angle normal fault that defines the northern contact of the Iberian Allochthon in SW Iberia (Díez Fernández and Arenas, 2015). Its fault line extends for a minimum of 400 km, whereas the fault zone is up to 150 m thick, dips to the SSW (30º mean dip), shows consistent top-to-the-SSW kinematics, and is made of graphite-bearing schist to the east. The movement of this fault should be later than the thrust tectonics that emplaced the Iberian Allochthon in the first place, estimated at ca. 340 Ma (see sections 3.4.2 and 3.4.4). A maximum age for the Puente Génave-Castelo de Vide Detachment is provided by Variscan granitoids and related rocks deformed into its fault zone (ca. 331 Ma; Larrea et al., 1999), while its minimum age is given by Moscovian granites cutting across the fault zone and dated at ca. 314 Ma and ca. 307 Ma (Carracedo et al., 2009; Solá et al., 2009).

Seismic and magnetotelluric data obtained in SW Iberia reveal the existence of a ~140 km long reflective and low-resistivity body located at the mid-crust, the Iberian Reflective Body (IRB; Simancas et al., 2003; Muñoz et al., 2008). IRB shows variable thickness (up to 5 km), making its shape wavy in some places. This body displays its maximum thickness to the SSW, wedges to the NNE, and dips about 5º to the SSW. Current ideas on the origin of IRB defend a hybrid (tectonic-magmatic) model, i.e. a layered, mantle-derived mafic intrusion in a detachment level (Simancas et al., 2003; Carbonell et al., 2004). This hypothesis may explain the Variscan alkaline, calc-alkaline, and metaluminous magmatism of the region (Carbonell et al., 2004; Cambeses et al., 2015), but no surface expression of the detachment level has been proposed.

The prolongation of the Puente Génave-Castelo de Vide Detachment to the south, following the seismic markers of the region (Simancas et al., 2003; Martínez Poyatos et al., 2012), reaches the northern edge of IRB (Fig. 3), which would account for the geophysical expression of this huge extensional shear zone, or at least that of a layered magmatic body shaped into the shear planes of that shear zone, either after or during intrusion. This correlation suggests that the extensional shear zone associated with the detachment widens with depth, conferring a listric geometry to the detachment, and reinforcing the influence of this fault on the current (down thrown) position of the Iberian Allochthon in SW Iberia (Díez Fernández and Arenas, 2015). Moreover, the prolongation of this fault further to the south, using as a guide the seismic markers (Simancas et al., 2003), unveils a potential structural correlation with the mid-crustal root of the basal decollement from which the south-directed thrusts of the South Portuguese Zone foreland may have derived. The age of these thrusts is uppermost Viséan to Moscovian (~330-307 Ma; Silva et al., 1990), a time interval that matches the chronology of the Puente Génave-Castelo de Vide Detachment (~330-314 Ma).

3.5.3. Beja-Acebuches Ophiolite and related rocks

There is a negligible age difference between the mafic protoliths of the Beja-Acebuches Ophiolite (ca. 340-332 Ma; Azor et al., 2008) and the subsequent ductile shearing that affected this ensemble at ca. 342-328 Ma (Dallmeyer et al., 1993; Castro et al., 1999). This ophiolitic unit is strongly overprinted by left-lateral, top-to-the-SW ductile shearing (South Iberian shear zone; Crespo-Blanc and Orozco, 1988, 1991) developed under low-P and mid- to low-T metamorphic conditions (Bard and Moine, 1979; Castro et al., 1996), although some of its sections seem to preserve early top-to-the-north shearing formed under a higher metamorphic grade (Fonseca and Ribeiro, 1993).

The regional fabrics developed throughout the Basal Allochthonous Units located just north of the Beja-Acebuches Ophiolite (age estimated at ca. 358-350 Ma, see section 3.4.4) and the high-grade rocks of the South Iberian Autochthon are affected by south- and southwest-verging folds and thrusts, which show a left-lateral component comparable to the structure of the Beja-Acebuches Ophiolite (Díaz Azpiroz et al., 2003;

Araújo et al., 2005; Borrego et al., 2005; Ponce et al., 2012). Folding was developed under greenschist facies conditions and produced additional flat-lying crenulation cleavages in the region.

3.6. Late Carboniferous (Bashkirian-Gzhelian)

3.6.1. Oroclines of the Iberian Massif

Some Variscan and pre-Variscan linear features that mark the structural trend of the Iberian Massif are curved into the shape of a plate-scale vertical fold to define a couple of oroclinal bends, namely the Ibero-Armorican arc and the Central Iberian arc (Fig. 1; Martínez Catalán, 2011). These arcs are delineated by some tectonostratigraphic domains of the Iberian Massif, by the first Variscan folds of the Iberian Autochthon, by low- and high-amplitude magnetic anomalies sourced from an unexposed crystalline basement (Aerden, 2004; Martínez Catalán, 2012), and by paleocurrents in Ordovician strata (Shaw et al., 2012). The structural grain and terranes of the Iberian Allochthon do not display such curved patterns for the case of the Central Iberian arc, either in NW or SW Iberia. However, the Iberian Allochthon occupies the core of the Central Iberian arc in the NW and flanks that orocline to the SW (Fig. 1). The southern boundary of this arc runs along the Puente Génave-Castelo de Vide Detachment, which appears to cut it at a high angle.

The nucleation of the Central Iberian arc is considered to have occurred later than ca. 335 Ma (age of the youngest folds affected by the arc), whereas its closure occurred at ca. 315-305 Ma (Martínez Catalán, 2011, 2012; Martínez Catalán et al., 2014). The age of the Puente Génave-Castelo de Vide Detachment allows further constrains on the age of this orocline, most of the vertical folding related to which should be older than the detachment (ca. 330-314 Ma). The Ibero-Armorican arc is slightly younger (Martínez Catalán, 2011), its age being constrained by means of paleomagnetic data at ca. 304-295 Ma (Weil et al., 2010).

3.6.2. Strike-slip shear zones of the Iberian Massif

Except for the Ibero-Armorican arc, all the previous Variscan record is variably affected by a series of intracontinental, strike-slip shear zones and related structures (Martínez Catalán, 2011). NW, Central, and SW Iberia exhibit a combination of dextral and sinistral shear zones (Fig. 2). Yet, left-lateral movements dominate in SW Iberia (Burg et al., 1981; Crespo-Blanc and Orozco, 1988; Pereira and Silva, 2001; Pérez-Cáceres et al., 2015a), while the major strike-slip systems in NW and Central Iberia are dextral in most of the cases (Iglesias Ponce de Leon and Choukroune, 1980; Ribeiro et al., 1980).

The strike-slip systems include zones with variable intensity of shearing. Various types of subvertical mylonites, and pervasive ductile deformation in their cores, give way to more spaced subvertical crenulation cleavages, overprinting the previous record at both sides of the shear zones. At a larger scale, the lateral displacements of these strike-slip systems deflect previous geological features such as contacts or tectonic fabrics, whereas the subhorizontal shortening experienced by the two blocks of the shear zone is accommodated by upright regional folds, most of which amplifies former extensional domes and structural basins. This is the case of the Iberian Allochthon, which is located in the core of open structural basins and it is surrounded by migmatized basement cropping out in structural domes (e.g., Padrón dome; Díez Fernández et al., 2012c). During the later stages of transcurrent shearing, many of the strike-slip systems evolved to subvertical faults with lateral and dip-slip motion.

The strike-slip shear zones show different relationships with the development of the Central Iberian arc. However, the trace of the upright folds associated with strike-slip deformation shows high correlation with the axial trace of this orocline, suggesting that eventual tightening of this vertical fold was accomplished by shortening related to strike-slip shearing (Martínez Catalán, 2011). Although not all the shear zones are coeval sensu stricto, their age and that of related folding as a whole is very consistent throughout the Iberian Massif, and ranges between ca. 315-305 Ma (Capdevila and Vialette, 1970; Martínez Poyatos et al., 1998; Rodríguez et al., 2003; Valle Aguado et al., 2005; Gutiérrez-Alonso et al., 2015).

3.7. South Portuguese Zone

The early Variscan deformation that is observed in the South Portuguese Zone is found in its northern section (Pulo do Lobo Unit). It consists of south verging folds developed under low-grade metamorphic conditions (Silva et al., 1990; Martínez Poza et al., 2012). Such folding affected both Silurian-Devonian and Frasnian series (Pereira et al., 2008b; Braid et al., 2011). Previous age estimations considered this deformation as Upper Devonian (Giese et al., 1999).

A subsequent phase of deformation in this region produced north- to south-southwest-verging folds and involved younger sedimentary series deposited (discordantly) on top of the previous folds (Silva et al., 1990, 2013; Fonseca, 2005; Martínez Poza et al., 2012). This younger series includes strata with ages ranging from upper Famennian, Tournaisian, and up to Viséan, as evidenced by fossil (Pereira et al., 2008b; Matas et al., 2015b) and detrital zircon data (youngest age population at ca. 347 Ma; Braid et al., 2011). Thus, the age of the first folds in the Pulo do Lobo Unit can be better constrained between ca. 380-359 Ma.

The South Portuguese Zone was affected by extension and related bimodal magmatism during the Tournaisian (~356-346 Ma; Barrie et al., 2002; Dunning et al., 2002; Rosa et al., 2008; Valenzuela et al., 2011). Later deformation progressed in a thin-skinned fashion up to the Moscovian and propagated from the Beja-Acebuches Ophiolite to the south via thrusts and related folds (~330-305 Ma; Silva et al., 1990, 2013). According to geochronological data of lithologies affected and non-affected by this later phase of deformation, south-directed thrusting must have been older in the northern part of the South Portuguese Zone (Pulo do Lobo Unit), where its age is estimated at ca. 345-335 Ma (Gladney et al., 2014). These thrusts represent a major tectonic inversion in the region and cut across the north- to south-southwest-verging folds of the Pulo do Lobo Unit (Martínez Poza et al., 2012). If so, the age of the latter folds should be Viséan (~347-335 Ma).

