IX.2

Reviewing the Arcuate Structures in the Iberian Variscides; Constraints and Genetical Models

Index

IX.2.1. Introduction

First order arcuate structures are a common feature in orogenic belts (Argand, 1924; Carey, 1955). Their understanding is an important issue in Plate Tectonics, because similar shapes could result from different processes. Primary arcs are induced by the formation of the fold belt, as happens when moulding the orogen around a promontory. When the curved shape is not a pre-orogenic feature but the result of an impressed strain on a previous linear belt, it is known as an orocline (Carey, 1955). Curved arcuations could also be considered as thick skinned or thin skinned: in the formers ones the strain pattern was developed both in the cover and the basement of the orogen, while the strain pattern in thin skinned arcs is restricted to the cover.

The understanding of orogenic arcs is usually easier in active orogens because the continuity of major structures between both branches is often visible. In such young tectonic environments, major arcuations are common in convergent settings, either related to ocean-ocean (*e.g.* Scotia Arc, Dalziel, 1971; De Wit, 1977), ocean-continent (*e.g.* the Central Andean orocline, Eichelberger and McQuarrie, 2014 or the Banda Arc, Vroon *et al.*, 1995; Harris, 2011) or continent-continent (*e.g.* western and eastern syntaxis of Himalayas; Tapponier and Molnar, 1976; Matte, 1986) collisions.

In old orogens major arcs are more difficult to emphasize because the original continuity is often:

- disrupted by the superposition of younger structures or magmatic batholiths;
- hidden below younger sediments;
- dismembered by the opening of new oceans.

Nevertheless, since the early works several major arcs have been described in the Variscan Belt, not only at the orogen scale but also in Iberia (*e.g.* Du Toit, 1937; Carey, 1955). However, at this moment there is still a lack of understanding concerning the formation of the first order Variscan arcs in Iberia, which is the main purpose of this work. A four step approach will be used:

- A historical review of the major arcuate structures;
- A critical review of the existing data (*e.g.* structural, deformation age, lithostratigraphic, and paleomagnetic);
- A discussion of the previous models;
- A unifying approach trying to conciliate the previous data.

The discussion of the Iberian Variscan arcs is crucial, mostly because since 2010 several papers strongly emphasize the so-called Central-Iberian Arc (*e.g.* Martínez Catalán, 2011a; 2011b; Johnston *et al.*, 2013). In spite of the weakness of the data supporting this Arc, several models were proposed for its formation, considering an Upper Carboniferous-Permian age for all the Iberian Arcs (*e.g.* Martínez Catalán, 2011c; Johnston *et al.*, 2013; Weil *et al.*, 2013; Martínez Catalán *et al.*, 2014), which is difficult to conciliate with most of the data.

This paper is also a contribution to the subject concerning the primary or secondary origin of curved orogenic belts, emphasizing the complexity of these structures and the care which should be taken when using simple models.

IX.2.2. Arcuate Variscan Patterns in Iberia; a Historical Approach

Any discussion concerning the geodynamical evolution related to the major Variscan arcs in Iberia should address the major zoning of this sector of the fold belt.

IX.2.2.1. Zones in Iberian Variscides

Since Lotze (1945) the Iberian Variscides has been divided in several zones based on stratigraphic, structural, metamorphic and magmatic features. Later works have led to some minor modifications in their number and boundaries (Fig. 1; Julivert *et al.*, 1972; Julivert and Martínez, 1983; Farias *et al.*, 1987; Arenas *et al.*, 1988; Martínez Catalán, 1990). This zoning reflects the complex evolution due to the superposition mainly of the Neoproterozoic Cadomian collision, the Lower Palaeozoic extensional tectonic activity and the Upper Palaeozoic Variscan Orogen.

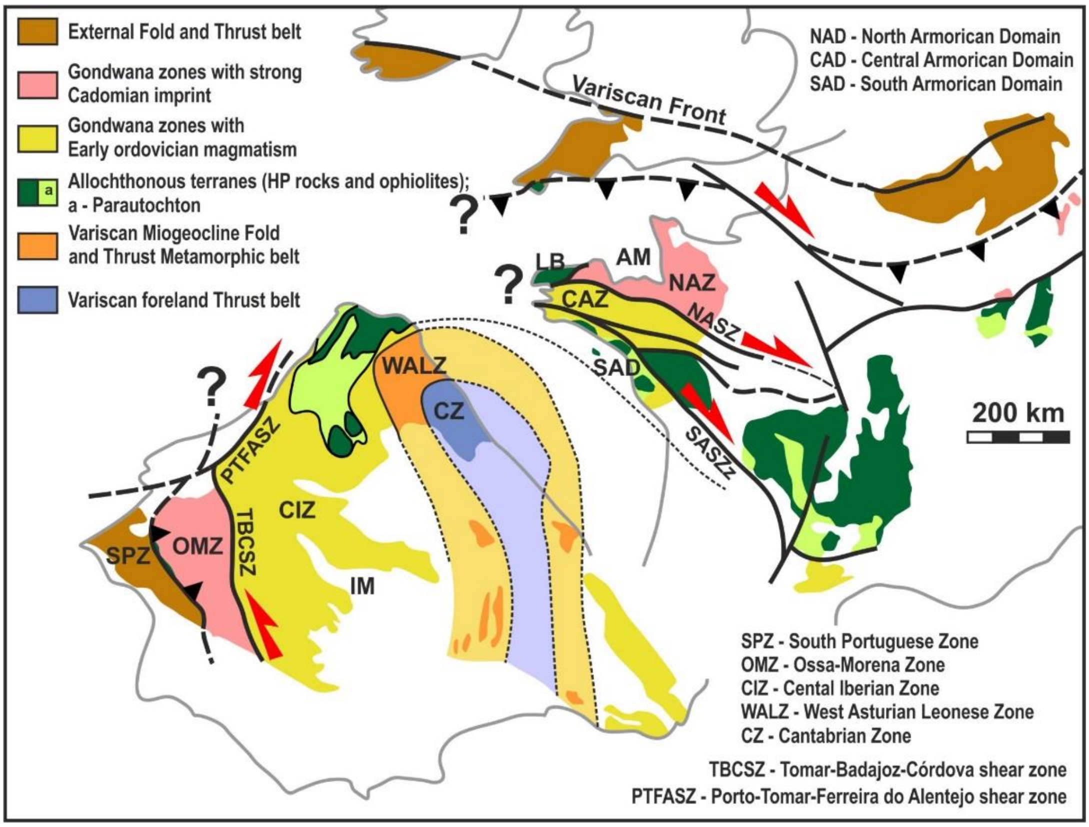

Figure 1 – Main Variscan tectonostratigraphic units in Iberia in the framework of the European Variscides (adapted from Lotze, 1945; Julivert *et al.*, 1972; Julivert and Martínez, 1983; Farias *et al.*, 1987; Arenas *et al.*, 1988; Martínez Catalán, 1990; 2011; Ribeiro *et al.*, 2007; Ballèvre *et al.*, 2014).

The Cantabrian Zone (CZ), considered the thin skinned foreland fold and thrust belt in the NW part of the Iberian Variscides (Pérez-Estaún, 1990), is characterized by a decollement level within the Palaeozoic cover above a non-exposed Precambrian basement overlaid by its fixed cover. The Lower Palaeozoic is composed of a passive margin sequence of shallow-marine to shoreface facies (Julivert and Marcos, 1973), which becomes progressively thinner towards the East (Pérez-Estaún *et al.*, 1991). This Cambro-Devonian sequence (Pérez-Estaún, 1990) was deformed in the Pennsylvanian, giving rise to a thin-skinned fold-and-thrust belt verging towards Gondwana craton (Pérez-Estaún *et al.*, 1988). The progressive emplacement of this imbricate complex structure, which is rooted below the West Asturian Leonese Zone (Pérez-Estaún *et al.*, 1991), controls the deposition of syn-kinematic marine to terrestrial foreland basin successions (Marcos and Pulgar, 1982; Pérez-Estaún *et al.*, 1988; 1991). The metamorphism related to this Variscan shortening is almost absent, although it could locally attain a low grade (Pérez-Estaún, 1990).

The boundary between the CZ and the West Asturian Leonese Zone, is considered the Neoproterozoic rocks outcropping in the complex Narcea Antiform (Julivert *et al.*, 1972).

The West Asturian Leonese Zone (WALZ) is often considered the transition between the foreland CZ and the more internal zones of the Variscan hinterland core (Pérez-Estaún *et al.*, 1991). It consists of a thick Upper Proterozoic flyschoid series unconformably overlain by a thick cover (Marcos *et al.*, 2004) of shallow-water Lower Cambrian to Lower Devonian deposits, where thick siliciclastic units are dominant (Pérez-Estaún *et al.*, 1990; 1991; Fernández-Suárez *et al.*, 2000). Such unconformity, has also been found in the Cantabrian Zone (Lotze, 1956; De Sitter, 1961), being a major characteristic of these two domains. The previous WALZ sequence presents a pervasive Variscan deformation due to the interference of at least three major tectonic events, that have originated a general structure (sometimes with large recumbent folds, like the Mondoñedo anticline and the Courel syncline) facing the external part of the orogenic belt (Bastida *et al.*, 1986; Martínez Catalán *el al.*, 1990; Pérez-Estaún *et al.*, 1991; Fernández *et al.*, 2007; Bastida *et al.*, 2010). The Variscan metamorphic grade increases towards the West, from greenschist to amphibolite facies (Suárez *et al.*, 1990). The granitoid plutonism exhibits a similar trend, becoming abundant in the western sectors of WALZ (Corretgé *et al.*, 1990; Pérez-Estaún *et al.*, 1991).

The western limit of the WALZ is still controversial. While some authors (*e.g.* Ábalos *et al.*, 2002) follow the initial proposal of Ollo de Sapo anticlinorium (Julivert *et al.*, 1972), others (*e.g.* Marcos, 2004) consider the more complex boundary proposed by Martínez Catalán (1985) formed by the main Vivero normal fault and its southern continuation in the Peñalba and Courel synclines (Martínez Catalán *et al.*, 1992; Fernández *et al.*, 2007).

The rather heterogeneous Central Iberian Zone (CIZ) is the axial domain of the Iberian Variscan Fold Belt with abundant granitic plutonism and metamorphism ranging from very low-grade to high-grade. Three main stratigraphic features have been fundamental for its individualization (Julivert *et al.*, 1972), because its Palaeozoic succession from Arenig to Middle Devonian is similar to the one found in adjacent zones (Ribeiro, 1990):

- The predominance of Pre-Ordovician sequences;
- The transgressive character of the Lower Ordovician quartzites;
- The presence of a pervasive unconformity which places the Lower Ordovician over Precambrian to Cambrian rocks (Douro-Beiras Super Group).

The presence of high-grade metamorphic allochthonous complexes and the abundant Silurian volcanism, led some authors to individualize in the NW of CIZ a Galicia - Trás-os-Montes domain, which has been considered either as a sub-Zone (Julivert *et al.*, 1972) or a Zone (Tex and Floor, 1971; Farias *et al.*, 1987). Whatever the option, its allochthonous behaviour over the CIZ autochthon is recognized (*e.g.* recent compilations of the Iberian Geology of Gibbons and Moreno, 2002; Vera, 2004; Dias *et al.*, 2013a). This Galicia - Trás-os-Montes Zone comprises two domains, bounded by a major thrust, with distinct paleogeographic and tectonometamorphic histories (*e.g.* Martínez Catalán *et al.*, 2004; Dias and Ribeiro, 2013):

- The parautochthonous "schistose" domain with clear lithostratigraphic affinities with the CIZ autochthon, mainly with its upper part;
- The domain of the allochthonous complexes with high-grade massifs.

Concerning the southern boundary of CIZ, Lotze (1945) proposed the elongated Los Pedroches granitic batholith. The Tectonic Map of the Iberia (Julivert *et al.*, 1972) still uses this batholith for the Spanish limit, while in Portugal it is marked by the Ferreira do Zêzere and Portalegre thrusts. However, some authors (*e.g.* Díez Balda *et al.*, 1990; Azor *et al.*, 1994; Martínez Catalán *et al.*, 2004) considered that the main stratigraphic and structural changes are marked by the main Tomar - Badajoz - Cordoba intra-continental shear zone (TBCSZ).

The Ossa-Morena Zone (OMZ), considered the southernmost zone of the hinterland orogenic domain of Iberian Massif, presents a magmatic, metamorphic and sedimentary complex evolution. One of its most distinctive features is the presence of two orogenic cycles (Cadomian and Variscan; Quesada, 1990; Ribeiro *et al.*, 2007; 2009; 2010), giving rise to three general stratigraphic successions (*e.g.* Quesada, 1990; Nance *et al.*, 2012; Araújo *et al.*, 2013; Moreira *et al.*, 2014):

- A Neoproterozoic sequence related to the Cadomian Cycle;
- A Lower Paleozoic anorogenic sequence related to the Rheic Ocean rifting and drifting;
- Syn-orogenic series of Lower-Middle Devonian to Carboniferous age linked to the Variscan convergence.

The magmatism also emphasizes three main different pulses: Neoproterozoic, Cambrian-Ordovician and Devonian-Carboniferous (*e.g.* Quesada, 1990; Moreira *et al.*, 2014). The geochemical signature and the temporal span of these rocks match the episodes recorded in the stratigraphic successions. The structure and metamorphism are complex, due to the superimposed of main Neoproterozoic and Paleozoic tectonometamorphic episodes with a heterogeneous distribution.

The classic southern limit of OMZ is emphasized by oceanic like rocks, which compose the Beja-Acebuches Amphibolites and, in their absence by the Ferreira-Ficalho-Almonaster thrust. These mafic rocks are interpreted either as an ophiolite complex (Quesada *et al.*, 1994; Fonseca *et al.*, 1999; Ribeiro *et al.*, 2010), or a narrow and very ephemeral realm of oceanic-like crust generated by mantle upwelling (*e.g.* Azor *et al.*, 2008).

The South Portuguese Zone (SPZ) is considered the SW foreland fold-and-thrust belt of Iberian Variscides, characterized by a thin-skinned SW facing structure, also emphasized by geophysical data (Ribeiro and Silva, 1983; Silva *et al.*, 1990; Simancas *et al.*, 2003; Ribeiro *et al.*, 2007). This zone comprises a stratigraphic sequence mainly composed of detrital rocks, occasionally with abundant magmatic rocks, with Devonian to Upper Carboniferous ages (Oliveira *et al.*, 2013). The deformation as a progressive behavior, which is older and more intense near the NE suture, and younger and less deformed towards SW (Ribeiro and Silva, 1983; Silva *et al.*, 1990; Dias and Basile, 2013). The metamorphism also presents a NE-SW progression, from greenschist facies at north and very low to absente in SW sectors (Oliveira *et al*., 2013).

IX.2.2.2. The Variscan Arcs in Iberia

Since the early regional studies it has become evident that the trend of the major geological structures have a strongly wavy pattern in Iberia. The remarkable pioneering geological map of Asturias by Schulz (1858), clearly shows a tight fold at northern Iberia scale (Fig. 2A). Such arc was detailed by the work of Barrois (1882) and used by Suess (1888), who was the first to recognize it as a mountain range bend. Soon it has become evident that this arcuate structure is not restricted to Iberia being part of a larger structure, the so-called Ibero-Armorican Arc (IAA) that continues in Brittany (Bertrand, 1887; Suess, 1888; Stille, 1924; Fig. 2B). Indeed, the NW-SE trend that predominates in most Iberia and rotates to a N-S orientation in NW Iberia, was assumed to continue in the E-W structures in Brittany (Choubert, 1935; Carey, 1955; 1958; Lotze, 1963; Cogné, 1967; 1971; Bard *et al.*, 1971; Lefort, 1979; Perroud and Bonhommet, 1981; Burg *et al.*, 1987). Although the continuity between the southern Iberian branch and the northern Armorican one is not possible to follow due to the opening of the oceanic rift of Biscay essentially during Upper Cretaceous (Ries, 1978), the arcuate structural pattern not only in the CZ, but also in the WALZ / CZ seems to confirm this interpretation (Staub, 1927, Fig. 2C; Carey, 1955, Fig. 2D). Indeed, here the continuity of the Variscan structures, completely underline an orocline known as the Asturian Knee (Staub, 1927; Julivert and Marcos, 1973; Julivert *et al.*, 1977), the Asturian Arc (Pérez-Estaún and Bastida, 1990; Aramburu and Garcia-Ramos, 1993; Ábalos *et al.*, 2002), the Cantabrian Arc (Parés and Van der Pluijm, 2004; Weil and Sussman, 2004), the Cantabrian - Asturian Arc (Parés *et al.*, 1994; Weil *et al.*, 2013) or, more recently, the Cantabrian orocline (Gutiérrez-Alonso *et al.*, 2008: 2011; Johnston *et al.*, 2013; Sengör, 2013; Weil *et al.*, 2013).

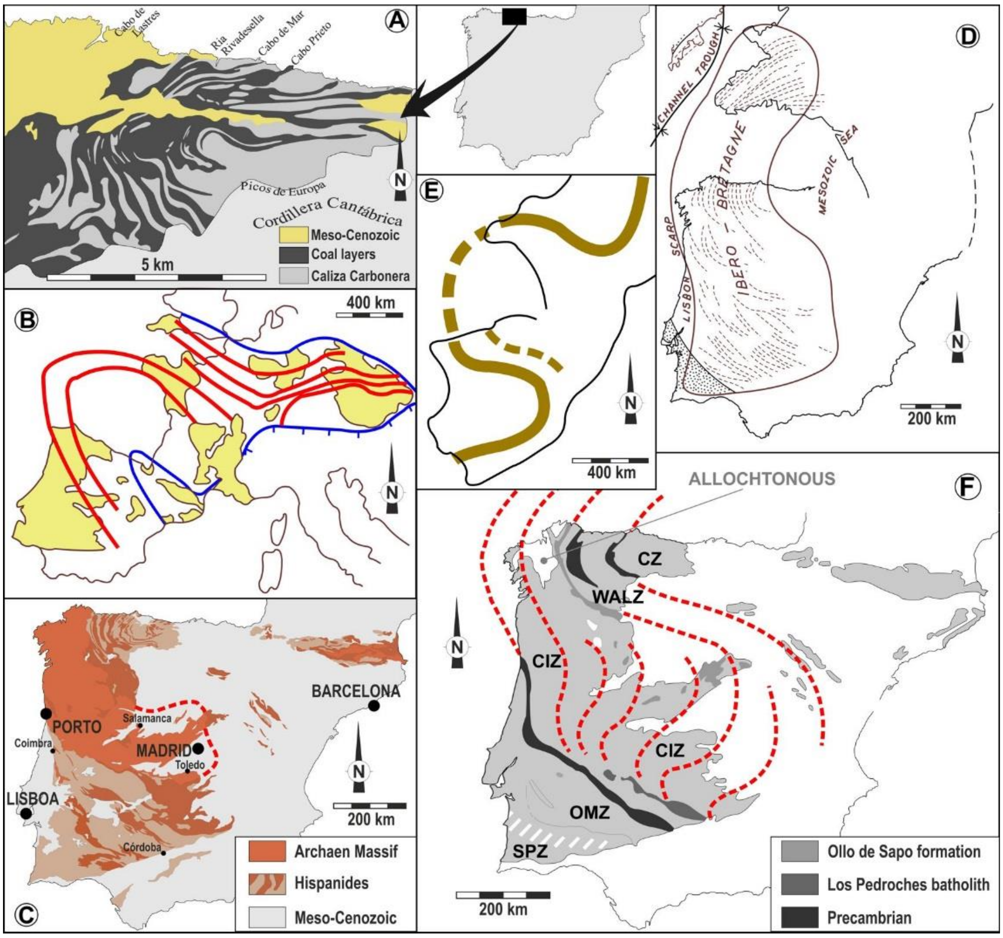

Figure 2 – Historical evolution of major Variscan Arcs proposed to Iberia:

A – Simplified version of the geological map of Eastern Asturias (redrawn from Schulz, 1858);

B – The structural continuity between Iberia and Brittany (Stile, 1924)

C – Asturian and Castilian major Iberian Arcs (Staub, 1927);

D – Matching geological structures across the Biscay Sea (redrawn from Carey, 1955);

E – Major structural arcuations in the Ibero - Brittany region (adapted from Du Toit, 1937);

F – Speculative sigmoidal pattern of the main Variscan structures in Iberia (adapted from Aerden, 2004).

As the Cantabrian Arc (CA) is located in the core of the IAA (Parés and Van der Pluijm, 2004) they are often considered the same structure, with a curvature increasing from external domains to its centre (Sengör, 2013). Such geometry led some authors to propose a common origin (Johnston *et al.*, 2013; Gutierrez-Alonso *et al.*, 2010; Sengör, 2013; Weil *et al.*, 2013; Martínez Catalán *et al.*, 2014).

Early works also emphasize another orocline in the Iberian Variscides, the Central-Iberian Arc (CIA). According to Staub (1927), the Caledonian and Hercynian folds have developed in lower metamorphic formations wraps around an Archaen core composed of schists and old granites, enhancing a Castilian Arc (Fig. 2C). This arc was also considered in the classical work of Du Toit concerning the world Palaeozoic fold systems (1937; Fig. 2E). Subsequent studies concerning the paleogeographic zoning of the Iberian Palaeozoic (*e.g.* Lotze, 1945; Julivert *et al.*, 1972) show that the original assumptions of Staub are no longer valid and the idea of a Castilian Arc was abandoned. More recently this major arcuate structure was again considered (Aerden, 2004). Indeed, extrapolating to SE the interpretation of a porphyroblasts study in NW Iberia he considered, in a "*still very speculative*" model (*op cit.* p. 194), that the observed structural relationships are apparently consistent with a late-Variscan sinistral transpression, delineating a sigmoidal shape (Fig. 2F) that partly follow the Staub proposal.

Since 2010 several papers consider the existence of the CIA or Central Iberian Orocline in the Iberian Variscides, but with two slightly different shapes:

- A short version where most of the southern branch is cut by the Tomar-Badajoz-Cordoba Shear Zone (Martínez Catalán 2011a; 2011b; 2011c; Simancas *et al.*, 2013; Martínez Catalán *et al.*, 2014; Fig. 3A)
- A long version, with the orocline assuming a major isoclinal shape and where the southern domain of the CIZ is considered a lateral equivalent of the WALZ (Johnston *et al.*, 2013; Weil *et al.*, 2013; Shaw *et al.*, 2012*a*; 2012*b*; 2014; Fig. 3B). The palinspatic restoration of both the Cantabrian-Central Iberian orocline pair, yields an initial linear ribbon of over 1500 km long (Shaw *et al.*, 2012*a*; 2012*b*) or even more than 2300 km (Shaw *et al.*, 2014; 2012b).

Although the coupled geometry of previous oroclines are sometimes considered the result of a coeval formation (Johnston *et al.*, 2013), some authors (*e.g.* Martínez Catalán *et al.*, 2014 and Simancas *et al.*, 2013) proposed that they could be slightly diachronic with the CIA older than the CA/IAA.

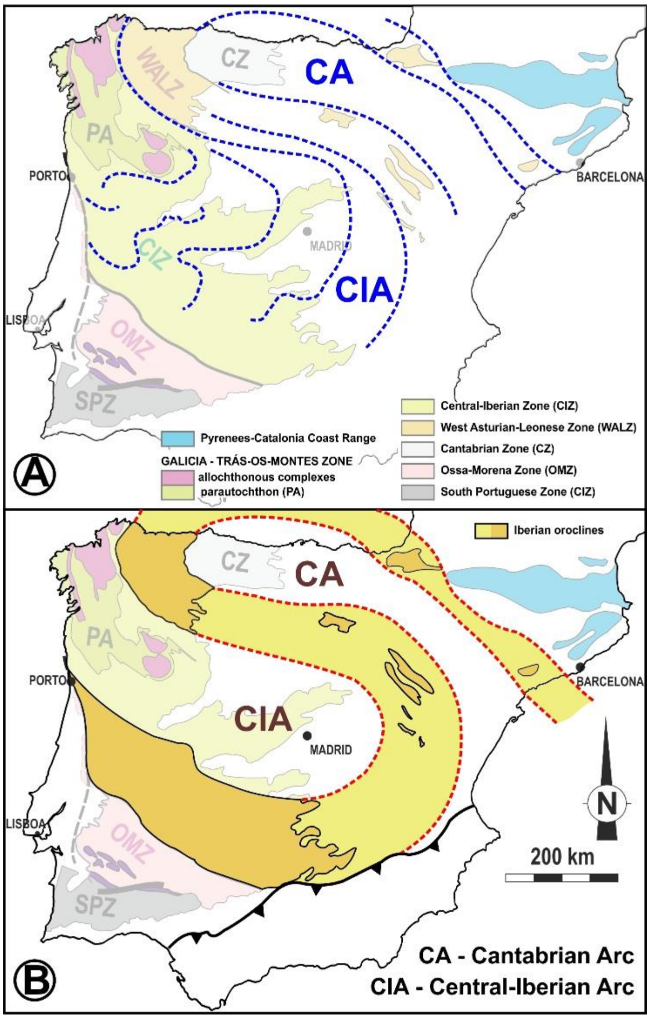

Figure 3 – The pattern of the Central-Iberian oroclines according to the:

A – short version (adapted from Martínez Catalán *et al.*, 2014);

B – long version (adapted from Johnston *et al.*, 2013).

IX.2.3. Reviewing the Data

In the last 40 years a huge amount of data concerning the Variscan geology of Iberia have been produced. A critical review of these data is fundamental to the main subject of this work.

IX.2.3.1. Variscan Folds and coeval Shear Zones

The strong shortening related to the Variscan orogeny gives rise to pervasive folding at all scales, well expressed in the lithostratigraphic units with higher competence (*e.g.* the Lower Ordovician quartzites).

Interference among folds in Iberia is often referred (*e.g.* Pérez-Estaún and Bea, 2004) and could be locally important. Nevertheless the general pattern at the Iberian scale is usually considered the result of the first and main Variscan tectonic event (D_1). However, this does not mean that all the folds are coeval because, not only there is a strong diachronism transversally to the orogen (Noronha *et al.*, 1981; Dallmeyer *et al*., 1997), but also their development could be slightly heterogeneous being older close to some major anisotropies either on the basement or on the cover.

Although the main trend of the regional D_1 folds have a simple arcuate pattern, geometrical and kinematical details emphasize a more complex behaviour, allowing the individualization of several sectors. We focus our analysis mostly on the CIZ, because this is a key sector to discuss the problem of the Iberian oroclines. It is not easy to make the Portuguese and Spanish data compatible mainly in the central and southern sectors of ZIC. This could partly reflect a change in the behaviour of the folds parallel to the trend of the orogen, related to the distance to the hinge zone of the IAA. Whatever the causes, six main distinct behaviours for the D_1 deformation in the CIZ could be emphasized (Fig. 4):

- Domain A (Ribeiro *et al.*, 1990; Dias, 1998; Moreira *et al.*, 2010a; Dias *et al.*, 2013b). It is a central segment where the D_1 Variscan deformation is very weak. The bedding is usually subhorizontal or presents open folds with subhorizontal hinges and subvertical axial planes. When present, the coeval cleavage is subvertical and spaced. The transition to the adjacent domains is sharp, usually marked by high dip sinistral shear zones sometimes with a thrusting component.
- Domain B (Ribeiro, 1974; Ribeiro *et al.*, 1990; Dias and Ribeiro, 1991; 1994; Pereira and Ribeiro, 1992; Pereira *et al.*, 1993; Dias, 1998; Moreira *et al.*, 2010a; Dias *et al.*, 2003; 2013b; Pamplona *et al.*, 2013). NE and N of this central segment the Variscan deformation strongly increases, becoming pervasive. The facing of the folds is towards NE and N and there is a continuous transition to the next domain. The pervasive S_1 cleavage is axial planar and has often developed a stretching lineation subparallel to the subhorizontal fold axes (*b* kinematical axes of Ramsay, 1967). Coeval of the D_1 folding, sinistral shear zones subparallel to the axial planes of the folds and a regional development have been developed. The finite strain ellipsoids estimated for the Armorican Quartzites are prolate.
- Domain C (Ribeiro, 1974; Díez Balda, 1986; Díez Balda *et al.*, 1990; Dias *et al.*, 2013b). Towards the NE foreland, despite preserving subhorizontal axes, the D_1 folds become recumbent with a NE to E facing. This geometry is similar to the WALZ Variscan general structure.
- Domain D (Ribeiro *et al.*, 1990; Dias and Ribeiro, 1994; 1995a; 1998; Dias, 1998; Dias *et al.*, 2013b). W and SW of the weakly deformed domain A, is found a narrow domain where the folds, still with a subhorizontal axes, have a monoclinic symmetry with a W to SW vergence. The deformation is intense and the axial plane S_1 cleavage is pervasive, mainly in the short and usually overturned limbs. As in domain B, there are frequent

sinistral shear zones with a trend parallel to the orientation of the D_1 folds, and the prolate finite strain ellipsoids are also dominant.

- Domain E (Ribeiro *et al.*, 1990; Dias, 1998; Metodiev *et al.*, 2009; Romão *et al.*, 2013). In most of the Portuguese southern sector of ZCI, the pervasive D_1 folds have subvertical axial planes and no clear vergence. The cleavage is usually present and a stretching lineation subperpendicular to the subhorizontal fold axes (*a* kinematical axes of Ramsay, 1967) predominates in deeply stepping axial plane cleavage. There is no evidence of the D_1 sinistral shear zones which are common farther north, and the finite strain ellipsoids are plane strain to slightly oblate.

- Domain F (Burg *et al.*, 1981; Azor *et al.*, 1994; Dias, 1998; Martínez Poyatos, 2002; Martínez Poyatos *et al.*, 2004; Pereira *et al.*, 2010). Adjacent to the boundary with the OMZ there is a narrow sector with an intense deformation and NE facing folds, which could attain recumbent shapes in the Spanish sector (the Puebla de la Reiña anticline and the Hornachos syncline). There is a penetrative S_1 axial plane foliation (often mylonitic), with a low dipping NW-SE stretching lineation subparallel to fold axes. In the Portuguese sector, the kinematical criteria indicate a predominant sinistral shear sense.

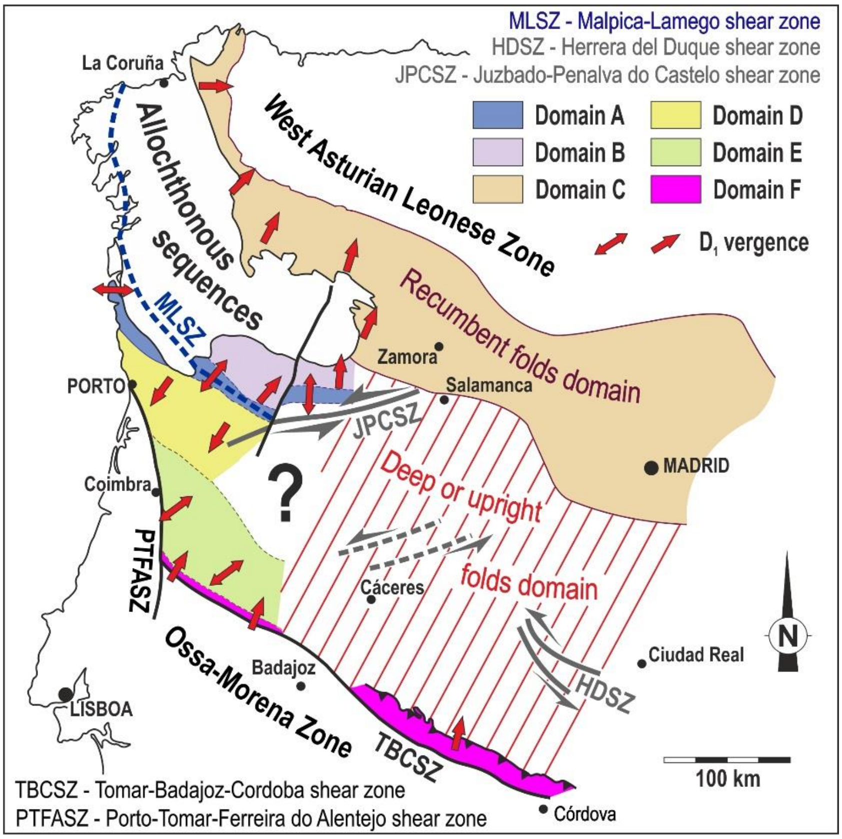

Figure 4 – Main fold pattern in the CIZ of NW Iberia (adapted from Azor *et al.*, 1994; Díez Balda *et al.*, 1990; Ribeiro *et al.*, 1990; Dias and Ribeiro, 1994; Dias, 1998; Dias *et al.*, 2013b; Romão *et al.*, 2013).

Despite the geometrical diversity of the regional D_1 folds of CIZ, their axes are always subhorizontal or low dipping. Such behaviour indicates either the lack of a pervasive superposition between Variscan deformation events, or a strong coaxiality of these events.

The predominance in northern Portugal of sinistral strike-slip shear zones subparallel to the axial planes of the D_1 folds, and the centripetal vergences around a less deformed sector has been interpreted as an asymmetrical positive flower structure (with a longer NE branch) centred in domain A (Dias, 1998; Moreira *et al.*, 2010a; Dias *et al.* 2013b) and developed in a transpressive regime (Dias and Ribeiro, 1994). This first order structure, which can be followed for more than 80 km (Moreira *et al.*, 2010a), was controlled by a Precambrian basement anisotropy (the Porto-Viseu-Guarda lineament; Dias, 1998) depicted by gravimetric (Mendes Victor *et al.*, 1993) and magnetic anomalies (Miranda, 1990; Miranda and Mendes Victor, 1990). Analogue modelling supports this interpretation (Richard and Cobbold, 1990). Towards NW this structure is in continuity with the Malpica-Tui unit, which seems to indicate the existence of an Early Variscan first order shear zone (the Malpica-Lamego shear zone - MLSZ in Fig. 4; Llana-Fúnez and Marcos, 2001; Pamplona *et al.*, 2016) controlling, not only the exhumation and emplacement of the high-pressure –low to intermediate temperature rocks of this unit (Llana-Fúnez and Marcos, 2002) but also the Variscan deformation in NW Iberia autochthon.

This CIZ flower structure is not observed SW of the first order Juzbado - Penalva do Castelo shear zone (JPCSZ; Fig. 4), showing that this structure has already been active since the beginning of the main folding event of the CIZ, as previously proposed (Iglésias and Ribeiro, 1981).

Previous geometrical and kinematical CIZ zoning must be expanded to the Iberian scale in order to understand its Variscan arcs. Indeed, the combination of the finite strain pattern of folded layers and the coeval kinematics is a powerful tool to discriminate among folding mechanisms (Dias and Ribeiro, 2008). Such approach has been used since the early models for the IAA (Matte and Ribeiro, 1975; Ries and Shackleton, 1976).

Using the relation of the finite strain axes with the coeval folds / thrusts, Matte and Ribeiro (1975) emphasize a major distinction between inner and external domains of the IAA, separated by a narrow transition zone less than 10 km wide (Ries and Shackleton, 1976). The inner domain (*i.e.* the Cantabrian, West Asturian-Leonese and previous domain C) is characterized by thrusts and folds with a well-developed vergence towards the core of the arc. The coeval maximum stretching lineation is subparallel to the *a* kinematical axes. In spite of this regional behaviour (well expressed in the Mondoñedo nappe unit), the complexity of the folding evolution could generate local anomalies as in Courel recumbent fold; here the stretching lineation coincides with the fold axes due to the superposition of an homogeneous strain after the active folding process (Bastida *et al.*, 2010). Concerning the external domain, the stretching lineation has been considered to be always subparallel to the pervasive subhorizontal fold axes. Such relation (stretching subparallel to the *b* kinematic axis), has been described not only for the Iberian branch of the IAA but also for the Armorican one. Nevertheless, Matte and Ribeiro (1975) emphasise a major contrast in both branches: while in the southern branch the folds are coeval of pervasive sinistral shears with the same trend of the folds, in Brittany the dominant regional shear is dextral. However, Audren *et al.* (1976) show that the regional shears in both branches are not synchronous: while in Iberia the sinistral shears are older than the Carboniferous granitic intrusions, the dextral shearing in Brittany is essentially coeval of this major magmatic event. Moreover as previously described, the external domain of Matte and Ribeiro (1975) presents a much more complex behaviour.