Some sections of the pre-Upper Devonian series of the South Portuguese Zone experienced Variscan high-P and low-intermediate-T metamorphism (Rubio Pascual et al., 2013b). These series are affected by the first phase of deformation recognized in this zone (ca. 380-365 Ma) and its Carboniferous cover does not show such tectonometanorphic imprint. Therefore, the high-P metamorphism must be either coeval or previous to the early south-directed folding (Lower to Middle Devonian?).

4. Tectonic evolution of Variscan Iberia: model and discussion

Lateral tectonics played a role on Variscan deformation affecting the Iberian Massif, impossible to quantify in full, but qualifiable. Many of the major shear zones that occur in SW Iberia (either flat-lying or subvertical) include a left-lateral component. Consequently, the position of SW Iberia -or that of the several blocks associated with strike-slip structures- relative to Central and NW Iberia before orogenesis should be located more to the northwest in present-day coordinates, i.e. west of the NW Iberian section s.l. (either southwest, purely west, or northwest). Such general assumption has been made to construct the composite section shown in Figure 3, and is strongly supported by semi-quantitative estimations on the left-lateral displacement accumulated in SW Iberia through the Variscan orogenesis (e.g., Burg et al., 1981; Pereira et al., 1998; Pérez-Cáceres et al., 2015b). Despite such restoration along-strike may result imprecise (e.g., lateral intracontinental displacements accumulated in some particular faults might have exceeded several hundreds of kilometers), the impact on the qualitative reconstruction of Variscan tangential tectonics is probably minor, as suggested by the synchrony of tectonic events across the Iberian Massif (see below).

Figure 4a shows a simplified restoration of Variscan thrusts and strike-slip shear zones and provides a general picture of the pre-collisional paleogeography across the margin of Gondwana. This reconstruction acknowledges the following ideas on the Variscan and pre-Variscan evolution of the Iberian Massif: (i) the recognition of the Iberian Allochthon across the Coimbra-Córdoba shear zone (Díez Fernández and Arenas, 2015); (ii) the Upper Allochthonous Units represent a section of the margin of

Gondwana that was rifted from its mainland during the Cambrian-Ordovician (Gómez Barreiro et al., 2007), but remained attached or geographically close to it (at least) up to the Lower Devonian (Robardet, 2003; López-Guijarro et al., 2008; Arenas et al., 2014b); (iii) the Cambrian-Ordovician rifting shaped the margin of Gondwana into a series of continental microblocks connected by stretched lithosphere (Díez Fernández et al., 2015); (iv) the onset of Variscan deformation is Lower Devonian (ca. 410-395 Ma); (v) the suture zone represented by the Allochthonous Ophiolitic Units accounts for the closure of an ephemeral oceanic basin opened after the onset of Variscan deformation (i.e. a second-order suture zone; Arenas et al., 2014a); and (vi) the Beja-Acebuches Ophiolite is the suture of a transient oceanic basin that separated most of the Iberian Gondwana from Laurussia during the early Carboniferous (Azor et al., 2008). Finally, the initial Variscan evolution of the South Portuguese Zone may not be directly related to that of the rest of the Iberian Massif, i.e. this section of putative Laurussia did not face the sections of Gondwana preserved in Iberia until Upper Devonian-Carboniferous times (Braid et al., 2011). The evolution of pre-Upper Devonian sequences of the South Portuguese Zone might be associated with NeoAcadian events recorded in Meguma (Van Staal et al., 2009). The palynological assemblages (Pereira et al., 2006b) and detrital zircon populations (Pereira et al., 2012a) found in syn-orogenic deposits at both sides of the Beja-Acebuches Ophiolite indicate that the Iberian Allochthon and its relative autochthon were close to the South Portuguese Zone during the Upper Devonian and early Carboniferous, so large distances along-strike (if any) are not expected between these two domains after closure of major oceanic basins by the Lower Devonian, such as the Rheic Ocean.

The data compilation presented in section 3 shows the synchronous character of compressional and extensional deformation events across the Iberian Massif during the Variscan orogenesis (see also Table 1). Although timing, geometry, and tectonic polarity coincides in many cases after unfolding the late oroclinal bends, the following evolutionary model is also aimed to integrate both similar and contrasted structural and metamorphic record by using age reference lines. In order to keep our model as realistic as possible, sketches presented in Figures 4 and 5 show the geometry and location of actual structures of the Iberian Massif, a reference of which has been given in section 3. The series of sketches culminates with Figure 3, which represents a synthetic cross-section of the Iberian Massif today and therefore a good approximation to the eventual structure after Variscan deformation.

4.1. Initiation of Variscan Orogeny (Fig. 4b)

The onset of the Variscan orogeny took place in the Lower Devonian and produced a fragmentary record across the Iberian Massif. The first phases of deformation related to the interaction of Laurussia and Gondwana can be observed in the Upper Allochthonous Units of NW and SW Iberia, although the structures and associated metamorphism are strikingly different depending on the region.

A stratigraphic gap and basin instability are the first hints on deformational processes heralding the Variscan orogenesis in Iberia (ca. 420-410 Ma). These can be tracked in the Upper Allochthon of SW Iberia, and are followed by the formation of a SW-verging train of recumbent folds and thrusts in a sinistral transpressional regime (Expósito et al., 2002). This major crustal thickening event took place in the outermost section of the margin of Gondwana during the Pragian-Emsian (ca. 410-395 Ma), following early events of syn-orogenic deposition associated with basement denudation in Lochkovian times. Top-to-the-SW kinematics (in present-day coordinates) of Lower Devonian structures implies underthrusting to the NE, i.e. subduction of Laurussia under Gondwana, where a magmatic arc developed in Emsian-Eifelian times (Silva et al., 2011). Shortening of the upper plate would be favored by the migration of the subduction hinge toward the upper plate (Doglioni et al., 2007), thus allowing the development of a shallower downgoing slab, as expected for upper plates of continental origin (Lallemand et al., 2005).

In Lower Devonian, a neck of stretched peri-Gondwanan lithosphere located inboard failed mechanically under the compressive regime derived from the interaction between Laurussia and Gondwana, thus creating an intra-Gondwana subduction zone for accommodating superimposed shortening throughout the continental margin. Intra-plate subduction was probably favored by a backstop effect exerted by thick (SW-verging folds) and more buoyant Variscan lithosphere located toward the Gondwana-Laurussia suture zone (Rheic suture). Lower Devonian continental subduction was oblique (dextral; Ábalos et al., 2003) and progressed up to high-P conditions under outboard sections of Gondwana (Albert et al., 2015b).

Some sections of the Upper Iberian Allochthon, such as the Obejo-Valsequillo Domain, seem to have escaped from penetrative Lower Devonian deformation. This is in agreement with its intermediate position across the Upper Allochthon inferred from restoration of Variscan thrusts (Fig. 4a), and confers a remarkable microplate-like entity to the whole Upper Iberian Allochthon during this stage. The apparent lack of widespread shortening affecting the upper plate of this subduction zone (uppermost allochthonous units) suggest that the subduction hinge remained relatively stationary, as expected for the onset of subduction zones (Doglioni et al., 2007). In this regard, sinistral lateral components acting over the external parts of the Upper Allochthon (sinistral transpression during SW-vergent folding), combined with coeval dextral movements affecting its inboard sections (oblique continental subduction), depict an overall setting of northwards escape tectonics for the case of the Upper Allochthon “microplate”.

4.2. Opening of a Devonian intra-Gondwana basin (Fig. 4c)

The geochemical signature of the Lower-Middle Devonian rocks of the Ophiolitic Allochthonous Units indicate that there were physical conditions for the opening of a marine basin following Late Devonian continental subduction (Arenas et al., 2014a). This interpretation ties into the coeval extensional record of the Iberian Autochthon (alkaline magmatism and basin subsidence; Gutiérrez-Alonso et al., 2008). Simple orthogonal restoration of the allochthonous pile reveals the location of the spreading center of this basin between the Upper Allochthonous Units and the pair constituted by the Cambrian Allochthonous Ophiolites and the Basal Allochthonous Units. According to the stratigraphic record, neither sediments were laid down at that time in the continental margins of that basin, nor do thick sedimentary series exist within the Lower-Middle Devonian ophiolites. Whether or not the lack of Middle Devonian sedimentary and volcanic rocks in the continental counterparts of the Iberian Allochthon is associated with deformation and denudation (emerged areas?), the absence of such stratigraphic record in all this domain may indicates a broad thermal uplift in relation to ridge inception.

The Lower Devonian high-P metamorphic belt that is preserved in the lowermost structural position of the Upper Allochthonous Units of NW Iberia was developed at ca. 410-390 Ma. This is virtually the same age (somewhat older) as that of the mafic protoliths of the Lower-Middle Devonian Allochthonous Ophiolitic Units (ca. 400-395 Ma), which are tectonically juxtaposed right underneath that high-P metamorphic belt. Exhumation of high- to ultra-high-P metamorphic rocks today is observed in regions subjected to high-rates of lithosphere extension and coeval ocean basin formation, such as the Woodlark rift (Davies and Warren, 1988; Wallace et al., 2004). Previously deep-seated high-P metamorphic rocks in these cases can reach lower crustal levels, and then the upper crust, in less than 3 and 5 Ma, respectively (Gordon et al., 2012). On the grounds of modern analogues, we propose that initial exhumation of the Lower Devonian high-P metamorphic rocks was strongly controlled (probably fuelled) by the opening of the intra-Gondwana oceanic basin shortly after their burial. Extension of the upper plate, triggered by a subduction hinge migrating away from the upper plate (Doglioni et al., 2007), could have facilitated both a fast exhumation and the opening of a basin, which could have then evolved as one of pull-apart type under dominant transcurrent movements (Arenas et al., 2014a).