IX.2.3.2. Folding Events and Ages

The age of the main Variscan folds in Iberia is a key issue for the genesis of the Iberian Variscan arcs. This is not easy being necessary to distinguish between local and regional tectonic events, a problem enhanced by the heterogeneity of deformation. At a more regional scale, the orogeny migrates both transversally and longitudinally to structures. The transverse migration operates from the suture zones to the relative foreland and, in the autochthon, from the axis of positive flower structures to their branches. The longitudinal migration operates from the hinge of the IAA towards its flanks.

In spite of these limitations and some minor discrepancies, three main compressive tectonic events are usually recognized in the autochthon of NW Iberia Variscides (Gibbons and Moreno, 2000; Vera, 2004; Dias *et al.*, 2013a):

- D_1, the only that is pervasive, usually induces the formation of upright to slightly overturned structures in the more internal domains (*i.e.* most of the CIZ) and recumbent folds and thrusts facing towards the core of the CA in the most external ones (*i.e.* domain C of CIZ, as well as the WALZ end CZ). As expected, D_1 intensity usually increases from the foreland towards the hinterland, where a well-developed cleavage is pervasive.
- D_2 has a very heterogeneous development being restricted to the vicinity of the NW Iberian allochthonous and parautochthonous units. It corresponds to a second folding phase associated to shallow dipping shear zones, developing an S_2 axial planar cleavage to schistosity which can completely transpose S_1. This behaviour, as well as the similarities between the kinematics of the structures found in the nappes and in the

autochthon, show that this tectonic event has been induced by the emplacement of the napes.

- D_3 is also non penetrative, being expressed by upright folds often linked to subvertical strike-slip shear zones, like the major ENE-WSW Juzbado-Penalva do Castelo with a left lateral kinematics. The widespread granitic plutonism is also related to the D_3 tectonic event.

Since the recognition of sin-D_2 subhorizontal shear zones with displacements parallel to the orogenic trend in the Tormes gneiss dome (Escuder Viruete *et al.*, 1994; Escuder Viruete, 1998) and in Martinamor antiform (Díez Balda *et al.*, 1995), both in the vicinity of Salamanca, widespread similar low dipping fabrics has been described. They are generally more frequent in the vicinity of the parautochthonous /allochthonous units of the NW Iberia, being usually related with large antiforms and domes where outcrop high grade rocks (Escuder Viruete *et al.*, 2004). A recent review (Martínez Catalán *et al.*, 2014), emphasizes two major extensional events, the first one coeval with the end of D_2, and the other sin to post-D_3. Although most authors considered such events related to the extensional collapse of the orogen (Escuder Viruete *et al.*, 2004 Martínez Catalán *et al.*, 2014; Ballèvre *et al.*, 2014) for others (*e.g.* Ribeiro *et al.*, 2007; Dias *et al.*, 2013b), they could result from local compressions related to the allochthonous / autochthonous units

The Variscan orogeny in Iberia has recently been consider the result of a major Late Carboniferous collision followed by a Permian wrenching (Schulmann *et al.*, 2014). This is different from previous models where the deformation began in Lower/Middle to Upper Devonian times (*e.g.* Burg *et al.*, 1981; Noronha *et al.*, 1981; Matte 1986; Pérez-Estaún *et al.*, 1991; Azor *et al.*, 1994; Ábalos *et al.*, 2002; Franke *et al.*, 2005). The D_1 age (Fig. 5) is crucial to understand the Iberian arcs.

The 359 Ma age is often considered the beginning of the D_1 deformation in the CIZ autochthon (Martínez Catalán *et al.*, 2014; Ballèvre *et al.*, 2014). This age (Fig. 5B) was obtained (Dallmeyer *et al.*, 1997) by $^{40}Ar/^{39}Ar$ in white micas on S_1 cleavage NE of Morais Massif, in the normal limb of the Ollo de Sapo anticline (Fig. 5A). It is similar to the ages obtained in Somosierra also in the CIZ autochthon (354 and 353 Ma; Rubio Pascual *et al.*, 2013; Fig. 5B). Using the ages obtained farther E in the WALZ (336 and 322 Ma; Dallmeyer *et al.*, 1997) they estimate the diachronism related to the migration of the D_1 deformation towards the foreland (*circa* 23 Ma difference per 125 km of present distance). This value is similar to the range they found in D_2 mylonites. Thus, an average convergence rate of *circa* 1-2 cm/year was then proposed.

Such evolution is consistent with the estimations for the West Asturian-Leonese and Cantabrian zones based on stratigraphical evidences in synorogenic deposits (Pérez-Estaún *et al.*, 1988; Fig. 5B). While in the WALZ the deformation began in Lower Carboniferous (Pérez-Estaún, 1974) in the rearmost units of CZ it could be Westephalian B (*i.e.* Lower Moscovian; Arboleya, 1981) or even Namurian (*i.e.* Serpukhovian; Alonso *et al.*, 2009). The emplacement of nappes in the more external sectors of CZ persists until Stephanian (*i.e.* Kasimovian: Maas, 1974; Marquínez, 1978; Alonso *et al.*, 2009). Nevertheless, there is still some controversy with some works proposing that the foreland basin system had already developed during the Late Devonian (Keller *et al.*, 2008). This eastwards progression of NW Iberia autochthon deformation is consistent with the emplacement of the allochthonous nappes in the same sense (Fig. 5B): beginning of the accretion at 400-390 Ma with the old continental arc preserved in the Upper Allochthon (Arenas *et al.*, 2014), followed at 390-380 Ma by the Middle Allochthon unit with ophiolitic affinities (Ballèvre *et al.*, 2014) and the Lower Allochthon emplacement at 370-365 Ma (Ballèvre *et al.*, 2014).

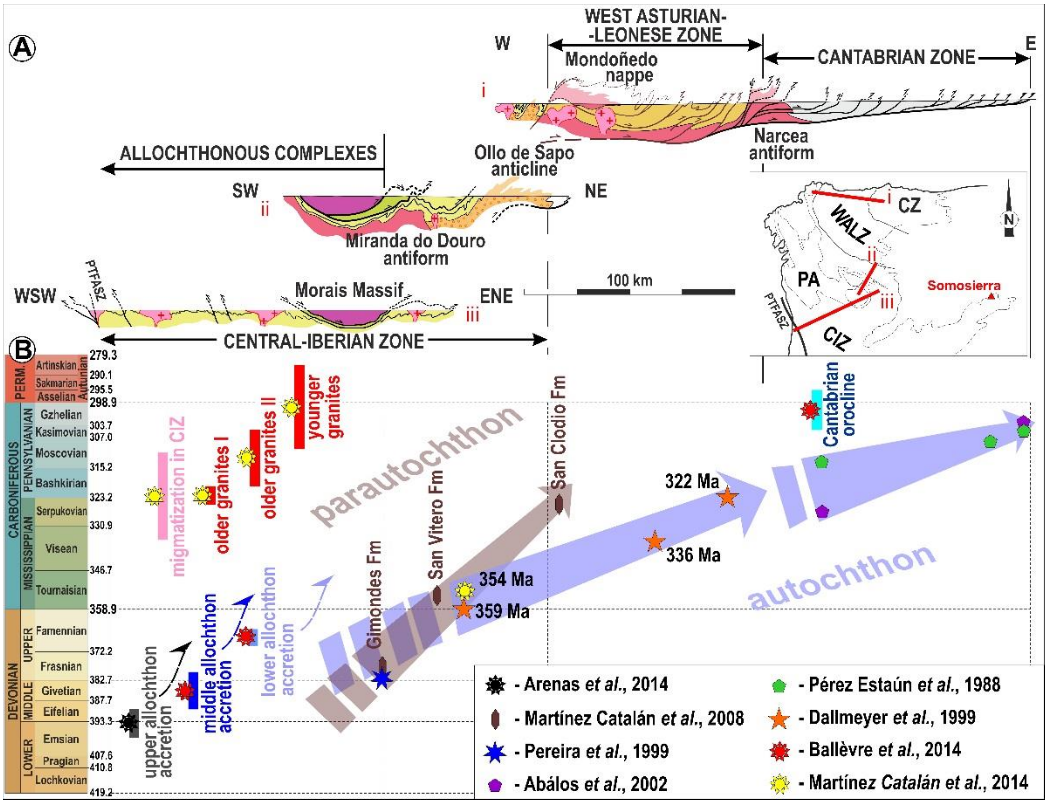

Figure 5 – Age distribution of main Variscan tectonic events in NW Iberia since Lower Devonian:

A – Simplified cross-sections emphasizing the main structural features of NW Iberia (adapted from Pérez Estaún *et al.*, 1991; Ribeiro *et al.*, 2005; Ballèvre *et al.*, 2014; Martínez Catalán *et al.*, 2014);

B – Main tectonic events evolution in the NW Iberia.

Such diachronism shows that it is not plausible to assume 359 Ma as the lower limit to the D_1 deformation in the CIZ autochthon, because such age has been obtained, at least 200 km away from any suture. Thus, the deformation must be older in the more far-travelling nappes rooted there and in the inner domains of the autochthon. This older age for the D_1 deformation in the autochthon is supported by further evidence:

(1) The Gimondes, San Vitero and San Clodio formations are considered flyschoid deposits related to the frontal thrusts of the allochthonous / parautochthonous unit of NW Iberia (Martínez Catalán *et al.*, 2008). NE of the Bragança Massif, the paleontological contend of the Gimondes formation (Ribeiro, 1974) indicates an age close to the Givetian - Frasnian boundary (Fig. 6B; Teixeira and Pais, 1973; Pereira *et al.*, 1999). This unit presents pebbles from the allochthonous complexes showing that they were already exhumed (Ribeiro and Ribeiro, 1974). Based on the youngest zircon contend of these three synorogenic deposits, it was emphasized a diachronism of the parautochthon deformation (Martínez Catalán *et al.*, 2008) with the younger ages towards the more external domains of the orogen (Fig 6B; 378 ± 6 Ma for Gimondes; 355 ± 8 Ma for San Vitero and 324 ± 7 Ma for San Clodio). Such behaviour is compatible with the observed in the autochthon.

(2) In the inner domains of CIZ, D_1 regional folds are cut by the imbricated basal thrusts of parautochthon (*e.g.* N of Marão; Pereira, 1987; 1989).

(3) The emplacement of the parautochthon in northern Portugal is subparallel to the trend of major D_1 folds and coeval sinistral strike slip shear zones (Pereira, 1987; 1989; Ribeiro *et al.*, 1990; Rodrigues *et al.*, 2005; Rodrigues, 2008; Rodrigues *et al.*, 2013). This indicates an already well structured autochthon where major anisotropies were reactivated has lateral ramps of the nappes. Such model is common in other orogenic domains, as in the Appalaches (Pohn, 2000) or the Greenville of Canada (Dufréchou, *et al.*, 2014).

(4) The stable platform environment that predominate in the CIZ since the Early Cambrian change by the end of Lower Devonian (Martínez Catalán *et al.*, 2008; Ballèvre *et al.*, 2014; Martínez Catalán *et al.*, 2014). The absence of Middle Devonian in northern Portugal is probably related to the beginning of the Variscan deformation in the CIZ autochthon (Pereira, 1988).

The beginning of a pervasive and intense deformation in the innermost sectors of CIZ, at least since the Middle / Upper Devonian, is also coherent with Iberian geodynamics. In the OMZ, the stratigraphic, metamorphic and magmatic data shows that the beginning of subduction, and consequently the first deformation episode, had Lower Devonian age. Indeed in Odivelas (southernmost domains of OMZ) the Emsian-Eifelian limestones, which are spatially associated with volcanic rocks with tholeiitic orogenic signature, show evidences of syn-sedimentary deformation (Machado *et al.*, 2009; 2010; Moreira *et al.*, 2010b; Silva *et al.*, 2011). Also in the OMZ the Lower Devonian Terena formation, in which the D_1 is absent, shows evidences of syn-sedimentary deformation (*e.g.* Oliveira *et al.*, 1991; Araújo *et al.*, 2013). Geochronological data (Pereira *et al.*, 2012; Braid *et al.*, 2011; Rodrigues *et al.*, 2014) also show clusters of inherited zircons around Lower-Middle Devonian, which seems to indicate that the orogenic magmatic processes was active during lower Devonian (*e.g.* Moreira *et al.*, 2014). The first deformation events in northern and central domains of OMZ are associated with a sinistral kinematic component (*e.g.* Expósito *et al.*, 2002; Araújo *et al.*, 2013). In the TBCSZ, where a pervasive sinistral kinematics is described (*e.g.* Ábalos, 1992; Quesada and Dallmeyer, 1994), Upper Devonian metamorphic ages were obtained in Neoproterozoic volcanic rocks (370 and 360 Ma by $^{40}Ar/^{39}Ar$ in amphibole; Quesada and Dallmeyer, 1994).

Thus, the Iberian Variscides is the result of a Lower Devonian - Upper Carboniferous complex evolution like its northeast extension and not an essentially Late Carboniferous collision as recently proposed (*e.g.* Schulmann *et al.*, 2014).

IX.2.3.3. The Regional CIZ Folding

Most of the regional folds of the CIZ autochthon have been traditionally attributed to the main D_1 Variscan shortening (Fig. 6A). However, recently a drastically different proposal was presented (Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014), where most of these folds are now considered to related to the D_3 event (Fig. 6B). This new definition of the folding events in CIZ is one of the main arguments favouring the existence of the CIA. The age of these CIZ folds is thus a key issue.

All the recent syntheses of the Iberian Geology, either from the Spanish groups (Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004) or from the Portuguese ones (Dias *et al.*, 2013b; Romão *et al.*, 2013) consider that these regional folds were formed during the early D_1 Variscan phase. A similar conclusion is obtained comparing the great number of works supporting their formation during D_1, with the scarcity of papers considering they are D_3 (Table I).

In order to discriminate between both models (*i.e.* Fig. 6A versus Fig. 6B) three situations will be discussed (see location in Fig. 6B): the Marão complex folded structure in the vicinity of the allochthon / parautochthon of northern Portugal, the Amêndoa - Carvoeiro syncline close to the intersection between the first order PTFA and TBC shear zones and the Mora - Madridejos region south of Toledo.

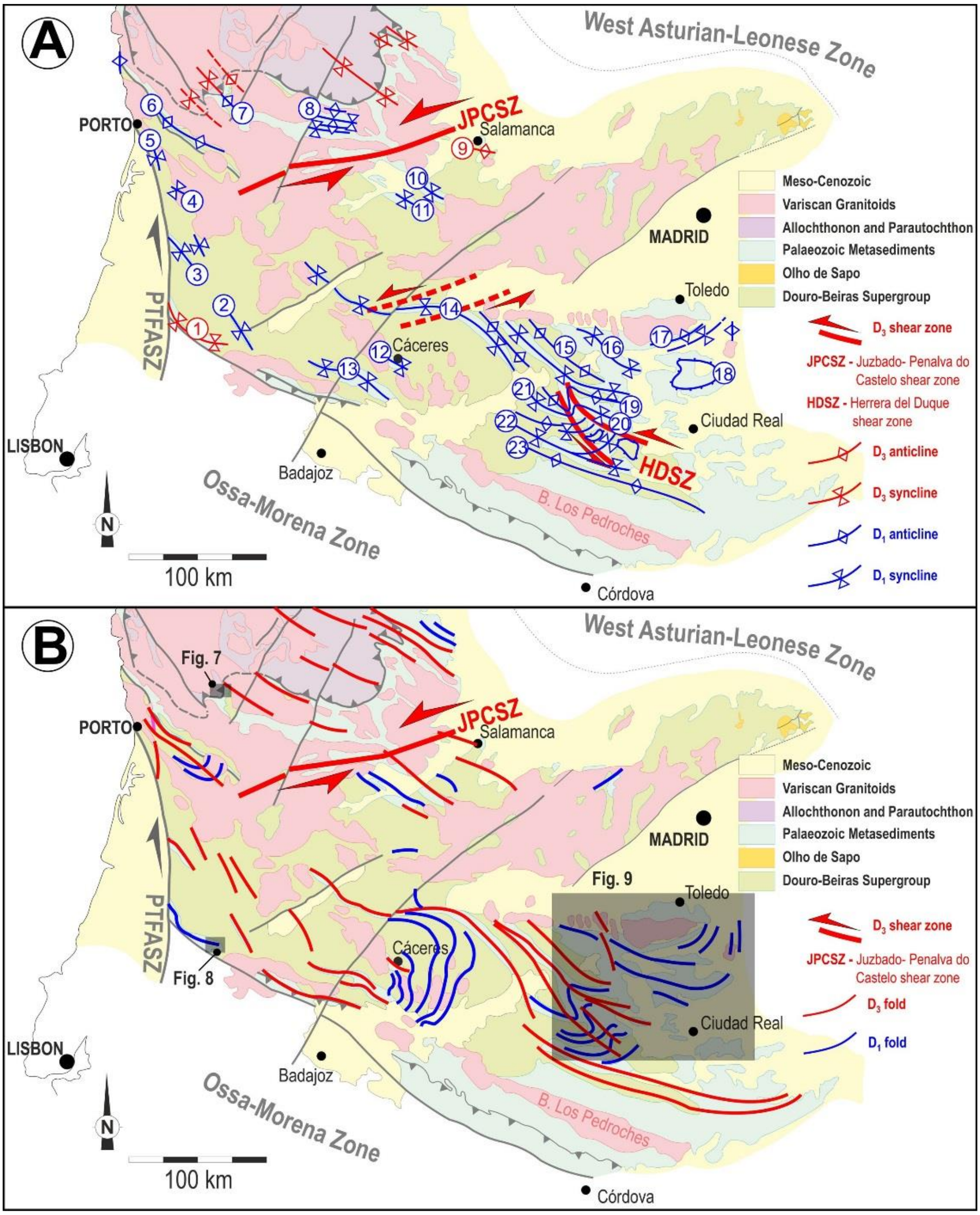

Figure 6 – Central-Iberian Zone major folds according to their Variscan tectonic event:

A – Classical interpretation (based in Díez Balda *et al.*, 1990; Ribeiro *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004; Dias *et al.*, 2013b; Romão *et al.*, 2013). See table I for the association between the numbers inside the circles and the structures;

B – Recent interpretation emphasizing the Central-Iberian Arc (based in Martínez Catalán 2011a; 2011b; Martínez Catalán *et al.*, 2014), with locations of figures 7, 8 and 9.

Table I – Relation between main CIZ Variscan folds and regional tectonic events according to different authors.

Ref.	structure	D_1	D_3
1	Amêndoa-Carvoeiro syncline	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014	Romão, 2000; Romão *et al.*, 2013
2	Vila Velha de Ródão syncline	Ribeiro *et al.*, 1990; Dias, 1994; Metodiev *et al.*, 2009; Romão *et al.*, 2013	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
3	Buçaco syncline	Ribeiro *et al.*, 1990; Dias and Ribeiro, 1993; Dias, 1994; Dias *et al.*, 2013b; Romão *et al.*, 2013	Martínez Catalán, 2011a; 2011b; 2011c, Martínez Catalán *et al.*, 2014
4	Caramulo syncline	Ribeiro *et al.*, 1990; Dias, 1994; Dias and Ribeiro, 1995a; Dias *et al.*, 2013b; Romão *et al.*, 2013	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
5	Oliveira de Azeméis syncline	Ribeiro *et al.*, 1990; Dias *et al.*, 2013b	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
6	Valongo anticline	Ribeiro *et al.*, 1990; Pereira and Ribeiro, 1992; Dias, 1994; Dias and Ribeiro, 1998; Dias *et al.*, 2013b; Pamplona and Ribeiro, 2013	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014; Valle Aguado, 1992
7	Marão anticline	Ribeiro *et al.*, 1990; Dias, 1994; Coke, 2000; Coke *et al.*, 2003; Rodrigues *et al.*, 2005; Moreira *et al.*, 2010a; Dias *et al.*, 2013b	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
8	Moncorvo and Poiares synclines	Díez Balda *et al.*, 1990; Ribeiro *et al.*, 1990; Dias and Ribeiro, 1991; Dias, 1994; Dias *et al.*, 2003; 2013b; Rodrigues *et al.*, 2005; Moreira *et al.*, 2010a	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014; Dias da Silva, 2013
9	Martinamor anticline		Díez Balda, 1986; Díez Balda *et al.*, 1990; Martínez Poyatos *et al.*, 2004 Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
10	Tamames and Salamanca synclines	Díez Balda, 1986; Rölz, 1975; Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004 Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014	
11	Peña de Francia syncline	Rölz, 1975; Macaya, 1981; Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
12	Cáceres syncline	Tena, 1980; Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
13	Sierra de S. Pedro syncline	Bascones *et al.*, 1980; Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	
14	Cañaveral syncline	Díez Balda *et al.*, 1990; Gil Toja and Pardo Alonso, 1991; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
15	Valdelacasa anticline	Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
16	Navalucillos syncline	Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
17	Mora anticline	Días Balda *et al.*, 1992; Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014	
18	Urda dome	Díez Balda *et al.*, 1990	
19	Navalpiño anticline	Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
20	Puebla de D. Rodrigo syncline	Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
21	Herrera del Duque syncline	Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004; Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014	
22	Almadén syncline	Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014
23	Alcudia anticline	Díez Balda *et al.*, 1990; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004	Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014

(1) In the region around Marão the regional Variscan fold pattern in the autochthon is well expressed in the Armorican quartzites formation. These folds have always been considered as D_1 (Pereira and Ribeiro, 1983; Ribeiro *et al.*, 1990; Pereira *et al.*, 1993; Coke *et al.*, 2000; 2003; Rodrigues *et al.*, 2005). Such conclusion is well supported because some of the folds are truncated by the basal thrusts of the NW Iberia parautochthon (Fig. 7; Pereira 1987; 1989), related with the nappe emplaced during the regional Variscan D_2 (Pereira 1987; Ribeiro *et al.*, 1990; 2007). These D_2 thrusts were locally folded during the D_3 shortening, a less intense and not pervasive deformation event (Pereira 1987; 1989). Nevertheless this clear cartographic interference pattern, the major D_1 Marão folds have been recently considered (Martínez Catalán, 2011a; 2011b; 2011c; Martínez Catalán *et al.*, 2014) due to the regional D_3 Variscan event (Fig. 7B), although they did not present any arguments to support this new interpretation.

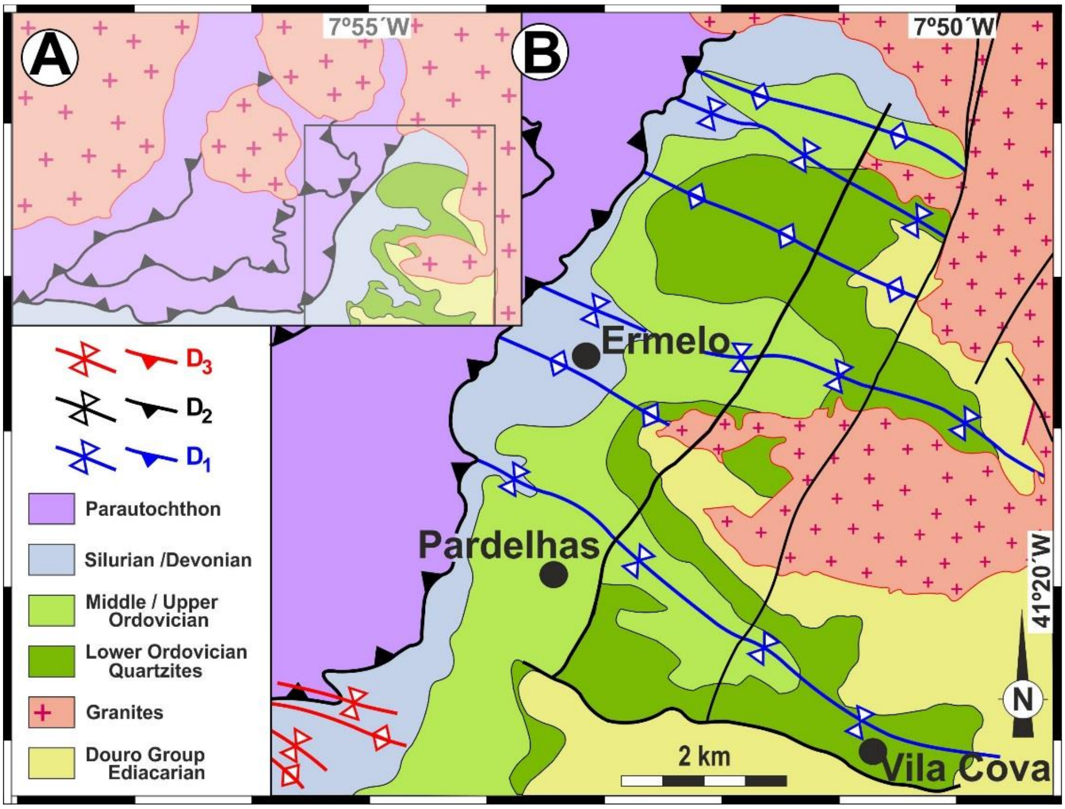

Figure 7 – General structural map northern of Marão Mountain showing the interference between D_1 and D_3 folds with D_2 thrusts (adapted from Pereira, 1989).

(2) In the models supporting the CIA, the E-W Amêndoa-Carvoeiro syncline (Fig. 8) has been considered (Martínez Catalán 2011a; 2011b; Martínez Catalán *et al.*, 2014) a D_1 structure deformed during D_3 by regional folds and the movement of the PTFASZ (Fig. 7B). Nevertheless, these authors never present any justification for such assumption.

Moreover, previous studies (Romão, 2000; Romão *et al.*, 2013; Fig. 8A), show a complex cartographic pattern due to the overprint of, at least three major tectonic events:

1º. NW-SE D_1 folds and thrusts with a NE facing (Fig. 8A);

2º. NNW-SSE D_2 folds and thrusts with a ENE facing (Fig. 8A) and clear interference patterns with D_1 structures (Fig. 8B);

3º. E-W D_3 folds and coeval thrusts, facing towards N (Fig. 8A). The interference with previous D_1 and D_2 structures (Fig. 8C) shows that the major Amêndoa syncline must be considered a D_3 structure and not D_1.

It should be emphasized that this local D_3 tectonic event expressed in the Amêndoa - Carvoeiro syncline, could be related to the Late Variscan deformation.

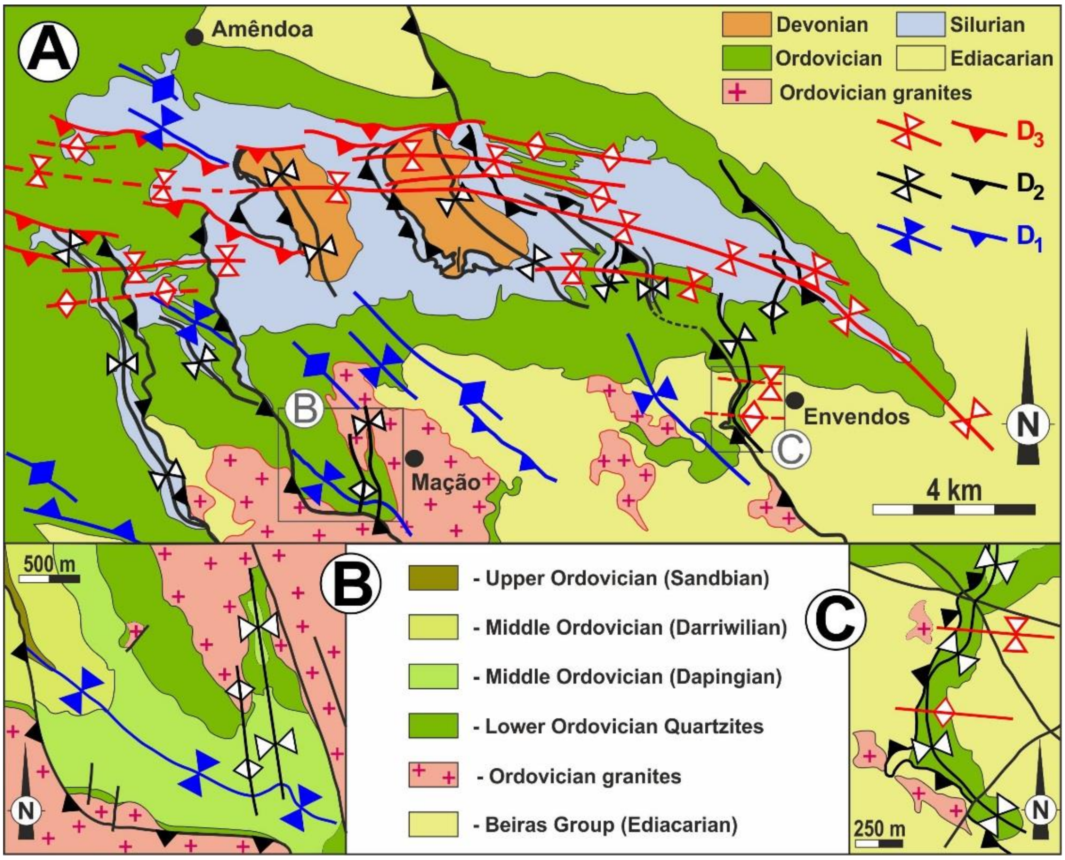

Figure 8 – The Amêndoa - Carvoeiro syncline structure:

A – General structural map (adapted from Romão, 2000);

B – D_1 syncline folded by D_2 structures at Monte de João Dias, W of Mação city (adapted from Romão, 2000);

C – D_2 structures folded by younger E-W D_3 folds at Sanguinheira sector, W of Envendos village (adapted from Romão, 2000).

(3) The Mora - Madridejos region is very important because it is close to the hinge zone of the supposed CIA. Detailed mapping has recently been presented trying to emphasize the regional development of D_3 folds in this sector (Fig. 9A; Martínez Catalán, 2011c). Nevertheless, the observed rotation of the D_1 folds from the regional NW-SE trend to the local N-S was already known (Díez Balda *et al.*, 1990; Díez Balda and Vegas, 1992; Julivert and Martínez, 1983; Julivert *et al.*, 1983) and its meaning is debatable. According to Martínez Catalán (2011c) it is the result of the intense D_3 folding related to the CIA (Figs. 6B, 9C). However, when considering the proximity to the D_3 Herrera del Duque shear zone (Fig. 9C; Martínez Poyatos *et al.*, 2004) such rotation could reflect the presence of another WNW-ESE D_3 sinistral shear zone (Fig. 9D).

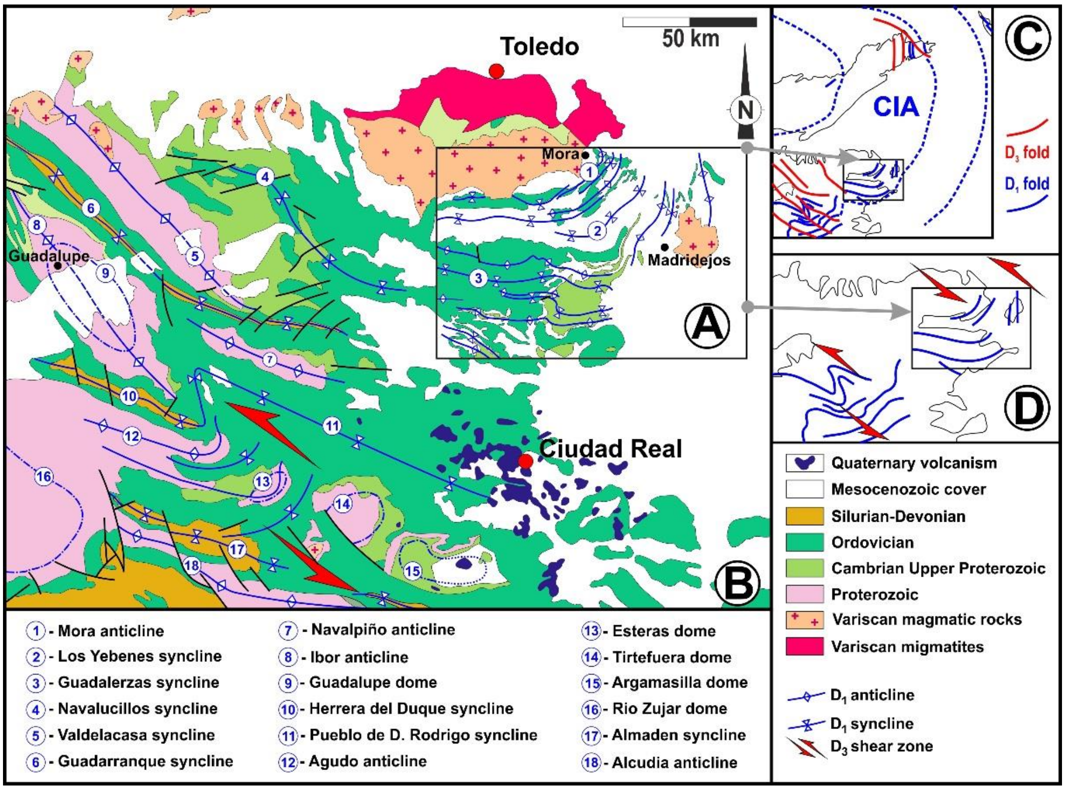

Figure 9 – Variscan structural map of Toledo - Ciudad Real region:

A – Rotation of main D_1 structures at Mora - Madridejos sector (simplified from Martínez Catalán, 2011c);

B – Geological map, emphasizing the general pattern of D_1 major folds (based on Julivert *et al.*, 1983; Días Balda *et al.*, 1992; Rodríguez-Fernández, 2004);

C – Mora - Madridejos D_1 structures in the framework of the CIA (redrawn from Martínez Catalán, 2011c);

D – Deflection of D_1 folds in Herrera del Duque and Mora sectors induced by D_3 WNW-ESE sinistral shear zones.

It is not easy to choose between both models because the deformation occurred at a low structural level without a pervasive cleavage and thus, the interference structures are rare. Moreover, whatever the proposals, the Cenozoic sediments of the Tagus basin hide the northern continuation of the N-S Mora structures (Fig. 9A). Thus, the deflection of the Variscan structures in the Mora - Madridejos region could not be used to support the existence of a pervasive regional D_3 folding (Fig 6B). Nevertheless, we favour the existence of a discrete D_3 sinistral shear zone, not only due to the vicinity of the Herrera de Duque one, but also due to the proximity to the major TBCSZ (Fig. 1).

As a major conclusion, we consider that the structural data show that the only pervasive Variscan tectonic event at the CIZ is D_1. D_2 and D_3 have an heterogeneous spatial distribution and are mostly important in the northern sectors (Julivert *et al.*, 1972: Díez Balda *et al.*, 1990; Ribeiro *et al.*, 1990; 2007; Ábalos *et al.*, 2002; Martínez Poyatos *et al.*, 2004; Rodrigues *et al., 2005; 2013;* Ribeiro *et al.*, 2013).