4.3. Closure of the intra-Gondwana Devonian basin (Fig. 4d)

In Upper Devonian times, renewed convergence between Gondwana and Laurussia led to the closure of Middle Devonian oceanic domains. The tectonic polarity for this event was ruled again by thicker Variscan crust located outboard mainland Gondwana, i.e. subduction to the W and SW in present-day coordinates. Understacking of young (Devonian) oceanic crust under the Upper Allochthonous Units was followed by accretion of older (Cambrian-Ordovician) tracts of transitional crust (Arenas et al., 2007a), and then by subduction of continental crust at ca. 380-370 Ma (Basal Allochthonous Units; Martínez Catalán et al., 1996). Regarding the latter process, insertion of more buoyant lithosphere under the (previous) Lower Devonian high-P metamorphic belt caused further exhumation of the bottom members of the Upper Allochthonous Units via tectonic denudation, which coupled to east-verging folding in response to simple shearing at the base of the upper plate (Gómez Barreiro et al., 2007). These processes continued the initial decompression experienced by the Lower Devonian high-P metamorphic rocks under the Lower-Middle Devonian rifting setting.

The Upper Devonian continental (intra-Gondwana) subduction system was formed with an angle of inclination between 15º and 30º (Alcock et al., 2005) and absorbed ongoing dextral convergence (Díez Fernández et al., 2012a). Initial exhumation within this system (Basal Allochthonous Units) was driven by crustal-scale ductile thrusting directed to the Gondwana mainland at ca. 370-360 Ma (e.g., Fervenza Thrust). Tangential deformation at this stage was concentrated on the upper part of the subducted plate, and it was likely coeval with further sinking of continental lithosphere. Ductile thrusting was assisted by erosion in the upper plate and it also forced the generation of a master, and/or a series of normal faults on top of the overthrusting high-P nappes (Díez Fernández et al., 2011). In this regard, top-to-Laurussia shear sense components of Famennian age affecting the Basal Allochthonous Units (e.g., older than ca. 358 Ma; Rosas et al., 2008) may account for normal, flat-lying shearing at the onset of decompression in response to upthrusting and extrusion of deep-seated continental nappes.

4.4. Development of the Iberian Allochthon (Figs. 5a and 5b)

Continental convergence remained during the latest Devonian and early Carboniferous (ca. 360-350 Ma) and was absorbed by W to SW (present-day coordinates) underthrusting of Gondwanan crust. Superimposed shortening probably created new contractional shear zones below the Upper Devonian subduction-exhumation channel. The onset of deformation in the Parautochthon and Autochthon of NW Iberia represents the transition from a purely continental subduction setting (recorded in the Basal Allochthonous Units) to a continent-continent collisional scenario.

At this stage, the progressive diminishing of initial high-P gradients down structure through the Variscan nappes favors a model of underthrusting of progressively thicker continental lithosphere. Protracted accretion of more buoyant continental crust to the base of the Upper Devonian subduction-exhumation system led to its progressive rotation about an horizontal axis and, consequently, to its deactivation. The early response to that exhumation process was the nucleation and propagation to the Gondwanan foreland of a train of recumbent folds within the high-P metamorphic belt and in its relative autochthon (Iberian Parautochthon and Autochthon). Convergence at this stage was also accompanied by dextral lateral movements, as indicated by tectonic fabrics associated with fold development in NW Iberia (Díez Fernández and Martínez Catalán, 2012) and probably the top-to-the-NW kinematics (Azor et al., 1994b) that dominated the exhumation process in the currently NE-dipping (originally SW-dipping) Central Unit.

Continuous constriction of the mantle wedge over the former subduction channel produced an eventual mechanical coupling between the Basal Allochthonous Units and the rest of the Iberian Allochthon, which, from this point on, would absorb general shear deformation associated with ongoing underthrusting more efficiently (ca. 350-340 Ma). In the Upper Allochthonous Units, the existence of a train of recumbent folds with comparable age, trend and vergence than those observed in the Basal Allochthonous Units supports this idea. Fold propagation across the Upper Allochthonous Units progressed toward Laurussia, reaching the lower parts of the Obejo-Valsequillo Domain shortly afterwards. However, some of those folds in the Upper Allochthonous Units were probably nucleated during the Upper Devonian, prior to the aforementioned mechanical coupling. This may be the case of the recumbent folds affecting the Lower Devonian high-P metamorphic belt exposed in NW Iberia, for which subsequent general shear after coupling would have produced additional amplification of their initial (overturned?) recumbent geometry. All these processes attest for an orogenic shortening that is propagating more pervasively into the lower plate, but that was already affecting the upper plate since the onset of continental subduction. This transition is observed in advanced stages of continental collision following a stage more dominated by subduction (Doglioni et al., 2006, 2007).

Large-scale ductile thrusts, such as the Lalín-Forcarei Thrust and the basal thrust of the Iberian Parautochthon, represent advanced stages of the accretion of mainland Gondwana under the Iberian Allochthon and Parautochthon, respectively (ca. 340 Ma). These structures moved the Iberian Allochthon onto inner domains of Gondwana in the first place (Martínez Catalán et al., 1996), and are responsible for the juxtaposition of the Iberian Parautochthon onto the Iberian Autochthon (Ribeiro et al., 1990). The prolongation of these broad ductile shear zones toward Laurussia is possible through the (top-to-the-Cantabrian Zone) strongly sheared sequences and tectonic fabrics that dominate the internal structure of the Basal Allochthonous Units of Iberia. The general mylonitic character of these fabrics accounts for pervasive ductile deformation along the lower structural levels of the Iberian Allochthon. A progressive ductile drag and stretching of the Basal Allochthonous Units during underthrusting would have conferred its apparent far-traveled nature to the Iberian Allochthon. During this process, the Allochthonous Ophiolites may have acquired some of its tectonically dismembered appearance. Such a broad ductile drag explains the great lateral continuity of the Upper Devonian high-P metamorphic belt across the Iberian Massif (Díez Fernández and Arenas, 2015), as well as the generation of unusually large allochthonous terranes like the Iberian Allochthon in a collisional orogeny.

Underthrusting continued during the Tournaisian-Viséan (ca. 340-330 Ma). However, the absence of regional, east to northeast verging folds of that age in the Upper Allochthonous Units located south of the Coimbra-Córdoba shear zone, suggests that simple shearing at the base of the Iberian Allochthon (if any) did not trigger folding in its outboard-most sections. In turn, continental convergence at this stage was accommodated by the nucleation of discrete reverse faults cutting across the upper plate. Among them we find the set out-of-sequence thrusts that bring pieces of an underlying suture zone to internal sections of the Upper Allochthonous Units, in the Obejo-Valsequillo Domain (Espiel Thrust; Apalategui and Pérez-Lorente, 1983; Martínez Poyatos et al., 2001). These type of faults have been also described in NW Iberia (Martínez Catalán et al., 2002), and altogether they depict an overthrusting event that transported most of the Iberian Allochthon further inboard Gondwana, thus enhancing its far-traveled nature. According to geological data, this out-of-sequence thrusting event was accomplished by taking the Upper Devonian, intra-Gondwana suture zone and its major tectonic boundaries as primary detachment levels (e.g., Díez Fernández et al., 2013a).

Due to limited structural and tectonostratigraphic record, the role of Laurussia in the course of all this continental accretion remains uncertain. However, the development of coeval folds (south-) vergent to its mainland in the South Portuguese Zone (Martínez Poza et al., 2012) suggests that the backstop effect exerted by the Rheic suture between Gondwana and Laurussia –dipping to Gondwana since Lower Devonian times– remained active up to the lowermost Carboniferous, at both sides of the suture zone. In this scenario, the late amplification and development of south-directed thrusts affecting the south-verging folds of the Upper Allochthonous Units of SW Iberia might have occurred during the Upper Devonian through the early Carboniferous.

In the South Portuguese Zone, folding of sedimentary series postdating the onset of the Rheic suture probably represents backs and forths in the far-field interaction between Gondwana and Laurussia, as demonstrated by alternating compressional and extensional events affecting the inner sections of Gondwana (see evolutionary model). Subsequent folds verging toward Gondwana attest to a switch in the inclination of the reference shear planes. The age and structural polarity of these folds fit the timing and kinematics of ongoing crustal underthrusting under the Iberian Allochthon. Thus, we speculate that understacking of Gondwanan lithosphere might have surpassed and interplayed with the Rheic suture by the Tournaisian-Viséan.

Some of the lithosphere extension and related magmatism observed in the South Portuguese Zone (not represented in Fig. 5) occurred in the early Carboniferous (ca.

360-330 Ma). At that age, a progressive Laurussia-directed underthrusting of Gondwanan lithosphere must have produced an incremental constriction in the mantle wedge resting over the Upper Devonian continental subduction system. Such constriction implies a lateral extrusion of that portion of mantle toward Laurussia, i.e. toward the South Portuguese Zone. A readjustment like this in the mantle lithosphere under Gondwana could have led to diffuse asthenosphere upwelling, extension, and magmatism in the Laurussian side of the orogen during the course of ongoing convergence. Some of the lowermost Carboniferous magmatism in SW Iberia might be explained by this large-scale mechanism, which might also have contributed to subsequent thermal anomalies in that region.