IX.2.3.4. Lithostratigraphic constraints in Pre-Orogenic Sequences

The lithostratigraphic Palaeozoic data are fundamental when trying to reconstruct the original shape of the proposed CIA where the continuity of the outcrops is often hidden by the overlap of younger sediments. This is particularly important in the "long-version" of the arc (Fig. 3B) where the southern domain of the CIZ (*i.e.* the Luso-Alcudian Zone of Lotze, 1945) is considered the lateral equivalent of the WALZ (Shaw *et al.*, 2012*a*; 2012*b*; 2014; Johnston *et al.*, 2013; Weil *et al.*, 2013). Thus, 2 critical review of the lithostratigraphy around the Iberian oroclines is thus fundamental to check the strength of previous proposals:

- To explain the different transition between the Lower Ordovician and older sequences in the WALZ (where there is stratigraphic continuity) and in CIZ (where a clear unconformity is pervasive), a possible "topographic high that was more proximal to the southern portion of the Gondwana margin" was suggested (Shaw *et al.*, 2012*a*), but no evidence of it has been presented.
- The correlation between (Shaw *et al.*, 2012*a*) the Lower Ordovician Ollo the Sapo volcanism (close to the boundary between CIZ and WALZ) and the Urra formation (in the vicinity between CIZ and OMZ) cannot be sustained. Indeed, in the proposed pattern of the CIA, the lateral equivalent of Ollo de Sapo must be in the central domains of CIZ (Fig. 3B).
- The S-pattern of paired CA and CIA oroclines is not compatible with the presence of the larger IAA. Indeed the continuity of the southern CIZ and OMZ in the French Armorican massif should imply a refolding of the Iberian oroclines (Shaw *et al.*, 2012*a*). Any model

constraining the continuity between the Iberian and Armorican branches of the IAA is difficult to accept due, not only to the strong correlation not only between the autochthonous formations of CIZ (*e.g.* the Lower Ordovician from Crozon in West Brittany and Buçaco in central Portugal; Robardet, 2002), but also between the allochthonous units exposed in the NW Iberian Massif and the southern Armorican Massif (Ballèvre *et al.*, 2014).

Previous correlations of paired CA and CIA oroclines are unable to propose any southward lateral correlation to the CZ, because OMZ or SPZ could not be considered its lateral equivalents (Shaw *et al.*, 2012*a; Fig. 3*). Several possibilities have been proposed to explain this major constraint: the vanishing of CZ along the strike (Shaw *et al.*, 2015), the possibility of OMZ and SPZ representing cryptic nappes not preserved in the core of the CA (Shaw *et al.*, 2012*a*) or an important offs*et al*ong the TBCSZ (Shaw *et al.*, 2012*a*). However, these proposals are never supported by observed data.

As a major conclusion, the existence of the CIA puts major lithostratigraphic constrains, which are difficult to solve.

IX.2.3.5. Lower Ordovician Paleocurrents

The Lower Ordovician paleocurrents in the CZ were considered as a western deposition in a N-S linear basin subsequently folded during the formation of the CA (Aramburu and Garcia-Ramos, 1993). Paleocurrent studies in Lower Ordovician rocks have recently extended to most of the NW Iberian autochthon. Although several papers have been published (Johnston *et al.*, 2013; Shaw *et al.*, 2012a; 2012b; 2014; Weil *et al.*, 2013) they all use the same data set concerning 50 localities from CZ, WALZ and CIZ. All these works considered the data compatible with a model of a long N-S linear basin (> 1500 km or even > 2300 km) filled from E. This basin was folded by a Carboniferous-Permian N-S shortening giving rise to two major oroclines (*e.g.* Johnston *et al.*, 2013; Weil *et al.*, 2013): the CA, preserving a centrifugal pattern of the Lower Ordovician paleocurrents, and the CIA with a centripetal pattern (Fig. 10A). However, the visual inspection of the published Rose diagrams of the paleocurrents, shows a great dispersion of the data in both branches of the CIA (sectors C and D in Fig. 10A). This led us to rework the same data (available at Shaw *et al.*, 2012a, *doi:10.1016/j.epsl.2012.02.014*), using a slightly different approach. When the trend of the arc limbs (sectors A, B, C and D of Fig. 11A) is projected against the expected paleocurrents (*i.e.* orthogonal to the trend), several linear segments are expected, one for each limb (dashed lines in Fig. 10B). When in this graphic the principal paleocurrents for each locality are projected, their dispersion around the theoretical dashed linear segments is a measure of the strength of the proposed model.

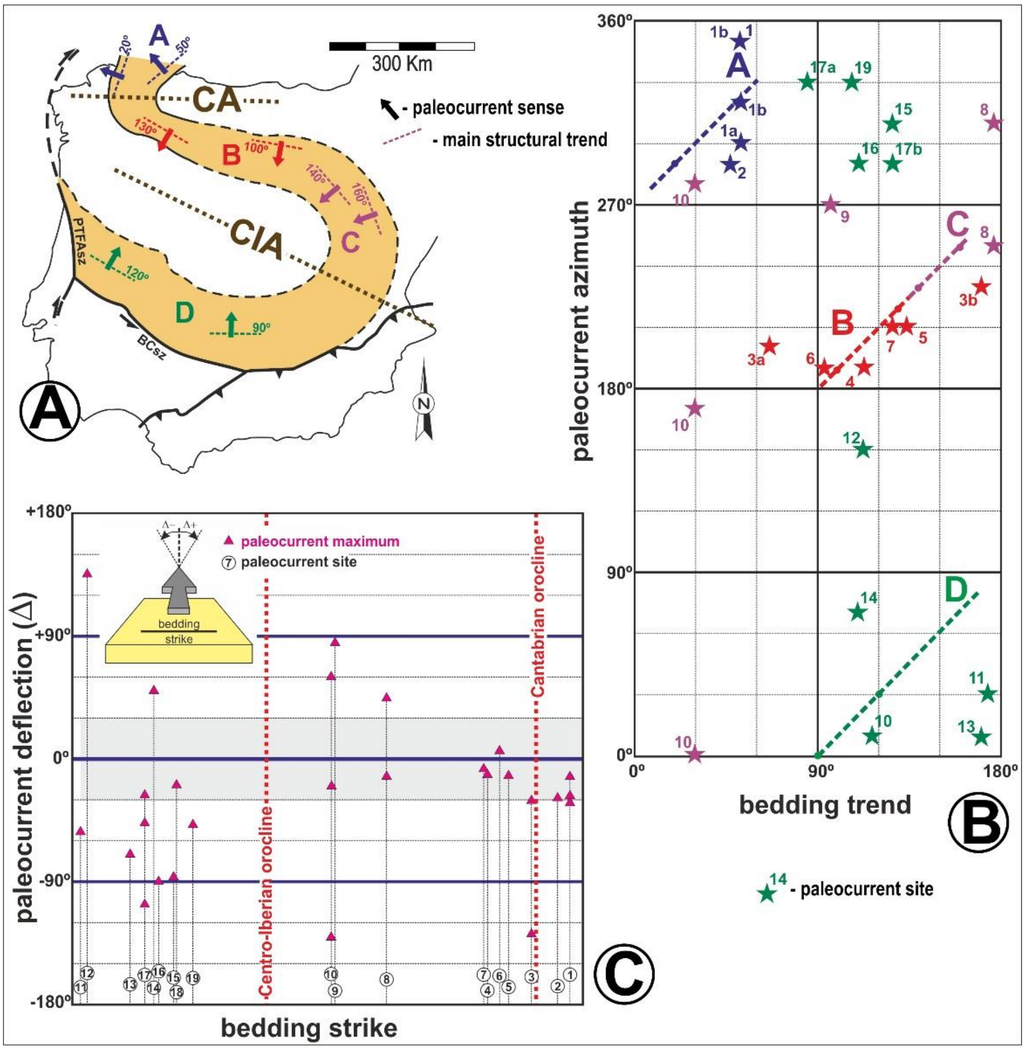

Figure 10 – The Lower Ordovician paleocurrents pattern around the Iberian oroclines:

A – Theoretical relation between paleocurrents and Variscan trend assuming an early N-S basin filled from E;

B – Predicted paleocurrent azimuths versus observed values (the numbers refer to sample localities as in Shaw *et al.*, 2012);

C – Divergence between the predicted and observed paleocurrents (the numbers refer to sample localities as in Shaw *et al.*, 2012).

The results (Fig. 10B and C) show that the paleocurrents around the Cantabrian orocline are compatible with a secondary origin for the arc. Indeed, there is a good linear correlation between the data and the theoretical predictions, either for the northern limb (sector A in fig. 10B) or the southern limb (sector B in fig. 10B). However, such behaviour is not found in the CIA data. Indeed, the great dispersion of the paleocurrents (sectors C and D in fig. 10B) seems to indicate that they could not be the result of uniform sedimentation in a N-S basin filled from East. The same conclusion was obtained when the regional structural trend is projected against the angular deflection (Δ) away from the perpendicularity to the strike of the bedding where the paleocurrent has been measured (Fig. 10C). While in both limbs of the CA the deflection is lower than 30º, in the CIA the deflection is much higher.

Some early paleocurrent studies in Lower Ordovician Quartzites of the Portuguese southern sectors of CIZ (Conde, 1966) are also not compatible with the CIA. Indeed, although located in the southern limb (sector D of Fig. 10A) they emphasize paleocurrents towards NW instead of NE.

In conclusion, although the Lower Ordovician paleocurrents support the existence of the Cantabrian orocline, they cannot be used to sustain the Central Iberian one as previously considered by several authors (Johnston *et al.*, 2013; Shaw *et al.*, 2012; 2013; 2014; Weil *et al.*, 2013).

IX.2.3.6. Variscan Paleomagnetism in Iberia

Paleomagnetism has always been considered fundamental to understand the Iberian Variscan arcs (Ries *et al.*, 1980; Bonhommet *et al.*, 1981; Perroud and Bonhommet, 1981; Hirt *et al.*, 1992). These early data demonstrate that, at least some of the CA curvature is secondary due to the Variscan deformation. Recent works show the weakness of this conclusion, not only because it was based on very few data, but mostly due to the inability of the used methodology to differentiate between secondary syn-tectonic and post-tectonic remagnetizations (Weil *et al.*, 2013). Such limitations have been overcome in more recent studies, mostly concentrated on the CA (see Weil *et al.*, 2013 for a detailed resume), which show:

- The oroclinal occurred after the regional orogenic folding / thrusting has been nearly completed;
- Undoing the orocline leads to a nearly straight original belt with a N-S trend;
- The bending occurred primarily about near-vertical axes, without much further tilting of the already folded thrust sheets;
- The orocline formed between the Late Carboniferous and the Early Permian (Moscovian to Asselian, *i.e.* around 310–297 Ma; Weil, 2006).

Although there is some consensus in previous statements, it is still debatable how far the orocline rotation perpetuates downward (Van der Voo, 2004). Moreover, several models assume that it seems dynamically impossible for the Cantabrian orocline to be a thin-skinned crustal structure, due to space problems related to a 180º bent of a previous linear belt (Weil *et al.*, 2013). However, it is even harder to understand how this difficulty is solved by the bending of all the lithosphere as proposed by these models.

The CIA paleomagnetic data are much scarcer. However, recent data from the supposed southern limb of this orocline (Pastor-Gálan *et al.*, 2015), show that the rotation during the Late Carboniferous - Early Permian is comparable to the one estimated in the southern limb of the CA. This indicates that, either this arc did not exist, or it must be older than the CA, being an inherited major structure (Pastor Gálan *et al.*, 2015).

IX.2.4. Iberian Arcs; reviewing the Models

The complexity of the first order arcuate Variscan structures in Iberia, gave rise to a wide range of interpretative models.. A brief critical review of the geodynamical mechanisms that have been used to explain the Iberian arc(s), helps to narrow the range of possibilities.

IX.2.4.1. How many Arcs?

When discussing the mechanisms that have induced the formation of the first order arcuate structures in Iberia, it is fundamental to identify how many Variscan Arcs could be recognized in Iberia.

IX.2.4.1.1. Cantabrian Arc

The CA, sometimes called Asturian Arc (Pérez-Estaún and Bastida, 1990; Aramburu and Garcia-Ramos, 1993; Ábalos *et al.*, 2002), is the only unquestionable Variscan Arc in Iberia. Indeed, due to the almost continuous outcrops of CZ, it is possible to follow an almost 180º trend rotation of the Variscan structures. Moreover, the existence of this arc is well supported by several other data, (discuss in detail in Alonso *et al.*, 2007; Weil *et al.*, 2013) like the centripetal trend of Lower Palaeozoic sediments (Aramburu and Garcia-Ramos, 1993) and coeval paleocurrents (Shaw *et al.*, 2012; Fig. 10), the spatial and temporal pattern of tensile joints in rock units with different ages (Pastor-Galán *et al.*, 2011) or paleomagnetic studies (Weil, 2006; Weil *et al.*, 2013). The CA mostly overlaps the region where the Variscan structures are related to a stretching lineation subparallel to the *a* kinematical axis (Matte & Ribeiro, 1975; Ries and Shackleton, 1976).

IX.2.4.1.2. Central-Iberian Arc

Although since 2010 this orocline became very popular (Johnston, *et al.*, 2013; Martínez Catalán, 2011a; 2001b; 2011c; Martínez Catalán *et al.*, 2014; Shaw *et al.*, 2012a; 2012b; 2014; 2015) its existence is highly doubtful:

- There is no continuity of the main structures from one branch to the other, because the supposed hinge zone is hidden below Cenozoic basins (Fig. 3).

- The supposed arcuate pattern of the geophysical anomalies is not clear. It often differs from study to study (*e.g.* Ardizone, 1989 versus Álvarez García, 2002). It is are almost impossible to ascribe such anomalies to an unquestionable Variscan age and, in some cases, they are clearly related, either to Alpine events (like in the vicinity of Béticas Chain), or inherited Cadomian basement structures.

- The new pattern of the D_1 and D_3 Variscan folding events (Fig. 6B; *e.g.* Martínez Catalán, 2011a) are in clearly contradicts most of the published structural studies and the new interpretation is not supported by new structural studies.

- The supposed centripetal trend of the Lower Ordovician paleocurrents around the orocline (*e.g.* Shaw *et al.*, 2012) results from a very crude interpretation of the data set (Fig. 10).

- The proposed correlation between the WALZ and the southern domains of the CIZ (Shaw *et al.*, 2012) is not supported by lithostratigraphic data (see compilations concerning WALZ in Pérez-Estaún *et al.*, 1990; Marcos *et al.*, 2004 and for the CIZ in Gutiérrez Marco *et al.*, 1990; San José*et al.*, 1990; Díez Montes *et al.*, 2004; Rodríguez Alonso *et al.*, 2004; Romão *et al.*, 2013 and Fig. 1). Thus, this correlation implies (Shaw *et al.*, 2012) along strike variation (not observed because it is hidden below the Cenozoic Douro and Tejo basins) in order to accomplish the differences between the supposed northern limb of the Arc (in the WALZ) and the southern limb (in the southern domains of the CIZ). Moreover, the lithostratigraphy of the CIZ southern domains (*e.g.* Romão *et al.*, 2013) is very similar to the one found in the northern ones (*e.g.* Dias *et al.*, 2013b), mainly concerning the post-Cambrian metasediments, which suggests a common basin.

- Recent paleomagnetic data (Pastor-Galán *et al.*, 2015) discard a major bend at the CIA.

Thus, the geological arguments supporting the CIA are highly questionable and this proposal should be abandoned.

IX.2.4.1.3. Ibero-Armorican Arc

The continuity between the northern Armorican and the southern Iberian branches of the IAA cannot be followed due to the Atlantic and Biscay Mesozoic oceanic rifting. Nevertheless, there is a strong similarity between the geological formations found in both sectors, not only of the autochthon but also of the allochthonous units (Ballèvre *et al.*, 2013). This led to a strong consensus concerning the IAA.

IX.2.4.2. Previews Models

The complexity of the Variscan arcs in Iberia, gave rise to a wide range of models. A major constraint results because some models explain such arcs using a plate tectonic framework with only one ocean (the Rheic), while others also considered a second minor ocean related to the southern Brittany suture (the Paleotethys of Stampfli and Borel, 2002, also called Galicia-Massif Central Ocean by Matte, 2002; Ballèvre *et al.*, 2009). A brief critical review of these models helps to narrow the range of possible mechanisms. In this approach, previous models have been grouped using a simplification of Macedo and Marshak classification (1999).

IX.2.4.2.1. Arcs due to Margin Irregularities

Even before the establishment of plate tectonics, Dana (1886) proposed that the arcuate shape of some mountain belts was due to their wrapping around nonlinear margins of a pre-existing craton. Since then, several works recognize that the heterogeneous compressive stress induced by the collision with an irregular continental margin should produce wavy fold-thrust belts (Marshak, 1988). Such arcuations could be either margin-controlled curves or indenter-controlled curves (Macedo and Marshak, 1999). In the first situation, the thrusting is related to an originally nonlinear continental margin, while in the second case the arc forms because at the lateral edge of the indentor the deforming rock layer is subject to a shear couple, whereas at the front of the indentor the deforming rock layer is subject to a normal stress in plan (Marshak, 1988). Both solutions have been used since the early attempts to explain the IAA formation (Fig. 11).

Using mostly paleomagnetic, paleontological, sedimentary and magmatic data Lorenz (1976) and Lorenz and Nichols (1984) consider a paleogeographic realm for Lower Carboniferous times (Fig. 11A_1) with a small and very elongated Southern Europe continental plate between two large ones: the northern North America - Europe and the southern Africa. The minor plate was limited by two oceans: the northern Mid - European (Buret, 1972) or Rheic Ocean (Mckerrow and Ziegler, 1972) and the southern Paleotethys (Stampfli, 1996). The metamorphic and magmatic processes induced by the two centripetal subduction in the overriding Southern Europe Plate (Fig. 11A_2) have weakened this continental plate due to widespread partial melting in the lower and middle crust (Lorenz and Nichols, 1984). Thus, when in the Visean to Westephalian times, Southern Europe collided with both large neighbouring North America-Europe and Africa plates the irregular continental margins of the larger and thicker major continental plates induced oroclinal bending of Variscan Europe (Fig. 11A_3). Such process gave rise to the IAA induced by the northern Brabant-Newfoundland embayment, due to clockwise rotation of the Armorican sub-plate (western France) and anticlockwise rotation of Iberian sub-plate, explaining the sinistral shears in Iberia and the dextral ones in Brittany. However, a major objection to such model, emphasizes the diachronism behaviour of the shear zones in both limbs of the IAA (Audren *et al.*, 1976), which are Devonian in Iberia and Carboniferous in Brittany (Dias and Ribeiro, 1995b). Another main problem concerns the proposed location of Iberia in Southern Europe minor plate (Fig. 11A$_3$), when most recent reconstructions show its close affinity to the major Gondwana (*e.g.* Ribeiro *et al.*, 2007).

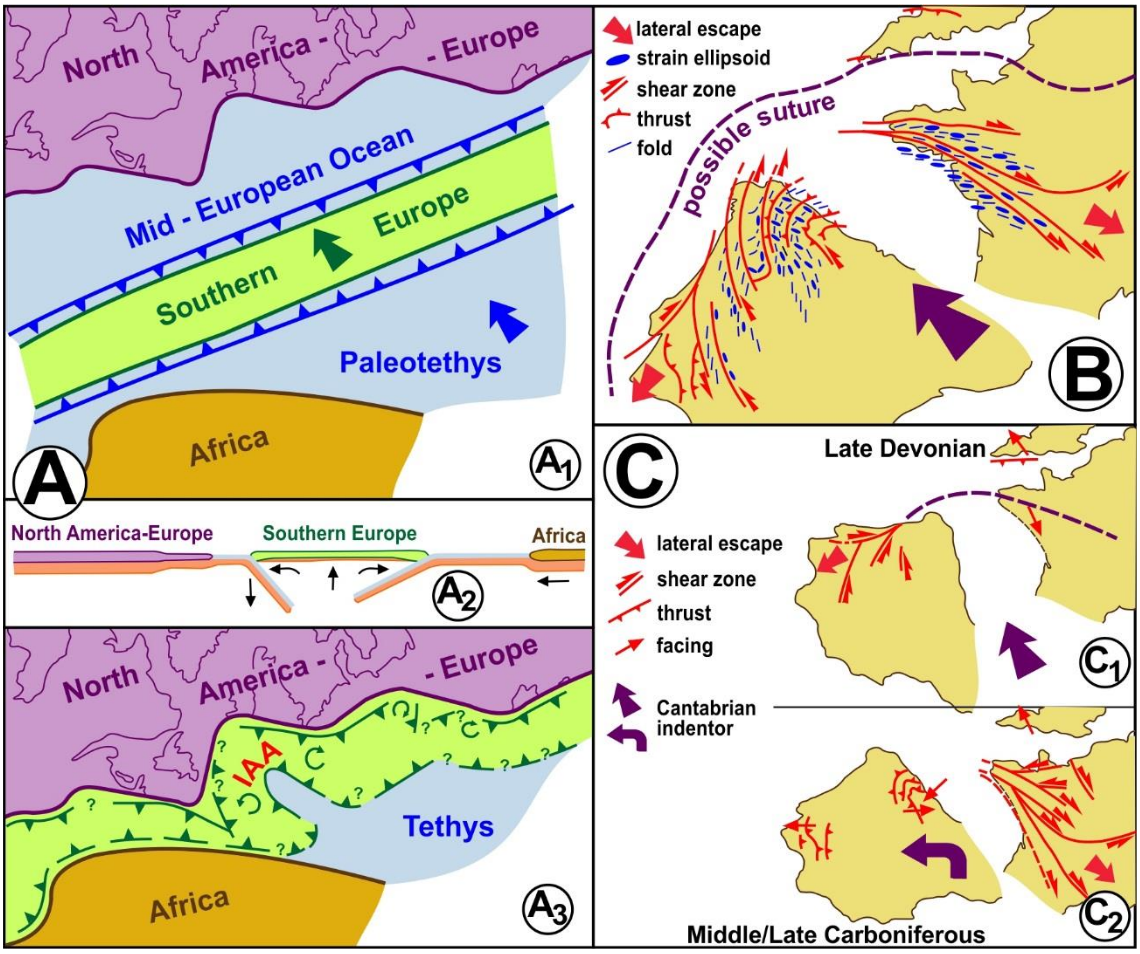

Figure 11 – Proposed models for the IAA formation in relation to major irregularities of continental margins:

A – Internal plastic distortion of the Southern Europe plate by the adjustment to irregular continental margins of large neighbouring plates (adapted from Lorenz and Nicholls, 1984);

B – Indentation related to one tectonic event (adapted from Matte and Ribeiro, 1982; Matte, 1991);

C – Indentation during two tectonic events (adapted from Dias and Ribeiro, 1995a).

An alternative approach considers the IAA the result of a wrapped Southern Europe microplate around a rigid indentor of northern margin of a major Africa plate. Using the opposed kinematics of the NW-SE sinistral shear zones in Iberia and the E-W dextral ones in Brittany, coupled with the Variscan finite strain pattern (orthogonal to the main structures in the core of the arc, and longitudinal in the more external domains) the IAA was explained (Matte and Ribeiro, 1975) as the result of an indentation process giving rise to a thin skinned arc with centripetal vergences (Fig. 11B). Although in this model the arc is essentially the result of a tectonic process, it was suggested that the deformed Palaeozoic basins and their boundary basement faults could have some initial curvature. This early indentation model also did not take into account the diachronism between shearing events in northern and southern branches of the arc (Audren *et al.*, 1976). Nevertheless, similar models were often used in later works where the arcuation was due, either to the impingement of a promontory of the African continent (Matte and Burg, 1981; Matte, 1991) or to the more rigid behaviour of the West African Precambrian craton (Lefort, 1989; Lefort and Van der Voo, 1981).

In order to account for the age disparity between the Iberia and Armorican shears, a two stage indentation model was presented (Dias and Ribeiro, 1995b; Ribeiro *et al.*, 1995). During Upper Devonian (Fig. $11C_1$) the northward displacement of the Gondwana indentor produced NW-SE sinistral transpression in Iberia and almost orthogonal collision farther north, either towards NNW (Lizard obduction in southwest Britain) or SE (Bretonic phase in the Armorican Massif). The collision of Iberia with the irregular southern margin of Laurentia/Baltica induced an anticlockwise rotation of Iberia during Late Carboniferous tightening the arc. This led to a change in the deformation regime in the IAA; in the northern branch dextral strike-slip shear zones were predominant, while in Iberia thrusts overprinted previous structures (Fig. $11C_2$). This model not only explains the differences in the structural behaviour between both branches during Devonian / Carboniferous times, but also the younger origin of the CA. Moreover, it is still consistent with the recent data concerning the southern Pulo do Lobo domain (located in the boundary between Ossa-Morena and South Portuguese Zones), which indicates it was a Southern Uplands terrane displaced along a major sinistral shear zone in Early to Middle Devonian times due to the indentation of the Iberian promontory in the British Caledonides (Braid *et al.*, 2011).

It has recently been it was proposed that the Cantabrian orocline was due to the westward drift of the Pyrenees between a major dextral E-W northern fault and the left-lateral shear system of inner Iberia during Late Carboniferous (Sengör, 2013). As in the early indentation models (Matte and Ribeiro, 1975) this also presents the major problem of diachronism between the NW-SE sinistral shears and the dextral E-W ones.

IX.2.4.2.2. Arcs controlled by Major Strike-Slip Shear Zones

Several models consider the formation of Iberian Variscan arcs mostly the result of the movement along major strike-slip shear zones. Slip on wrench faults could induce pronounced curvature of fold-thrust belt (Marshak, 1988; Macedo and Marshak, 1999), as the Cenozoic South Island orocline in New Zealand (Sutherland, 1999; Hall *et al.*, 2004). Such process could result, either from a younger fault overprinting a previous belt, or from the fact that the movement on the fault and the growing belt are contemporaneous (Marshak, 2004).

One of the first proposals (Ries and Shackleton, 1976) considered the IAA a secondary arc due to a counter-clockwise rotation of Iberia relative to Brittany, with the amount of rotation around the arc increasing southwards in direction to a supposed transform fault (Fig. 12A). Such simple mechanism, in spite of explaining the arcuate shape and the finite strain contrast between the inner and the external domains of the arc (Fig. 5A), is unable to explain the main regional sinistral shear of northern CIZ (Fig. 5B). Moreover, it only considered one tectonic event, which is incapable of explaining the variability of deformation observed in both branches of the IAA.

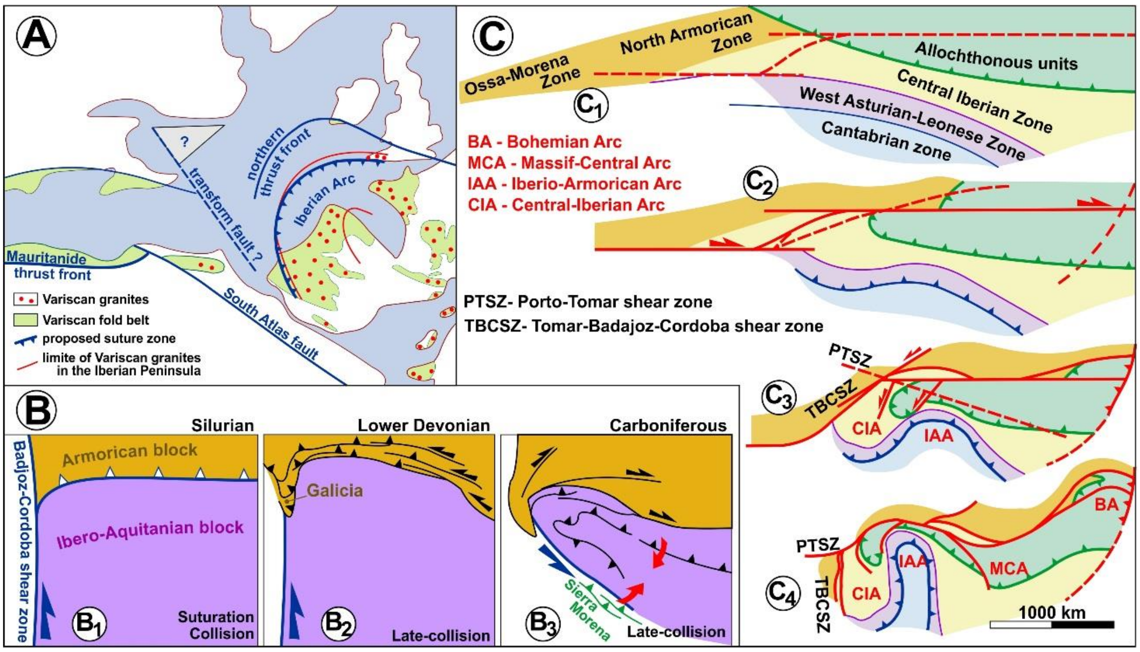

Figure 12 – Proposed models for the IAA formation in relation to major strike-slip shear zones:

A – Iberian rotation in relation with a major E-W transform fault (adapted from Ries and Shackleton, 1976);

B – The predominance of the Badajoz-Cordoba shear zone (adapted from Brun and Burg, 1982; Burg *et al.*, 1987);

C – The major role of E-W major shears (adapted from Martínez Catalán, 2011c).

Later models mostly explore the role of major TBCSZ and PTFASZ Iberian shear zones (Fig. 1). One of the early approaches (Brun and Burg, 1982; Burg *et al.*, 1987) considers a long-lasting interaction between the TBCSZ and an E-W southwards dipping intra-oceanic subduction zone, related to the emplacement of the Limousin ophiolites. During oceanic subduction the TBCSZ behaves has a sinistral transform fault (Fig. $12B_1$). With the collision of the two continents In Lower Devonian, the deformation becomes intra-continental increasing the interaction transform fault / continental subduction. As the subduction of the continental crust is limited, a so-called "corner effect" was produced at the intersection of the transcurrent shear zone and the thrust zone (Fig. $12B_2$). During Carboniferous collision, the intra-continental deformation increases the previous incipient curvature (Fig. $12B_3$). This tightening of the arc is coeval of the foreland thrusting on its core and of the important dextral and sinistral shearing on the outer Armorican and Iberian domains. This model has the main advantage of considering two main stages with different kinematics for the Devonian and Carboniferous deformation. Nevertheless, the correlation between the different tectonostratigraphic zones of the southern and northern limbs of the arcs is difficult to explain.

A different proposal relates the formation of the IAA with a 4000-5000 km dextral displacement of Laurentia around northern Gondwana during closure of the Rheic and the formation of Pangaea (Shelley and Bossière, 2000; 2002). The arc was formed, either by wrapping the mobile dextral transpressive shear belt around a rigid Iberian basement block (Shelley and Bossière, 2000), or counter clockwise rotation of the south-western part of Iberia around a possible extensional bend of the major dextral shears (Shelley and Bossière, 2002). Whatever the options, they do not exclude some indentation of an Iberian promontory synchronous of the major dextral transpressive shearing. Such idea is questionable, because oblique indentation originates asymmetric arcs (Marshak, 2004), which is not the case. Even if one does not question the existence of the proposed huge shear for the Variscan Fold Belt origin (a highly debatable topic; *e.g.* Stampfli and Borel, 2002), some assumptions are difficult to accept. One is the proposal that the sinistral faults of Iberia (*e.g.* the TBCSZ) are bookshelf-type related to the dextral shearing along the PTFASZ, considered the most important shear zone of Iberian Variscides, because it does not explain the pervasive D_1 sinistral kinematics in northern sectors of the CIZ. As they assume some part of the arcuation is primary, the degree of bending of the arc should be much less than the estimated using recent paleomagnetic data (section 3.4; Weil *et al.*, 2013).

A recent model (Martínez Catalán, 2011c; Martínez Catalán *et al.*, 2014) also relates the formation of the Iberian arcs with major E-W dextral shearing but only during Middle to Upper Carboniferous (Fig. 12C). This approach does not only explain the formation of the arcuate geometry and the correlation between the different variscan domains, but also the paleomagnetic behaviour established in the CA (Weil *et al.*, 2013). The deformation starts with the emplacement of the allochthonous nappes oblique to the zones in the autochthon (Fig. $12C_1$) giving rise to the thickening of the northern Gondwana continental margin. The tectonic setting changes with the onset of two major E-W right-lateral strike-slip faults with a stair-case geometry (Fig. $12C_2$). The evolution of the previous pattern gave rise to the bending of the wrench system in the sector where the major shears overlap, creating a restraining bend forming the CIA as a fault-bend or a fault-propagation fold. Subsequent movement along dextral major shears induced domino-style antithetic structures (Fig. $12C_3$), explaining the last motion along the TBCSZ. Such event was superposed on the pre-existing CIA and is older than the PTFASZ. Finally, the tightening of the IAA (Fig. $12C_4$), occurred due to the counter clockwise ductile rotation of the bounding shear zones to the west (TBCSZ) and to the east (Moldanubian thrusts), perhaps because their orientation was no longer able to allow the gliding of the blocks. The formation of the Bohemian and Massif Central arcs was produced during this last intra-continental deformation. The major problems with the previous model is, not only the existence of the highly dubious CIA, but also the assumption that all the Variscan deformation in Iberia has a Middle to Upper Carboniferous age.

IX.2.4.2.3. Arcs Related to Lithospheric Delamination

A different approach was recently proposed mostly based on the location of the CA in the core of Pangaea supercontinent precisely, on the western tip of Paleotethys (Fig. $13A_1$). If the first proposals only considered the CA (Gutiérrez-Alonso *et al.*, 2004; 2008; Fig. $13A_2$), soon these models evolved to include also the CIA with a "S" general pattern for the paired oroclines (Gutiérrez-Alonso *et al.*, 2011; 2012; Johnston *et al.*, 2013; Weil *et al.*, 2013; Fig. $13A_3$).

In spite of some variations, these models have always considered two tectonic events (Fig. 13B). During Carboniferous the closure of Rheic Ocean between Gondwana and Laurussia, due to an E-W shortening (in present-day coordinates), produces a N-S linear orogen facing East (Fig. $13B_1$). Close to the Permian-Carboniferous boundary (from 315 to 299 Ma) the shortening becomes N-S inducing orocline buckling around a vertical axis (Fig. $13B_2$). From the early works concerning the IAA (Matte and Ribeiro, 1975; Ries and Shackleton, 1976), the resultant strain field was considered to be close to tangential longitudinal strain leading to strong space problems in the inner arc (Fig. 13C). The extensive magmatism coeval of the CA formation was considered to reflect a thick-skinned process with the involvement of all the lithosphere (Gutiérrez-Alonso *et al.*, 2004; 2008; 2011; Fig. $13D_1$). The outer arc extension and the inner arc compression gave rise to a strong lithospheric thinning in the outer arc coupled with a strong thickening in the CA core. The continuity of the process induced delamination of the mantle lithosphere under the core of the CA, solving some space problems (Fig. 13D_2).