4.5. Opening of the Beja-Acebuches basin and the onset of orogenic collapse (Fig. 5c)

The Viséan is a stage of major changes in the dynamics of the Variscan orogen. A former period ruled by convergence between Gondwana and Laurussia gives way to a new phase characterized by intra-orogenic extensional activity (Simancas et al., 2006; Pereira et al., 2012b). Two main processes stand out: the opening of the Beja-Acebuches basin (named after the Beja-Acebuches Ophiolite) and the start of orogenic gravitational collapse.

Though speculative, a look into the mantle topography before the switch to an extensional regime may offer a tectonic perspective about the origin of the latter. In our model, Viséan extension followed the underthrusting of Gondwanan crust toward Laurussia. Regardless of the amount of crustal material seated under the Iberian Allochthon, such tectonic polarity favors a thicker crustal root toward Gondwana, i.e. a higher mantle topography toward Laurussia (Fig. 5b). The constriction and lateral extrusion of the mantle resting below peri-Gondwana, in response to a protracted underthrusting of mainland Gondwana, may have also favored such a higher mantle topography.

The numerous occurrences of mafic to intermediate magmatism of Tournaisian-Viséan age in SW Iberia have been related to a large-scale extensional event (Simancas et al., 2003), in which the opening of the Beja-Acebuches basin, floored with mafic and some ultramafic rocks, represents an eloquent proof of lithosphere necking (Azor et al., 2008). In this process, the mantle certainly played a role, as shown by varied petrological evidence from coeval mafic to intermediate magmatism (e.g., Moita et al., 2009; Pereira et al., 2009, 2015; Cambeses et al., 2015). But there is no consensus on whether or not extension was triggered by thermal anomalies in the mantle (e.g., plumes; Simancas et al., 2006), by subduction (Bard, 1977; Quesada et al., 1994), by a process of transcurrent slab break-off after collision (Pin et al., 2008), or due to intracontinental rifting in a transtensional (pull-apart?) setting (Bard, 1977; Azor et al., 2008; Cambeses et al., 2015). A higher mantle topography toward Laurussia, as suggested before, not only would imply a major thermal anomaly in the region, but also explains the location of the Beja-Acebuches basin, which could have been opened using this broad “lithosphere orogenic neck” as a trigger. The upwelling of the asthenosphere was probably responsible for the decompressional melting of the lithospheric mantle, which had already been metasomatized by a subducted slab (Rheic Ocean) leading to the generation of mafic parental magmas (Pereira et al., 2015). Simultaneously, the underplating of mafic magmas caused partial melting of continental crust. Time-equivalent and mantle-influenced magmatic activity in other parts of the Iberian Massif was apparently not related to the opening of additional oceanic basins (e.g., Pyrite belt;

Mitjavila et al., 1997; Martin-Izard et al., 2016), so lithosphere extension at this stage was probably heterogeneous (additional minor lithosphere necks might have existed) and/or the far-field influence of the mantle high here suggested was not restricted to the Gondwana-Laurussia suture.

A simple restoration of Variscan thrusts in SW Iberia indicates that the opening of the Beja-Acebuches basin cut off the tectonic pile culminated by the Iberian Allochthon. This implies that the former Rheic suture between Gondwana and Laurussia in Iberia became part of a different plate than the rest of the orogen. The actual location of that suture zone is a matter of debate, because traces of the Rheic Ocean crust are yet to be found. The terrane capable of sourcing sediments dispersed on both sides of the Rheic suture is interpreted to have been completely removed by erosion in SW Iberia (Pereira et al., 2012a). In this regard, erosion of rift shoulders during the opening of the Beja-Acebuches basin and/or subsequent crustal understacking during its closure are two likely mechanisms capable of hiding most of the previous orogenic record associated with the Rheic suture. This is particularly true for the lower plate to the north-dipping suture of the Beja-Acebuches basin. Interestingly, that plate is the one where the initial suture between Gondwana and Laurussia was located after the intracontinental rifting that gave way to the Beja-Acebuches basin. Therefore, even if separating Gondwanan and Laurussian domains, the suture of the Beja-Acebuches basin should be considered as a reworked one.

Lessening of the gravitational disequilibrium created after the transference of the Iberian Allochthon onto the Gondwana mainland was conducted by extensional detachments in the upper part of the tectonic pile. The former understacking, thickening and pressurization of its relative autochthon favored the thermal disequilibrium of the orogenic crust. This crust started to flow laterally and melt fertile crustal layers as response to thermal reequilibration, giving rise to migmatitic domes across the Iberian Autochthon and to felsic (crust-derived) magmatism after a period of thermal maturation (Alcock et al., 2009; Pereira et al., 2009; Martínez Catalán et al., 2014). The consequent crustal extension forced the mantle to rise to compensate for lithosphere attenuation. The opening of the Beja-Acebuches basin was roughly coincidental with the initiation of the gravitational collapse of the Variscan thrust pile. With this perspective, we think that positive feedback probably existed between the early stages of thermal and gravitational reequilibration of the orogen and the lithosphere extension that led to the opening of that basin.

Thermal models suggest that the overriding of the Iberian Allochthon can explain alone the extensional collapse of the Variscan crust and the generation of abundant (crust-derived) magmatism of Serpukhovian-Bashkirian age (Alcock et al., 2015). Yet, the mafic to intermediate (alkaline, calc-alkaline, and metaluminous) magmatism of SW Iberia is slightly older (Tournaisian-Viséan) and shows fair mantle input. Consequently, the mechanism(s) in place for the development of such mafic to intermediate magmatism (e.g., plumes, transient arc, slab break-off, and/or orogenic transtension) could have also contributed to the collapse of the orogenic hinterland in the first place. In this regard, the existence of a higher mantle topography toward Laurussia would explain not only older (mantle-induced?) extensional activity in SW Iberia but also larger (petrological) mantle contributions in the lack of a thick crustal root underneath this region. Remarkably, mantle contributions diminish toward the north and northeast of the Iberian Massif, where extensional syn-orogenic magmatism is slightly younger and is clearly dominated by crustal sources (e.g., Villaseca et al., 1998). That petrological and geochronological trend across the orogen accords well with the existence of a thicker crustal root toward Gondwana, supports the existence of irregular mantle topography before extension, and is consistent with protracted underthrusting of Gondwanan crust toward Laurussia.

4.6. The collapse of Variscan orogenic crust (Fig. 5d)

Once the cohesion of the orogen was lost to its thermal re-equilibrium, the gravitational collapse gained importance through Serpukhovian-Bashkirian times. This stage is characterized by extensive felsic magmatism, which occurred preferentially in areas subjected to severe denudation, i.e. under the Iberian Allochthon. Extensional faults in the upper crust drove further tectonic denudation (e.g., Pico Sacro and Puente Génave – Castelo de Vide detachments), whereas lower crustal flow distributed vertical flattening and lithosphere attenuation across the orogen. Regions dominated by felsic magmatism of Serpukhovian-Bashkirian age occur in the core of dome structures (e.g., Padrón dome), revealing the contribution of diapiric flow to the reequilibration process.

The extensional collapse of the Variscan orogen has been classically viewed as a syn-convergent process (Franke, 2000). One of the main reasons sustaining this idea in the case of Iberia is that extension was preceded and followed by indisputable phases of continental convergence (Martínez Catalán et al., 2002). Indeed, thermal and gravitational re-equilibration were not acting alone on the overthickened crust. The onset of the Central Iberian arc has been framed in this stage too (Martínez Catalán, 2012). Orogen-parallel extensional flow dominated the gravitational re-equilibration of the hinterland, and has been interpreted as the result of ongoing oblique plate movements in the course of orogenic collapse (Díez Fernández et al., 2012c). Additionally, the development of strike-slip shear zones coeval with extension, furthers the role exerted by lateral tectonics at this stage (Pereira et al., 2009). Any of the aforementioned structural records could have been developed either under plate-scale transtension and/or transpression. There is, however, a major geodynamic event on which convergence setting at plate scale relies during Serpukhovian-Bashkirian times, the closure of the Beja-Acebuches basin.

Dipping under the Iberian Allochthon and Autochthon, the Beja-Acebuches Ophiolite has an age of accretion (ca. 342-328 Ma) that matches the age of the onset of Variscan orogenic collapse. Hence, the subduction of oceanic lithosphere formed in the Beja-Acebuches basin provides a convergence geodynamic setting under which ongoing gravitational collapse must have evolved. The closure of the oceanic domain represented in the Beja-Acebuches Ophiolite has been widely considered as related to a subduction process (e.g., Munhá et al., 1986; Eden and Andrews, 1990; Silva et al., 1990; Fonseca and Ribeiro, 1993; Quesada et al., 1994; Simancas et al., 2003; Díaz Azpiroz et al., 2006; Braid et al., 2010), although part of the tectonic evolution of this domain could be also related to an obduction event (Pérez-Cáceres et al., 2015a). Recent findings of lawsonite-bearing rocks in the Pulo do Lobo Unit (Rubio Pascual et al., 2013b) suggest the formation of a pressure-dominated metamorphic belt during the accretion of the Beja-Acebuches Ophiolite, thus providing additional support to models that acknowledge subduction as a driving mechanism during the closure of the Beja-Acebuches basin.

The development of SW-verging folds and thrusts affecting the previous record, both in the Iberian Allochthon and Autochthon of SW Iberia during the Serpukhovian-Bashkirian, can be explained by a NE-directed tectonic polarity for the closure of the Beja-Acebuches basin. Convergence at this stage probably occurred in a transpressional setting, as suggested by sinistral lateral movements along major tectonic boundaries of SW Iberia (Crespo-Blanc, 1992). In this scenario, convergence may have also facilitated reactivation of previous thrusts, particularly in the upper crust (e.g., out-of-sequence thrusts with Serpukhovian-Bashkirian age). Later pronounced extension within the orogenic hinterland facilitated the widening of former sedimentary basins over the Variscan allochthonous nappes (e.g., Los Pedroches basin).

The opening of the Beja-Acebuches basin also covers the start of the orogenic collapse (see section 4.5). Either a mantle upwelling in response to the inception of the Beja-Acebuches basin, and/or the subsequent consumption of that same basin by NE-directed subduction, are two expected contributors of deep-sourced material to bear on the orogenic collapse. The ca. 328-317 Ma calc-alkaline to adakitic-like magmatism (Pavia pluton; Lima et al., 2013) lying to the north of the Beja-Acebuches Ophiolite may represent the product of such later subduction. In this way, we find older (and much more abundant) evidence of such crustal growth toward SW Iberia (closer to the Beja-Acebuches Ophiolite; e.g., Pereira et al., 2009; Cambeses et al., 2015) than to Central and NW Iberia (toward the advancing front of Variscan allochthonous nappes; e.g., Dias et al., 2002; Rodríguez et al., 2007). This makes a petro-geochronological trend that may express either the lag in the rise of the mantle after maximum crustal thickening (earlier in SW Iberia by favorable mantle topography after rifting), and/or the arrival of mantle-derived melts related to a downgoing oceanic tract that sinks progressively to the northeast (consumption of the Beja-Acebuches basin), i.e. toward the advancing front of Variscan allochthonous nappes. That sector of the orogen would be equivalent to a broad back-arc region relative to the subduction zone closing the Beja-Acebuches basin. Eventual migrations of its subduction hinge (e.g., Doglioni et al., 2007) might explain transient extensional or compressional regimes affecting that section of the orogen.

The east- and northeast-directed collapse of the eastern orogenic hinterland is roughly contemporaneous with the early stages of east-directed thrusting in the western part of the foreland of the Cantabrian Zone (Dallmeyer et al., 1997; Martínez Catalán et al., 2003). Such coupling between orogen-perpendicular hinterland extension and foreland compression in the Gondwanan flank of the orogen has its equivalent in the Laurussian side (South Portuguese Zone). South-directed extension along the Puente Génave-Castelo de Vide Detachment is coeval with the southerly propagation of thrusts and folds in the foreland of the South Portuguese Zone. Therefore the lateral spreading of the orogenic crust has been a fundamental cause for triggering Variscan shortening across foreland basins at both sides of the orogen (Cantabrian and South Portuguese zones).

4.7. Late strike-slip tectonics (Fig. 3)

Convergence persisted during the orogenic collapse, which, in turn, waned as extensional flow reduced gradients of potential energy. Thermal equilibrium was not fully achieved in the process, since syn-orogenic magmatism remained throughout the Bashkirian and Moscovian (Pereira et al., 2009, 2015; Martínez Catalán et al., 2014; Cambeses et al., 2015). Subhorizontal extension was eventually outpaced by superimposed subhorizontal compression in Moscovian times. Some of the magmatism at this stage occurred in close relation to strike-slip shear zones (e.g., Aranguren et al., 1997; Valle Aguado et al., 2005; Carracedo et al., 2009), which accommodated most of the lateral components of convergence along their central parts and distributed shortening in their tectonic blocks, thus producing open upright folds all over the Variscan hinterland. Transcurrent deformation reworked previous thrusts and normal faults and partly redrew the map of tectonic blocks. Dextral and sinistral strike-slip shear zones acted together and created escape tectonics settings at local scale (Iglesias Ponce de Leon and Choukroune, 1980). Such settings in the hinterland were coeval with further propagation of thrusts and folds to both the Gondwanan and Laurussian forelands. Oroclinal bending of the orogen occurred in the course of all of this deformation, starting from the Serpukhovian-Bashkirian extensional collapse and culminating with the folding of the latest syn-orogenic deposits of the Gondwanan foreland.

5. Conclusions

Strike-slip deformation during the Moscovian segmented the hinterland of the Variscan orogen into new tectonic blocks, partly different from those operating during previous convergence processes between Gondwana and Laurussia. Becoming fully aware of this particular switch in the architecture of the orogen (even if it was transitional) is essential for understanding the common structural history linked to previous tangential tectonics at both sides of major transcurrent shear zones, such as the Coimbra-Córdoba shear zone. Shear zones accommodating large amounts of tangential deformation, transported pieces of continental and oceanic crust located at the periphery of Gondwana that were affected by previous Variscan deformation. These shear zones are envisaged as the rulers during the early stages of Pangea amalgamation, which, however, did not seal Gondwana and Laurussia once for all. Inception of short-lived oceanic basins following periods of convergence provides solid evidence on a complex amalgamation process in southern Europe, hardly explainable by a single collisional process.

Based on our integration of structural and geochronological data, the Variscan tectonic evolution of the Iberian Massif can be summarized as follows (Paleozoic geographic coordinates):

1. Following the closure of the Rheic Ocean, Gondwana and Laurussia collided in the Lower Devonian. Kinematics of major structures that developed toward the most external margin of Gondwana support that Laurussia was the lower plate to the Rheic suture.
2. Contraction over the margin of Gondwana initiated an intra-Gondwana, continental subduction zone dipping to the north, which progressively spread under the Rheic suture.
3. A transient period of extension after continental subduction led to the opening of an intra-Gondwana oceanic basin in the Lower-Middle Devonian. Such intra-orogenic rifting coupled with the initial exhumation of high-P rocks within the former continental subduction system.
4. Closure of the intra-Gondwana basin in the Upper Devonian caused the accretion to the north of Devonian oceanic crust, then Cambrian-Ordovician transitional crust, and finally the subduction of inner sections of Gondwana to the north.

5. Continuous convergence between Gondwana and Laurussia during the early Carboniferous was accommodated by underthrusting of Gondwanan lithosphere to the north, below the peri-Gondwanan domain that had been previously involved in the collisional orogenesis. Protracted underthrusting locked the Upper Devonian intra-Gondwana subduction first, and then forced the mechanical coupling between the lower and upper tectonic plate. Coeval shearing throughout the orogenic crust generated a series of south-directed folds and a series of extensional faults in the upper plate. Ductile drag exerted by the lower plate extended the Upper Devonian subduction system and the intra-Gondwana suture zone under the upper plate, thus shaping this whole ensemble of peri-Gondwanan terranes into a set of allochthonous units.
6. Further convergence nucleated a system of out-of-sequence thrusts, which reworked the intra-Gondwana suture in the course of its obduction onto the Gondwana mainland.
7. Rifting of the resulting overthickened crust led to the opening of a short-lived oceanic basin (Beja-Acebuches Ophiolite) near the Gondwana-Laurussia suture zone formed in the Lower Devonian.
8. Intra-continental extension was followed by or coeval with the gravitational collapse and thermal re-equilibration of the orogen, which remained active up to the late Carboniferous. Continental convergence resumed shortly afterwards, and forced the closure of newly-formed oceanic basins. Deformation propagated via thrusts and folds toward the mainland of both Gondwana and Laurussia and favored the reactivation of former thrusts.
9. Lithosphere extension in the hinterland was progressively replaced by strike-slip deformation. Oroclinal bending of the orogen started in this transition. Lateral

tectonics at this stage was manifested in discrete, subvertical shear bands and in the upright folding of previous flat-lying structures. Foreland propagation of deformation continued during this period. Variscan deformation concluded with the development of late oroclinal bends affecting the whole structural grain of the orogen.

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Figure Captions

Fig. 1. Zonation of the Variscan orogen after Martínez Catalán et al. (2007) and Díez Fernández and Arenas (2015). The locations of the Coimbra-Córdoba Shear Zone and the oroclinal bends of the orogen are shown.

Fig. 2. Geological map showing the main zones of the Iberian Massif (after Díez Fernández and Arenas, 2015). Abbreviations: AF — Azuaga Fault; BToIP — Basal Thrust of the Iberian Parautochthon; BAO — Beja–Acebuches Ophiolite; CA — Carvalhal Amphibolites; CF — Canaleja Fault; CMU — Cubito–Moura Unit; CO — Calzadilla Ophiolite; CU — Central Unit; EsT — Espiel Thrust; ET—Espina Thrust; HF— Hornachos Fault; IOMZO —Internal Ossa-Morena Zone Ophiolites; J–PCSZ — Juzbado-Penalva do Castelo Shear Zone; LFT — Lalín-Forcarei Thrust; LPSZ — Los Pedroches Shear Zone; LLSZ — Llanos Shear Zone; MLSZ — Malpica–Lamego Shear Zone; MF — Matachel Fault; OF — Onza Fault; OVD — Obejo–Valsequillo Domain; PG–CVD — Puente Génave–Castelo de Vide Detachment; PRSZ— Palas de Rei Shear Zone; PTSZ — Porto–Tomar Shear Zone; RF — Riás Fault; SISZ —South Iberian Shear Zone; VF — Viveiro Fault; ZSI — Zalamea de la Serena Imbricates.

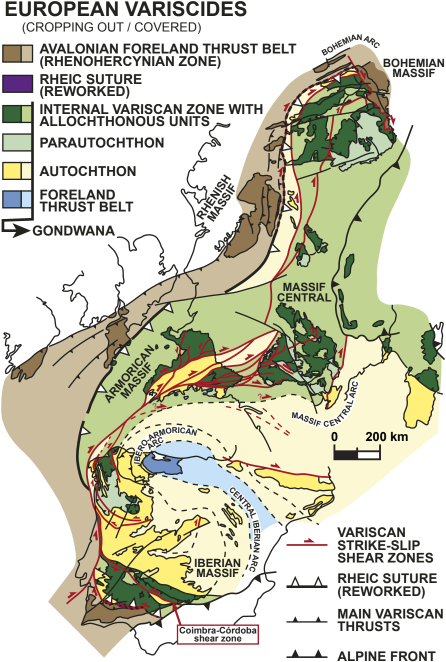
Fig. 3. Composite cross-section of major tectonic elements of the Iberian Massif (after Díez Fernández and Arenas, 2015). Abbreviations follow Figure 2. The location of the Iberian Reflective Body is shown.