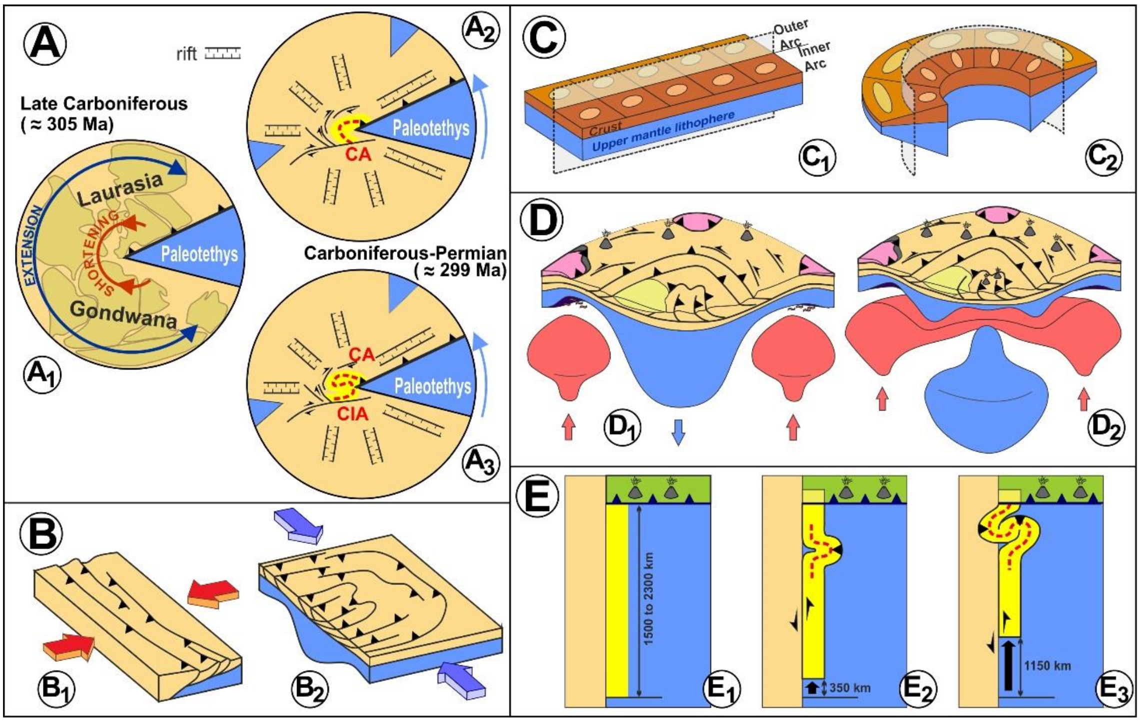

Figure 13 – Proposed models for the IAA formation by lithospheric delamination:

A – Schematic diagrams showing simplified Pangaea reconstructions (adapted from Gutiérrez-Alonso *et al.*, 2012; Johnston *et al.*, 2013);

B – Stress rotation inducing secondary Cantabrian orocline formation by buckling (adapted from Johnston *et al.*, 2013; Weil *et al.*, 2013);

C – Neutral surface model (adapted from Gutiérrez-Alonso *et al.*, 2012; Weil *et al.*, 2013);

D – Schematic block diagram illustrating Cantabrian orocline development (adapted from Gutiérrez-Alonso *et al.*, 2012; Weil *et al.*, 2013);

E – Geometric tectonic model for secondary development of coupled orocline (adapted from Johnston *et al.*, 2013).

Despite always coupling the CA and CIA, these models never discuss how to solve the space problems in the southern CIA. Another major problem concerns the mechanism that has induced a 90º rotation of the major compression from normal to subparallel to the orogenic belt. This is not an easy task and several mechanisms have been proposed:

- The so-called self subduction of Pangaea (Gutiérrez-Alonso *et al.*, 2004; 2008 Fig. 13A_2) where the Upper Carboniferous compressive stress field inside the plate was induced by the oceanic margin of a continent being subducted beneath the continental edge at the other end of the same plate.

- The buckling of a ribbon continent between Laurussia and Gondwana during the final amalgamation of Pangaea (Johnston and Gutierrez-Alonso, 2010; Weil *et al*., 2010; Fig. 13E).

If in the first case a possible explanation for the strong reorientation of the stress field was advanced, in the second, the rotation of the very long (at least 1500 km) Cantabrian - Central Iberian ribbon continent from the Early Palaeozoic E-W trend to its N-S pre-orocline position during the Late Carboniferous is unclear (Shaw *et al.*, 2014).

However the rotation mechanism isn't the only problematic situation (Sengör, 2013):

- There are frequent Late Carboniferous - Permian E-W dykes, showing that σ_1 (maximum compression) is oriented E-W and not N-S, as predicted by the models.
- Lithostratigraphic arguments show that the core of the CA was never very high as it should be expected in a region above a delaminated lithospheric mantle.
- The Early Permian volcanics which have been considered in NW Iberia induced by the lithospheric detachment are not exclusive of this region, being characteristic of the entire Late Variscan magmatism of Europe.
- It is difficult to explain the preservation of the highest supracrustal sedimentary rocks in the core of the Cantabrian orocline, if a lower lithospheric detachment has occurred.

IX.2.5. A Unifying Approach

Any discussion of Iberian arcs models must take into account some major constraints:

- There is no evidence concerning the existence of the CIA and so, only the IAA and the CA should be considered.
- The age of the first and main Variscan tectonic event is at least middle Devonian in the inner domains of the Iberian Variscides and it propagates towards its external domains where it has an Upper Carboniferous age (Fig. 5).
- Although the data show that the CA was formed during Carboniferous / Permian (*e.g.* Weil *et al.*, 2013), this does not mean that the IAA has a coeval formation, representing an earlier stage of arc generation.
- The E-W trend of the axial planes of IAA and CA was used to deduce a N-S major compressive stress, inducing orocline buckling around a vertical axis (Fig. $13B_2$). However, the same geometry could be obtained by bending (Hobbs *et al.*, 1976; Twiss and Moores, 1992). So, a westwards indentation is also compatible with the CA data, and it is going to be used here because it is more able to explain the observed features.

- Although recent models concerning the Variscan evolution mostly focus on Rheic Ocean without considering an Armorica plate, the existence of a Paleotethys / Galicia - Massif Central Ocean is well established in Brittany and is going to be used (Fig. 14).

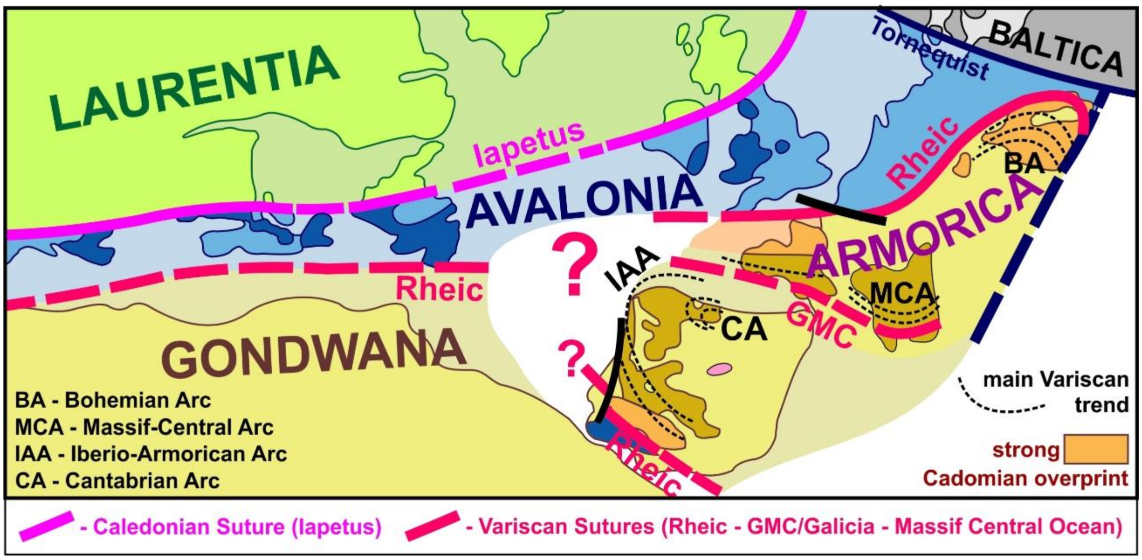

Figure 14 – Major plates and Palaeozoic sutures in the vicinity of the IAA on an Early Mesozoic reconstruction of the North Atlantic (adapted from Matte, 2001; Ribeiro *et al.*, 2007; Ballèvre *et al.*, 2009; Nance *et al.*, 2012).

The previous assumptions have been integrated in a general model. In Upper Ediacarian, oblique subduction below the northern margin of Gondwana gives rise to the Cadomian arc-continent collision (Fig. 15A). The sectors closer to the trench were strongly deformed, being included in the so-called Cadomia microplate (Linnemann *et al.*, 2008; Nance *et al.*, 2012). During Lower Palaeozoic stretching predominates in this continental margin, inducing a widespread thinning. In Lower / Middle Cambrian, Avalonia began to drift (Linnemann *et al.*, 2008) due to the opening of Rheic Ocean (Fig. 15B), mostly following an old Neoproterozoic suture (Murphy *et. al.*, 2006). Farther north, the subduction of Iapetus below Laurentia compensate the widening of Rheic and the continuous stretching in northern margin of Gondwana (Fig. 15C), giving rise to the Iberia and Armorica paleogeographic domains (Fig. 1). The position of these domains is difficult to establish due to the intense Variscan deformation of Peri-Gondwana Terranes during Pangaea assemblage.

Another ribbon continent, Armorica, began to be isolated from Gondwana due to the opening of Galicia-Massif Central Ocean (Fig. 15D; Matte, 2001; Ballèvre, 2013). Such opening could be related to the subduction of the Rheic below the northern margin of Armorica (Ribeiro *et al.,* 2007). Due to the obliquity between previous paleogeographic zones and the major trend of this new ocean, some of the zones appear in both margins, while others are restricted to the Iberian side (*i.e.* CZ and WALZ).

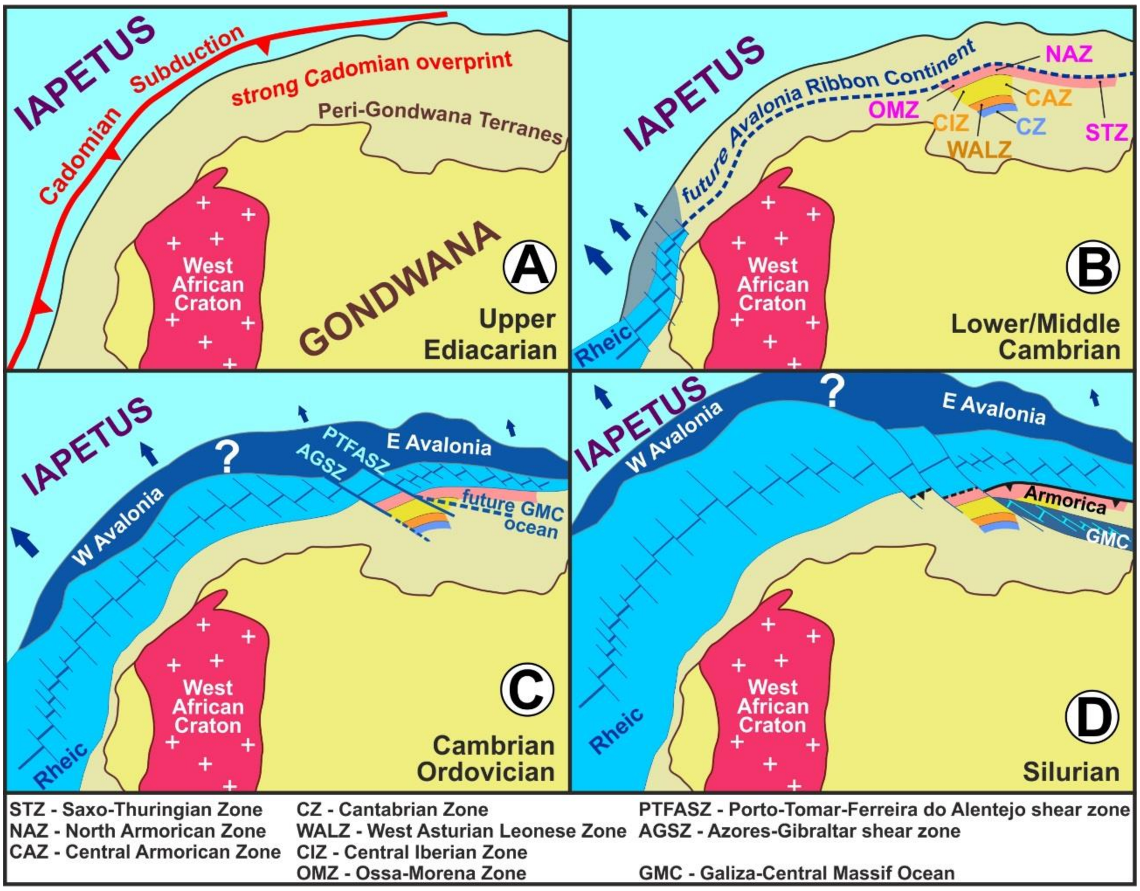

Figure 15 – Schematic Lower Palaeozoic evolution of northern margin of Gondwana (adapted from Matte, 2001; Linnemann, *et al.,* 2008; Nance *et al.*, 2012).

A – The Cadomian Arc during Upper Ediacarian;

B – Individualization of Avalonia by the opening of Rheic Ocean;

C – The northward drift of Avalonia due to widening of Rheic;

D – The drift of Armorica due to the Galicia-Massif Central Ocean.

During Devonian / Carboniferous the convergence of the continents to form Pangaea led to a strong inversion of initial basins. In the Devonian, the subduction of Variscan oceans produce (Fig. 16A) an almost orthogonal thrust tectonics in the Brittany branch (Burg *et al.*, 1987; Dias and Ribeiro, 1995b), with sense of movement either towards the NNW (Lizard obduction in southwest Britain) or SE (Bretonic phase in the Armorican Massif). At the same time in Iberia due to the interaction between the subduction of the Rheic and the wavy northern Gondwana margin an oblique collision occurs, inducing the predominance of a sinistral transpressive regime. Due to the location of the Iberian suture in the external boundary of OMZ, the deformation propagates from there towards the CZ (Fig. 5). However, due to a strong strain partitioning, the Devonian deformation in Iberia was very heterogeneous, with regions where sinistral transpression was predominant, adjacent to regions deformed by almost pure compression (Dias and Ribeiro, 1994; 1995b; Dias *et al.*, 2013b). This explains the coexistence of regions dominated by almost a pure thrusting tectonics, like happens in ZOM (e.g. Simancas *et al.*, 2003 and references therein).

In Early Carboniferous the tectonic setting is similar, but previous plate convergence has evolved to continental collision (Fig. 16B). The allochthonous units were then emplaced with the movement above the CIZ subparallel to the main D_1 structures, which were often reworked as lateral ramps (Rodrigues *et al.* 2005; 2013; Dias *et al.*, 2013b).

In the Upper Carboniferous / Lower Permian a drastic change occurs (Weil *et al.*, 2013). The irregular shape of the southern margin of Laurentia/Baltica block has probably induced an anticlockwise rotation of Iberia during the intra-continental deformation (Dias and Ribeiro, 1995b). Thus, in the northern branch of the IAA major dextral strike-slip shear zones occur, while in Iberia thrusts overprinted previous structures. Such process is not unusual and has already been proposed not only for the Variscan orogen (Lorenz and Nicholls, 1984), but also for the Alpine Himalayan Fold Belt (Treloar *et al.*, 1992) and the Oligocene collision between Africa and Euro-Asian plates (Carvalho *et al.*, 1983-85). The Cantabrian basement was displaced westwards producing indentation on the WALZ and CZ metasediments above the indenter, giving rise to a tight first order arcuate shape, the CA (Fig. 16C). A thin-skinned arc was produced with the major thrusts displaced towards E above a decollement located within the limestones and dolomites of the Lancara Formation of Lower-Middle Cambrian age (Pérez-Estaún and Bastida, 1990; Alonzo *et al.*, 2009). This indentation also affects the NW Iberian units, not only the allochthonous nappes but also the autochthon ones. The more open shape of this western thick-skinned IAA, reflects not only the higher metamorphic grade of the rocks of the inner domains of the Iberian Variscides, but also the greater distance to the indentor. Moreover, it should be stressed that the arcuate shape of this arc, between the NW-SE structures in Iberia, and the E-W ones in Brittany, did not represent a strong rotation of a previous liner belt, but two major trends of mostly independent structures that have been at a high angle in the early stages, and were slightly bent at a latter phase.

We are well aware that this model didn't take into account the interpretation of the axial zone of Tomar-Badajoz-Córdoba as an Eo-Variscan suture (Simancas *et al.*, 2001; Gomez Pugnaire *et al.*, 2003). However, we disagree with such interpretation, because we consider this structure is a Cadomian suture reactivated as a transpressive intraplate flower structure, for reasons exposed in detail elsewhere (Ribeiro *et al.*, 2007) and confirmed by recent geochronological data (Henriques *et al.*, 2015).