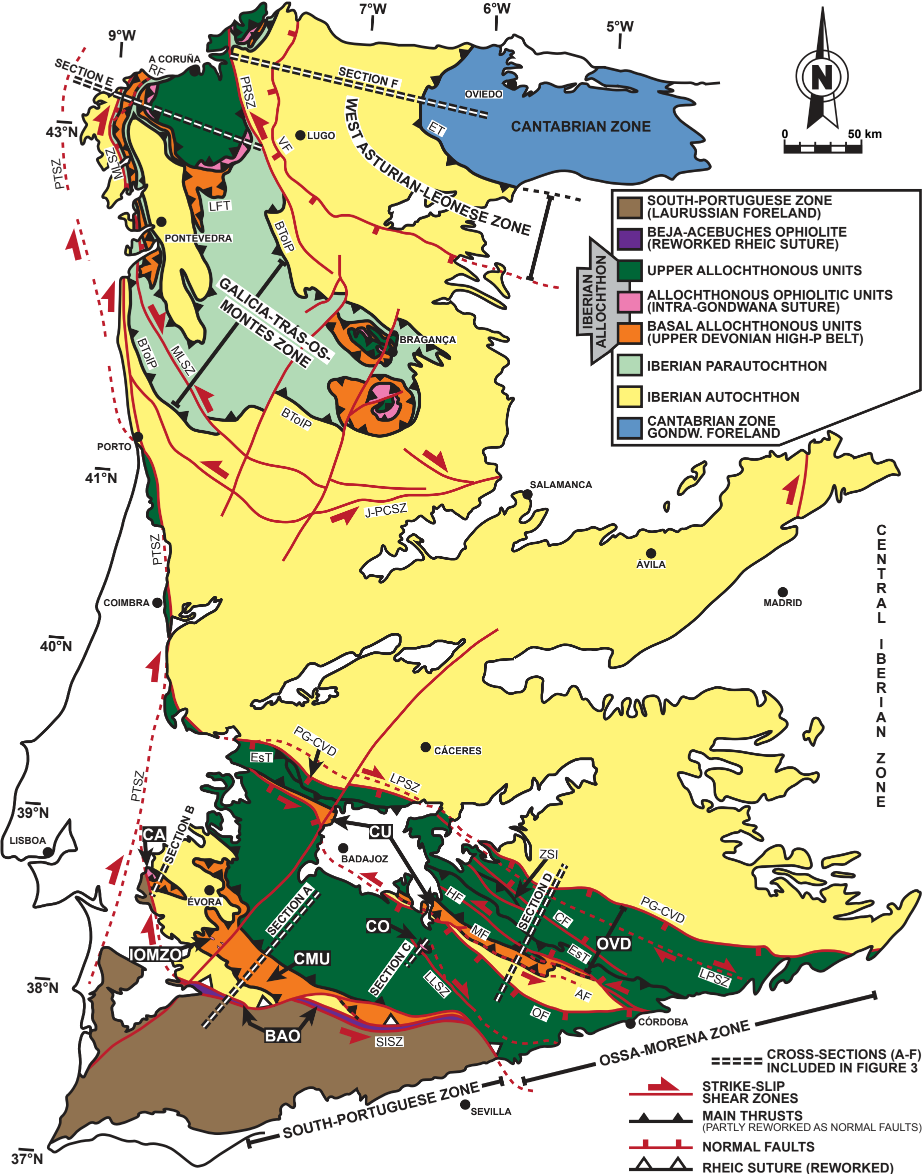
Fig. 4. Idealized Variscan evolution of the Iberian Massif during the Devonian (see text for explanation). Circled numbers refer to specific structures and basins of the Iberian Massif. (a) Simplified pre-collisional paleogeography across the margin of Gondwana after restoration of Variscan deformation. Note the model is at 50% scale relative to the rest of the drawings (b) Tectonic setting showing the onset of Variscan deformation in the peri-Gondwana realm after the closure of the Rheic Ocean. Note the north-directed tectonic escape proposed for the peri-Gondwanan domain. (c) Opening of an intra-Gondwanan oceanic basin in the Lower-Middle Devonian. (d) Closure of the intra-Gondwana basin by accretion of different tectonic slices of oceanic crust (ophiolitic units) and followed by subduction of Gondwanan continental crust under the previously stacked ophiolites.

Fig. 5. Idealized Variscan evolution of the Iberian Massif during the Carboniferous (see text for explanation). Circled numbers refer to specific structures and basins of the Iberian Massif (numbering continues list of Figure 4). (a) Thrust and fold nappe tectonics during the early stages of the emplacement of the Iberian Allochthon. Propagation of Gondwana-verging folds in the Allochthon accompanied the underthrusting of the Iberian Autochthon and Parautochthon. (b) Climax of Gondwanan lithosphere underthrusting and onset of out-of-sequence thrusts. Note the proposed mantle topography near the Rheic suture that separates Gondwana from Laurussia. Time lines in sections b and c overlap each other because they both represent processes that may have occurred simultaneously. (c) Inception of the Beja-Acebuches basin, beginning of the orogenic extensional collapse and advance of out-of-sequence thrusts. (d) Closure of the Beja-Acebuches basin, widespread collapse of the orogen and

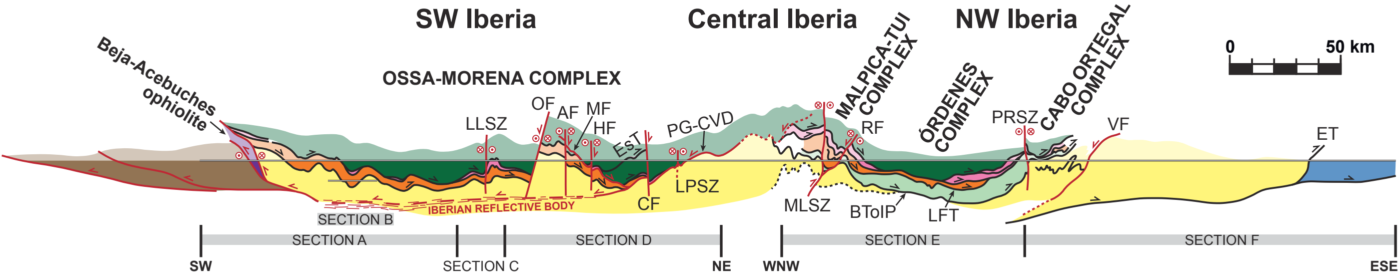
reactivation of out-of-sequence thrusts. Along-strike movements are not represented for extensional faults. Location and kinematics of later strike-slip shear zones is shown. Abbreviations: IRB — Iberian Reflective Body; SZ — Shear Zone.

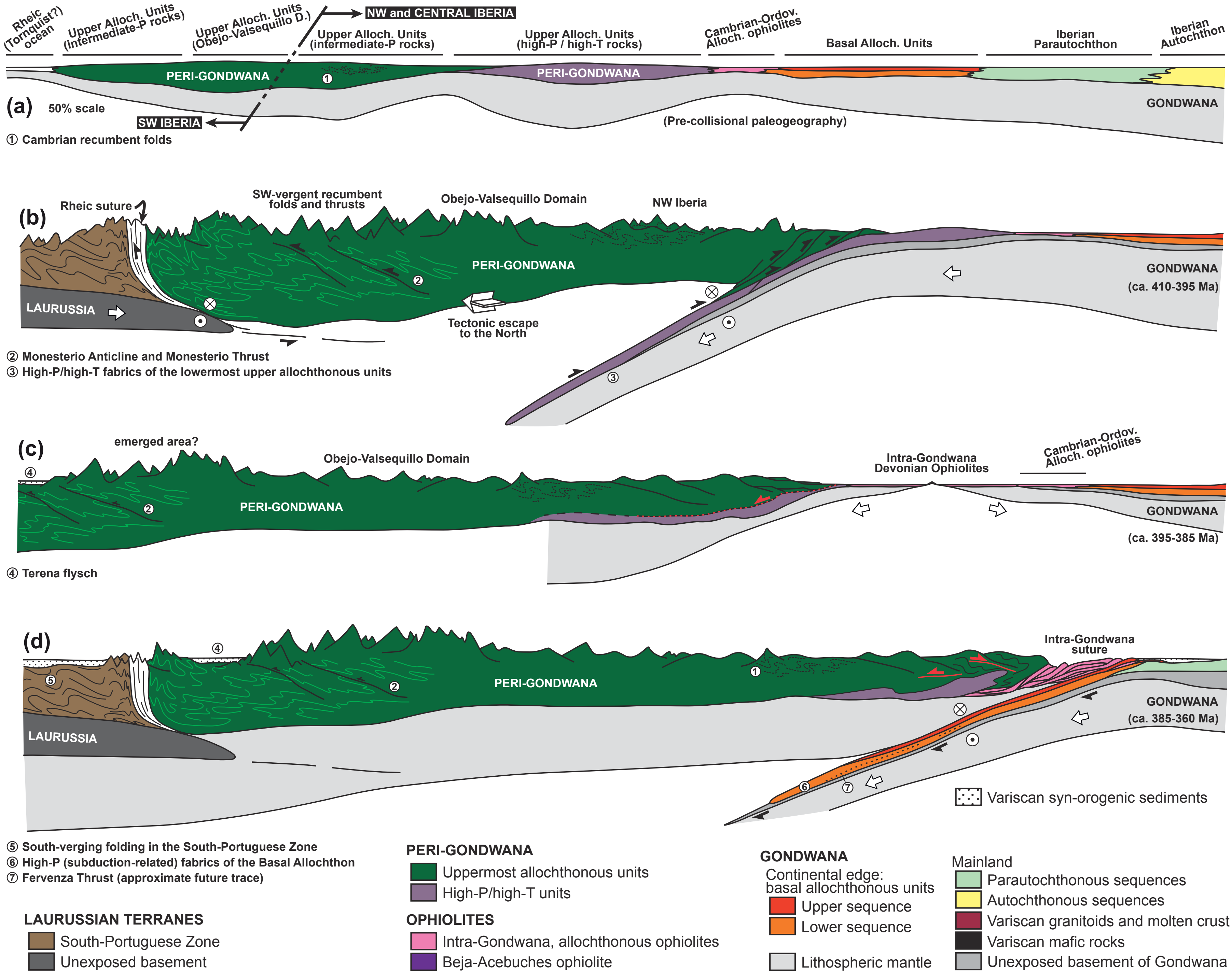
Table 1. Summary of the main tectonic events recognized on each geotectonic zone of the Iberian Massif during the Variscan orogeny. Dashed lines show a time-based correlation. Names in capital letters refer to the nomenclature utilized in this work for tectonic integration, while the commonly used regional names are shown in grey boxes below (the terms autochthon and allochthon inside parentheses refer to the allochthonous or autochthonous nature after Díez Fernández and Arenas, 2015).


EUROPEAN VARISCIDES
(CROPPING OUT / COVERED)
AVALONIAN FORELAND THRUST BELT (RHENOHERCYNIAN ZONE)
RHEIC SUTURE (REWORKED)
INTERNAL VARISCAN ZONE WITH ALLOCHTHONOUS UNITS
PARAUTOCHTHON
AUTOCHTHON
FORELAND THRUST BELT
GONDWANA
BOHEMIAN ARC
BOHEMIAN MASSIF
RHENISH MASSIF
MASSIF CENTRAL
ARMORICAN MASSIF
MASSIF CENTRAL ARC
IBERO-ARMORICAN ARC
CENTRAL IBERIAN ARC
IBERIAN MASSIF
0
200 km
VARISCAN STRIKE-SLIP SHEAR ZONES
RHEIC SUTURE (REWORKED)
MAIN VARISCAN THRUSTS
ALPINE FRONT
Coimbra-Córdoba shear zone

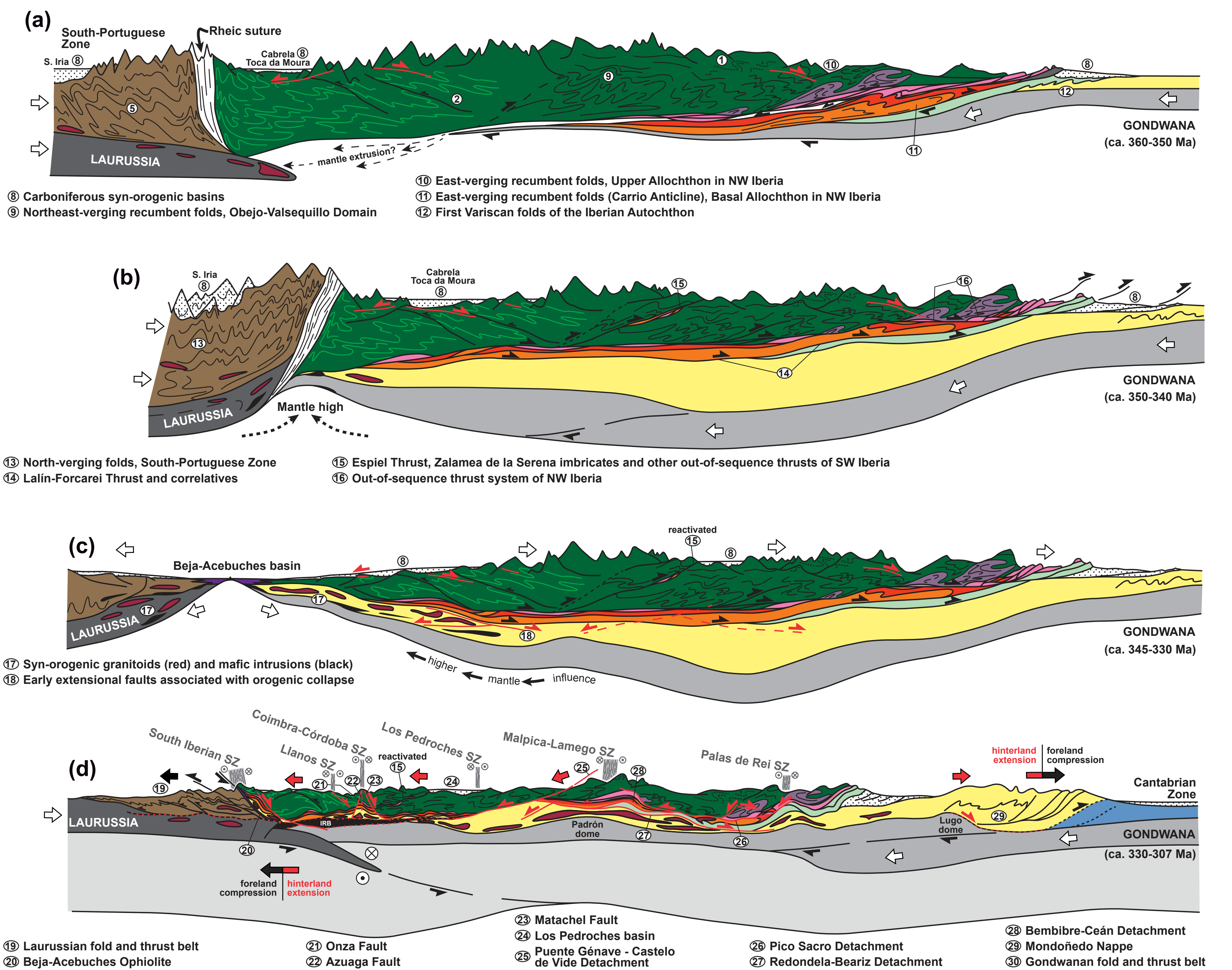


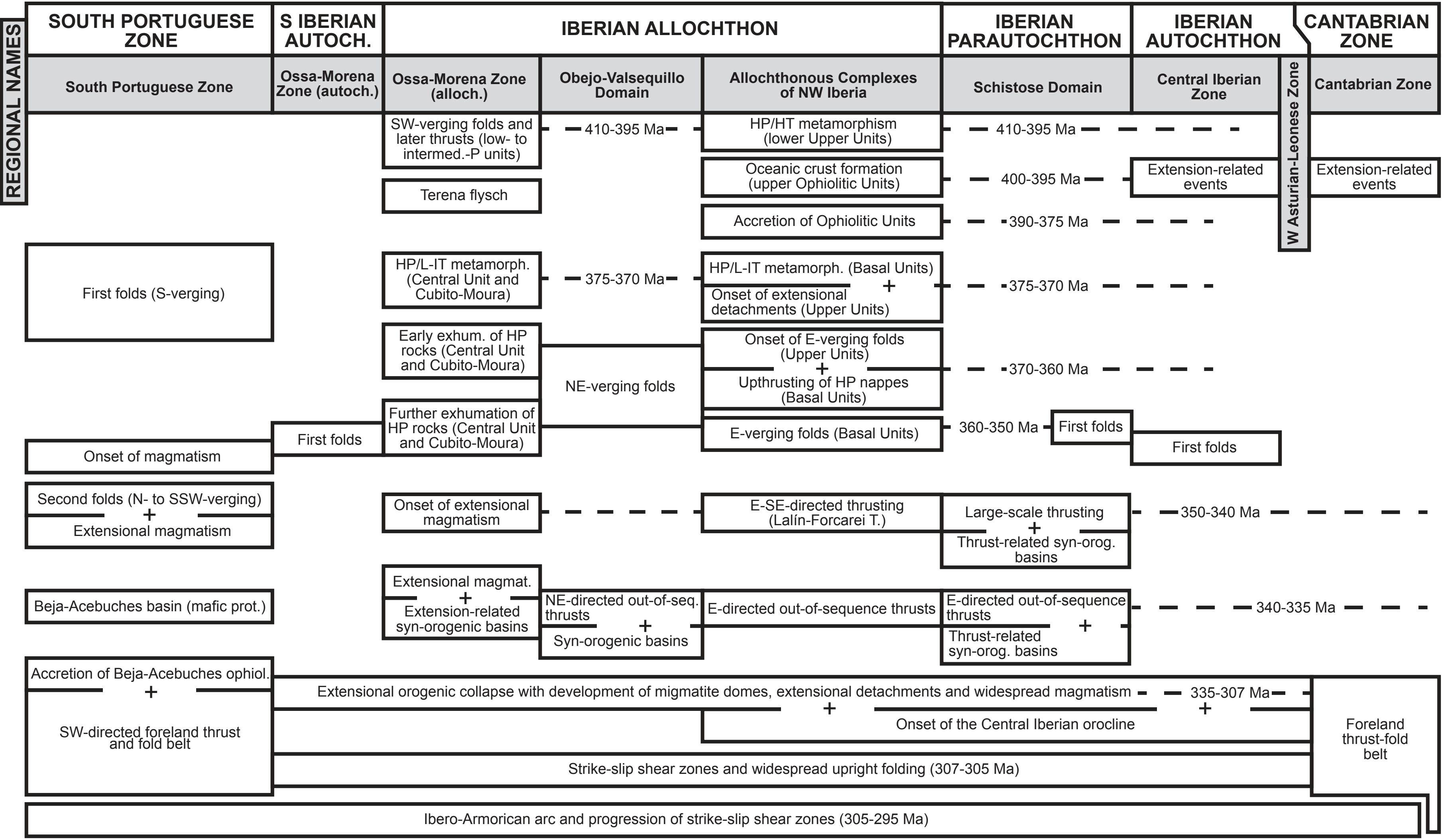
9°W
7°W
6°W
5°W
N
0
50 km
43°N
41°N
40°N
39°N
38°N
37°N
A CORUÑA
RF
SECTION E
SECTION F
OVIEDO
CANTABRIAN ZONE
ET
PRSZ
VF
LUGO
WEST ASTURIAN-LEONESE ZONE
MLSZ
PTSZ
LFT
PONTEVEDRA
BToIP
GALICIA-TRÁS-OS-MONTES ZONE
BRAGANÇA
PORTO
SALAMANCA
J-PCSZ
ÁVILA
MADRID
COIMBRA
CENTRAL IBERIAN ZONE
CÁCERES
PG-CVD
EsT
LPSZ
LISBOA
CA
SECTION B
CU
BADAJOZ
ZSI
ÉVORA
SECTION A
HF
SECTION D
CF
CO
MF
OVD
IOMZO
CMU
SECTION C
LLSZ
AF
OF
BAO
SISZ
CÓRDOBA
OSSA-MORENA ZONE
SOUTH-PORTUGUESE ZONE
SEVILLA
SOUTH-PORTUGUESE ZONE (LAURUSSIAN FORELAND)
BEJA-ACEBUCHES OPHIOLITE (REWORKED RHEIC SUTURE)
IBERIAN ALLOCHTHON
UPPER ALLOCHTHONOUS UNITS
ALLOCHTHONOUS OPHIOLITIC UNITS (INTRA-GONDWANA SUTURE)
BASAL ALLOCHTHONOUS UNITS (UPPER DEVONIAN HIGH-P BELT)
IBERIAN PARAUTOCHTHON
IBERIAN AUTOCHTHON
CANTABRIAN ZONE GONDW. FORELAND
CROSS-SECTIONS (A-F) INCLUDED IN FIGURE 3
STRIKE-SLIP SHEAR ZONES
MAIN THRUSTS (PARTLY REWORKED AS NORMAL FAULTS)
NORMAL FAULTS
RHEIC SUTURE (REWORKED)


SW Iberia
Central Iberia
NW Iberia
0
50 km
Beja-Acebuches ophiolite
OSSA-MORENA COMPLEX
MALPICA-TUI COMPLEX
ÓRDENES COMPLEX
CABO ORTEGAL COMPLEX
LLSZ
OF
AF
MF
HF
EsT
PG-CVD
RF
PRSZ
VF
ET
CF
LPSZ
MLSZ
BToIP
LFT
IBERIAN REFLECTIVE BODY
SECTION B
SECTION A
SECTION C
SECTION D
SECTION E
SECTION F
SW
NE
WNW
ESE



Rheic (Tornquist?) ocean
Upper Alloch. Units (intermediate-P rocks)
Upper Alloch. Units (Obejo-Valsequillo D.)
NW and CENTRAL IBERIA
Upper Alloch. Units (intermediate-P rocks)
Upper Alloch. Units (high-P / high-T rocks)
Cambrian-Ordov. Alloch. ophiolites
Basal Alloch. Units
Iberian Parautochthon
Iberian Autochthon
PERI-GONDWANA
PERI-GONDWANA
(a)
50% scale
SW IBERIA
GONDWANA
(Pre-collisional paleogeography)
① Cambrian recumbent folds
(b)
Rheic suture
SW-vergent recumbent folds and thrusts
Obejo-Valsequillo Domain
NW Iberia
PERI-GONDWANA
GONDWANA
(ca. 410-395 Ma)
LAURUSSIA
Tectonic escape to the North
② Monesterio Anticline and Monesterio Thrust
③ High-P/high-T fabrics of the lowermost upper allochthonous units
(c)
emerged area?
Obejo-Valsequillo Domain
Intra-Gondwana Devonian Ophiolites
Cambrian-Ordov. Alloch. ophiolites
PERI-GONDWANA
GONDWANA
(ca. 395-385 Ma)
④ Terena flysch
(d)
Intra-Gondwana suture
PERI-GONDWANA
GONDWANA
(ca. 385-360 Ma)
LAURUSSIA
Variscan syn-orogenic sediments
⑤ South-verging folding in the South-Portuguese Zone
⑥ High-P (subduction-related) fabrics of the Basal Allochthon
⑦ Fervenza Thrust (approximate future trace)
LAURUSSIAN TERRANES
South-Portuguese Zone
Unexposed basement
PERI-GONDWANA
Uppermost allochthonous units
High-P/high-T units
OPHIOLITES
Intra-Gondwana, allochthonous ophiolites
Beja-Acebuches ophiolite
GONDWANA
Continental edge: basal allochthonous units
Upper sequence
Lower sequence
Lithospheric mantle
Mainland
Parautochthonous sequences
Autochthonous sequences
Variscan granitoids and molten crust
Variscan mafic rocks
Unexposed basement of Gondwana


(a)
South-Portuguese Zone
Rheic suture
S. Iria ⑧
Cabrela ⑧
Toca da Moura
LAURUSSIA
mantle extrusion?
GONDWANA
(ca. 360-350 Ma)
⑧ Carboniferous syn-orogenic basins
⑨ Northeast-verging recumbent folds, Obejo-Valsequillo Domain
⑩ East-verging recumbent folds, Upper Allochthon in NW Iberia
⑪ East-verging recumbent folds (Carrio Anticline), Basal Allochthon in NW Iberia
⑫ First Variscan folds of the Iberian Autochthon
(b)
S. Iria ⑧
Cabrela
Toca da Moura ⑧
LAURUSSIA
Mantle high
GONDWANA
(ca. 350-340 Ma)
⑬ North-verging folds, South-Portuguese Zone
⑭ Lalín-Forcarei Thrust and correlatives
⑮ Espiel Thrust, Zalamea de la Serena imbricates and other out-of-sequence thrusts of SW Iberia
⑯ Out-of-sequence thrust system of NW Iberia
(c)
Beja-Acebuches basin
reactivated ⑮
LAURUSSIA
GONDWANA
(ca. 345-330 Ma)
higher mantle influence
⑰ Syn-orogenic granitoids (red) and mafic intrusions (black)
⑱ Early extensional faults associated with orogenic collapse
(d)
South Iberian SZ
Coimbra-Córdoba SZ
Llanos SZ
reactivated ⑮
Los Pedroches SZ
Malpica-Lamego SZ
Palas de Rei SZ
hinterland extension
foreland compression
Cantabrian Zone
LAURUSSIA
IRB
Padrón dome
Lugo dome
GONDWANA
(ca. 330-307 Ma)
foreland compression
hinterland extension
⑲ Laurussian fold and thrust belt
⑳ Beja-Acebuches Ophiolite
㉑ Onza Fault
㉒ Azuaga Fault
㉓ Matachel Fault
㉔ Los Pedroches basin
㉕ Puente Génave - Castelo de Vide Detachment
㉖ Pico Sacro Detachment
㉗ Redondela-Beariz Detachment
㉘ Bembibre-Ceán Detachment
㉙ Mondoñedo Nappe
㉚ Gondwanan fold and thrust belt


REGIONAL NAMES
SOUTH PORTUGUESE ZONE
S IBERIAN AUTOCH.
IBERIAN ALLOCHTHON
IBERIAN PARAUTOCHTHON
IBERIAN AUTOCHTHON
CANTABRIAN ZONE
South Portuguese Zone
Ossa-Morena Zone (autoch.)
Ossa-Morena Zone (alloch.)
Obejo-Valsequillo Domain
Allochthonous Complexes of NW Iberia
Schistose Domain
Central Iberian Zone
W Asturian-Leonese Zone
Cantabrian Zone
SW-verging folds and later thrusts (low- to intermed.-P units)
410-395 Ma
HP/HT metamorphism (lower Upper Units)
410-395 Ma
Terena flysch
Oceanic crust formation (upper Ophiolitic Units)
400-395 Ma
Extension-related events
Extension-related events
Accretion of Ophiolitic Units
390-375 Ma
First folds (S-verging)
HP/L-IT metamorph. (Central Unit and Cubito-Moura)
375-370 Ma
HP/L-IT metamorph. (Basal Units)
+
Onset of extensional detachments (Upper Units)
375-370 Ma
Early exhum. of HP rocks (Central Unit and Cubito-Moura)
NE-verging folds
Onset of E-verging folds (Upper Units)
+
Upthrusting of HP nappes (Basal Units)
370-360 Ma
Further exhumation of HP rocks (Central Unit and Cubito-Moura)
First folds
Onset of magmatism
E-verging folds (Basal Units)
360-350 Ma
First folds
First folds
Second folds (N- to SSW-verging)
+
Extensional magmatism
Onset of extensional magmatism
E-SE-directed thrusting (Lalín-Forcarei T.)
Large-scale thrusting
+
Thrust-related syn-orog. basins
350-340 Ma
Beja-Acebuches basin (mafic prot.)
Extensional magmat.
+
Extension-related syn-orogenic basins
NE-directed out-of-seq. thrusts
+
Syn-orogenic basins
E-directed out-of-sequence thrusts
E-directed out-of-sequence thrusts
+
Thrust-related syn-orog. basins
340-335 Ma
Accretion of Beja-Acebuches ophiol.
+
SW-directed foreland thrust and fold belt
Extensional orogenic collapse with development of migmatite domes, extensional detachments and widespread magmatism
335-307 Ma
+
+
Onset of the Central Iberian orocline
Foreland thrust-fold belt
Strike-slip shear zones and widespread upright folding (307-305 Ma)
Ibero-Armorican arc and progression of strike-slip shear zones (305-295 Ma)