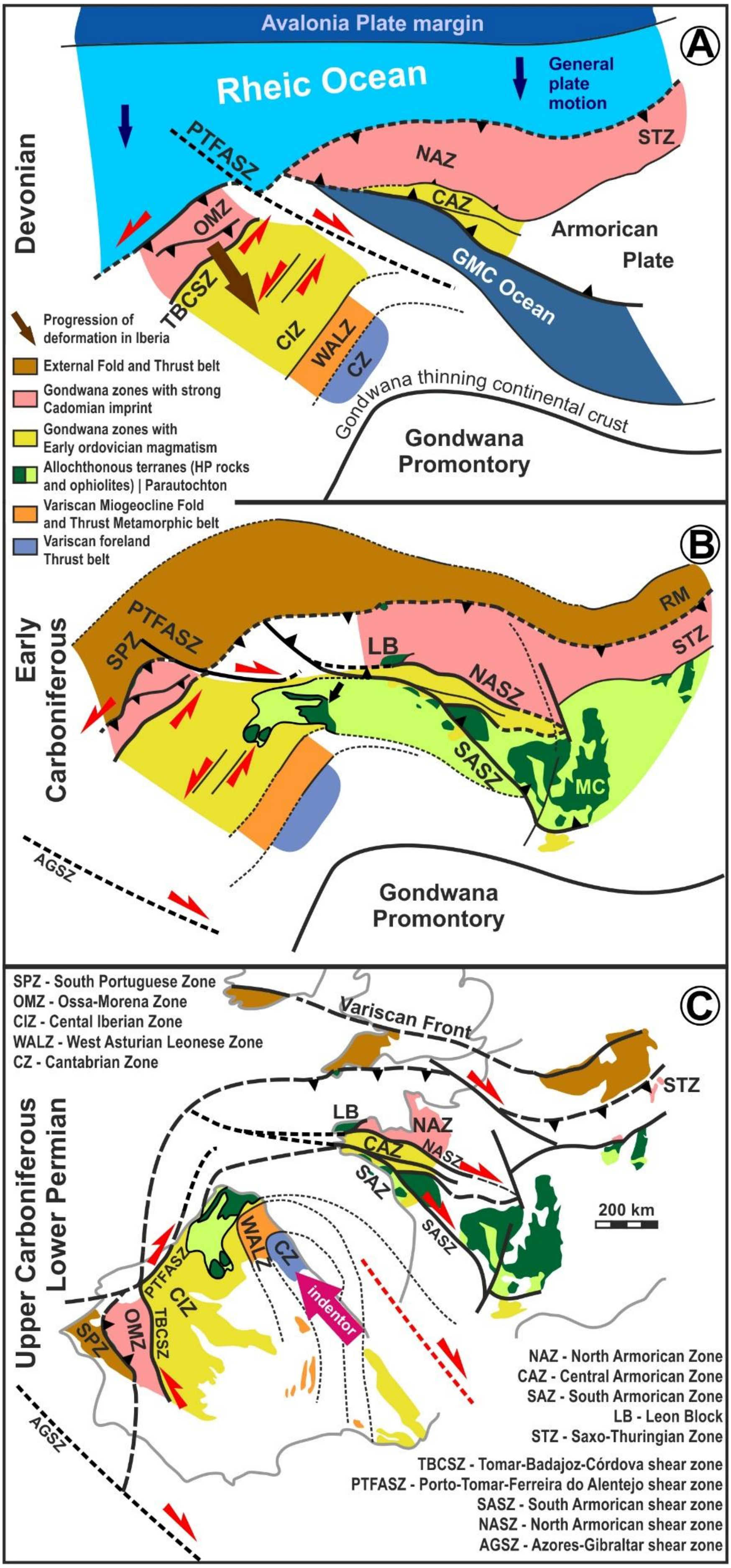
Avalonia Plate margin
Rheic Ocean
General plate motion
A
Devonian
PTFASZ
NAZ
STZ
CAZ
OMZ
Armorican Plate
GMC Ocean
TBCSZ
CIZ
WALZ
CZ
Progression of deformation in Iberia
External Fold and Thrust belt
Gondwana zones with strong Cadomian imprint
Gondwana zones with Early ordovician magmatism
Allochthonous terranes (HP rocks and ophiolites) | Parautochton
Variscan Miogeocline Fold and Thrust Metamorphic belt
Variscan foreland Thrust belt
Gondwana thinning continental crust
Gondwana Promontory
B
Early Carboniferous
PTFASZ
SPZ
LB
NASZ
RM
STZ
SASZ
MC
AGSZ
Gondwana Promontory
C
SPZ - South Portuguese Zone
OMZ - Ossa-Morena Zone
CIZ - Cental Iberian Zone
WALZ - West Asturian Leonese Zone
CZ - Cantabrian Zone
Variscan Front
STZ
LB
NAZ
CAZ
NASZ
SAZ
SASZ
200 km
Upper Carboniferous Lower Permian
PTFASZ
WALZ
CZ
indentor
CIZ
TBCSZ
OMZ
SPZ
AGSZ
NAZ - North Armorican Zone
CAZ - Central Armorican Zone
SAZ - South Armorican Zone
LB - Leon Block
STZ - Saxo-Thuringian Zone
TBCSZ - Tomar-Badajoz-Córdova shear zone
PTFASZ - Porto-Tomar-Ferreira do Alentejo shear zone
SASZ - South Armorican shear zone
NASZ - North Armorican shear zone
AGSZ - Azores-Gibraltar shear zone

Figure 16 – Schematic interaction between Avalonia / Armorica and northern Gondwana during Upper Palaeozoic (colours and abbreviations like in figure 1).

A –Convergence between Iberia and Armorica during Middle Devonian;

B –The collision stage with all the plates assembled;

C –The westward indentation of the Cantabrian basement and the formation of major Iberian Arcs.

References

Ábalos, B. (1992). Variscan shear-zone deformation of late Precambrian basement in SW Iberia: implications for circum-Atlantic pre-Mesozoic tectonics. J. Struct. Geol., 14(7), 807-823.

Ábalos, B., Carreras, J., Druguet, E., Viruete, J., Pugnaire, M., Alvarez, S., Quesada, C., Fernández, L., Gil-Ibarguchi J. (2002). Variscan and Pre-Variscan Tectonics. In: Gibbons, W., Moreno, M.T. (Eds.), The Geology of Spain. Geol. Soc. London, 155-183.

Aerden, D. (2004). Correlating deformation in Variscan NW-Iberia using porphyroblasts; implications for the Ibero-Armorican Arc. J. Struct. Geol., 26, 177-196.

Álvarez García, J. (2002). Análisis gravimétrico e isostético en el Macizo Hespérico. Diploma de Estudios Avanzados, Univ. Complutense de Madrid. 72 p.

Aramburu, C., Garcia-Ramos, J. (1993). La sedimentación Cambro-Ordovícica en la Zona Cantábrica (NO de España). Trab. Geol., 45-73.

Araújo, A., Piçarra de Almeida, J., Borrego, J., Pedro, J., Oliveira, J. T. (2013). As regiões central e sul da Zona de Ossa-Morena. In: Dias, R., Araújo, A., Terrinha, P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1). Escolar Editora, 509-549.

Arboleya, M. (1981). La estructura del manto del Esla (Cordillera Cantábrica, León). Bol. Inst. Geol. Min. Esp., XCII/I, 19-40.

Ardizone, J., Mezcua, J., Socías, I. (1989). Mapa Aeromagnético de España Peninsular 1:1.000.000. Instituto Geográfico Nacional, Madrid.

Arenas, R., Farias, P., Gallastegui, G., Gil Ibarguchi, J., González Lodeiro, F., Klein, E., Marquinez, J., Martin Parra, L., Martinez Catalán, J., Ortega, E., Pablo Macia, J., Peinado, M., Rodriguez Fernández, L. (1988). Características geológicas y significado de los dominios que componen la Zona de Galicia-Trás-os-Montes. Simposio sobre Cinturones Orogénicos, II Congresso de Geologia de España, Granada, 75-84.

Arenas, R., Díez Fernández, R., Sánchez Martínez, S., Gerdes, A., Fernández-Suárez, J., Albert, R. (2014). Two-stage collision: Exploring the birth of Pangea in the Variscan terranes. Gondwana Research, 25, 756–763.

Argand, E, 1924. La Tectonique de l'Asie. In: Proc. XIII Int. Geol. Congr., Brussels, 171-173.

Audren, C., Brun, J. Cobbold, P., Cogné, J., Iglésias, M., Jegouzo, P., Le Correm C., Le Metour, J., Le Théoff, N., Rabu, D. (1976). Données complémentaires sur la géométrie du plissement et sur les variations de forme et orientation de l'ellipsoïde de déformation dans l'arc Hercynien Ibéro-Armorican, Bol. Soc. Géol. Fr., XVIII, 757-762.

Azor, A., González Lodeiro, F., Simancas, J. (1994). Tectonic evolution of the boundary between the Central Iberian and Ossa-Morena zones (Variscan belt, southwest Spain), Tectonics, 13, 45–61.

Azor, A., Rubatto, D., Simancas, J.F., González Lodeiro, F., Martínez Poyatos, D., Martín Parra, L.M., Matas, J. (2008). Rheic Ocean ophiolitic remnants in southern Iberia questioned by SHRIMP U-Pb zircon ages on the Beja-Acebuches amphibolites. Tectonics 27, TC5006.

Ballèvre, M., Bosse, V., Ducassou, C., Pitra, P. (2009). Palaeozoic history of the Armorican Massif: Models for the tectonic evolution of the suture zones. C. R. Geoscience, 341, 174–201.

Ballèvre, M., Bosse, V., Dabard, M., Ducassou, C., Fourcade, S. (2013). Histoire Géologique du massif Armoricain : Actualité de la recherche. Bull. Soc. Géol. Min. Bretagne, D/10-11, 5-96.

Ballèvre, M., Martínez Catalán, J., López-Carmona, A., Pitra, P., Abati, J., Díez Fernández, R., Ducassou, C., Arenas, R., Bosse, V., Castiñeiras, P., Fernández-Suárez, J., Gómez Barreiro, J., Paquette, J., Peucat, J., Poujol, M., Ruffet, G., Sánchez Martínez, S. (2014). Correlation of the nappe stack in the Ibero-Armorican arc across the Bay of Biscay: a joint French–Spanish project. In: Schulmann, K., Martínez Catalán, J., Lardeaux, J., Janousek, V., Oggiano, G. (Eds.), The Variscan Orogeny: Extent, Timescale and the Formation of the European Crust. Geological Society, London, Special Publications, 405, 77–113.

Bard, J., Matte, P., Capdevila, R. (1971). Le structure de la Chaîne Hercynienne de la Meseta Ibérique: comparaison avec les segments voisins. Publ. Inst. Fr. Pétrole, Collections Coll. et Sémin., 22 (Histoire Structurale du Golfe de Gascogne, 1, 1-68.

Barrois, C. (1882). Recherches sur les Terrains Anciens des Asturies et de la Galice, Mémoires de la Société Géologique du Nord, 2, 630 p.

Bastida, F., Martínez Catalán, J., Pulgar, J. (1986). Structural, metamorphic and magmatic history of the Mondoñedo nappe (Hercynian belt, NW Spain). J. Struct. Geol., 8, 415–430.

Bastida, F., Aller, J., Pulgar, J., Toimil, N., Fernández, F., Bobillo-Ares, N., Menéndez, C. (2010). Folding in orogens: a case study in the northern Iberian Variscan Belt. Geol. J., 45, 597–622.

Bertrand, M. (1887). La Chaîne des Alpes et la formation du continent européen. Bull. Soc. Géol. France, 15(3), 423-447.

Bonhommet, N., Cobbold, P., Perroud, H., Richardson, A. (1981). Paleomagnetism and cross-folding in a key area of the Asturian Arc (Spain). Journal of Geophysical Research, 86, 1873–1887.

Braid, J., Murphy, J., Quesada, C., Mortensen, J. (2011). Tectonic escape of a crustal fragment during the closure of the Rheic: U-Pb detrital zircon data from the Late Palaeozoic Pulo do Lobo and South Portuguese zonas, southern Iberia. J. Geol. Soc. London, 168, 383-392.

Brun, J., Burg, J. (1982). Combined thrusting and wrenching in the Ibero-armorican arc — a corner effect during continental collision. Earth and Planetary Science Letters, 61, 319–332.

.Burg, J., Iglesias, M., Laurent, P., Matte. P., Ribeiro, A. (1981). Variscan intracontinental deformation: the Coimbra-Córdoba shear zone (SW Iberian Peninsula). In: Lister, G. S., Behr, H. J., Weber, K., Zwart, H. J. (Eds.), The effect of deformation on rocks. Tectonophysics, 78, 161-177.

Burg, J., Bale, P., Brun, J., Girardeau, J. (1987). Streching lineations and transport direction in the Ibero-Armorican Arc during the Siluro-Devonian collision. Geodinamica Acta, 1, 71-81.

Carey, S. (1955). The orocline concept in geotectonics. Proceedings of the Royal Society of Tasmania, 89, 255–288.

Carey, S. (1958). The tectonic approach to continental drift. In: S.W. Carey (Ed.), Continental Drift, a Symposium, Univ. Tasmania, 177-356.

Carvalho, A., Ribeiro, A., Cabral, J. (1983-85). Evolução paleogeográfica da bacia Cenozóica do Tejo-Sado. Bol. Soc. Geol. Portugal 24, 209-212.

Choubert, B. (1935). Recherche sur la genèse des chaînes paléozoïque et antécambriennes. Rev, Géogr. Phys. Géol. Dynam. 8(1), 1-50.

Cogné, J. (1967). L'Arc hercynien ibéro-armoricain. Publ. Inst. Fr. Pétrole, Collections Coll. et Sémin., 22 (Histoire Structurale du Golfe de Gascogne), 1, 1-23.

Cogné, J. (1971). Le massif Armoricain et sa place dans la structure des socles ouest européens: l'arc hercynien ibéro-armoricain. In Histoire structurale du Golfe de Gascogne. Editions Technip, Paris (ed. Debyser), I, 1-24.

Coke, C., Dias, R., Ribeiro, A. (2000). Evolução Geodinâmica da bacia do Douro durante o Câmbrico e o Ordovícico; um exemplo de sedimentação controlada pela tectónica. Comum. Inst. Geol. Min., 87, 5-12.

Coke, C., Dias, R., Ribeiro, A. (2003). Rheologically induced structural anomalies in transpressive regimes, J. Struct. Geol., 25/3, 409-420.

Conde, L. (1966). Direcções de correntes na base do Ordovícico do afloramento de Amêndia-Mação e sua importância paleogeográfica. Mem. Not., Publ. Mus. Lab. Geol. Univ. Coimbra. 61, 44-55.

Corretgé, L., Suarez, O., Galan, G. (1990). West Asturian-Leonese Zone; igneous rocks. In: Dallmeyer, R.D., Martínez García, E. (Eds.), Pre-Mesozoic Geology of Iberia. Springer-Verlag, Berlin, 115–128.

Dallmeyer, D., Martínez Catalán, J., Arenas, R., Gil Ibarguchi, J., Gutiérrez Alonzo, G., Farias, P., Bastida, F., Aller, J. (1997). Diachronous Variscan tectonothermal activity in the NW Iberian Massif: Evidence from $^{40}Ar/^{39}Ar$ dating of regional fabrics. Tectonophysics, 277, 307-337.

Dalziel, I. (1971). Evolution of the Scotia Arc. Nature, 233/5317, 246-252.

Dana, J. D. (1866). A texbook of geology. Philadelphia, Theodore Bliss, 354 p.

De Sitter, L. (1961). Le Précambrien dans le Chaîne Cantabrique. C. Rend. Soc. Géol. France, 9, 253 p.

De Wit, M. (1977). The evolution of the Scotia Arc as a Key to the reconstruction of southwestern Gondwanaland, Tectonophysics, 37(1-3), 53-81.

Dias da Silva, I. (2013). Geología de las Zonas Centro Ibérica y Galicia – Trás-os-Montes en la parte oriental del Complejo de Morais, Portugal/España. Ph. D. Thesis. Salamanca University, 389 p.

Dias R. (1998). Estrutura Varisca do autóctone do Terreno Ibérico, uma herança Precâmbrica?, Comun. Serv. Geol. Portugal, 85, 29-38.

Dias, R., Basile, C. (2013). Estrutura dos sectores externos da Zona Sul Portuguesa; implicações geodinâmicas. In: Dias, R., Araújo, A., Terrinha, P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1) Escolar Editora, 787-807.

Dias, R., Ribeiro, A. (1991). Finite-strain analysis in a transpressive regime (Variscan autochthon, NE Portugal). Tectonophysics, 191, 389-397.

Dias, R., Ribeiro, A. (1993). Porto-Tomar shear zone, a major structure since the begining of the Variscan orogeny. Comun. Serv. Geol. Portugal, 79, 29-38.

Dias, R., Ribeiro, A. (1994). Constriction in a transpressive regime: an example in the Ibero-Armoricain Arc. J. Struct. Geol., 16/11, 1543-1554.

Dias, R., Ribeiro, A. (1995ª). Caramulo region: a transition domain between trasnpressive constriction and plane strain in the Centro-Iberian zone. Gaia, Museu Nacional de História Natural, 11, 35-42.

Dias, R., Ribeiro, A. (1995b). The Ibero-Armorican arc: a collisional effect against an irregular continent? Tectonophysics, 246(1-3), 113-128.

Dias, R., Ribeiro, A. (1998). Interaction between major sinistral wrench faults and coeval folds in a variscan transpressive regime (NE Portugal), Comun. Serv. Geol. Portugal, 85, 19-27.

Dias, R., Ribeiro, A. (2008). Heterogeneous strain behaviour in competent layers during folding in transpressive regimes. Geodinamica Acta, 21(4), 219-229.

Dias, R., Ribeiro, A. (2013). O Varisco do sector norte de Portugal. In: Dias, R., Araújo, A., Terrinha, P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1), Escolar Editora, 59-71.

Dias, R., Mateus, A., Ribeiro, A. (2003). Strain partitioning in transpressive shear zones in the southern branch of the Variscan Ibero-Armorican Arc. Geodinamica Acta, 16, 119-129.

Dias, R., Araújo, A., Terrinha, P., Kullberg J. (Eds.) (2013a). Geologia de Portugal, Escolar Editora, vol. 1, Geologia Pré-Mesozóica de Portugal, 807 p.

Dias, R., Ribeiro, A., Coke, C., Pereira, E., Rodrigues, J., Castro, P., Moreira, N., Rebelo, J. (2013b). Evolução estrutural dos sectores setentrionais do autóctone da Zona Centro-Ibérica. In: Dias, R., Araújo, A., Terrinha, P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1), Escolar Editora, 73-147.

Díez Balda, M. (1986). El Complejo Esquisto-Grauváquico, las series paleozóicas y la estructura hercínica al Sur de Salamanca. Acta Salmanticensia, Sección Ciencias, Ediciones Universidad de Salamanca, Spain, 52, 162 p.

Díez Balda, M., Vegas, R. (1992). La estructura del Dominio de los pliegues verticales de la Zona Centro-Ibérica. In: Gutiérrez-Marco, J.C., Saavedra, J., Rábano, I. (Eds.), Paleozoico Inferior de Ibero-América, Universidad de Extremadura, 523-534.

Díez Balda, M., Vegas, R., González Lodeiro, F. (1990). Structure of the autochthonous sequences of Central Iberian-Zone. In: Dallmeyer, R. D., Martínez García, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, New York, 172–188.

Díez Balda, M., Martinez Catalán, J., Ayarza Arribas, P. (1995). Syn-collisional extensional collapse parallel to the orogenic trend in a domain of steep tectonics: the Salamanca Detachment Zone (Central Iberian Zone, Spain). J. Struct. Geol., 17, 163-182.

Díez Montes, A., Navidad, M., González Lodeiro, F., Martínez Catalán, J. (2004). Estratigrafia do Dominio del Ollo de Sapo. in: Vera, J. (Ed.), Geologia de España. SGE-IGME, Madrid, 69-75.

Dufréchoua, G, Harris, L., Corriveau, L. (2014). Tectonic reactivation of transverse basement structures in the Grenville orogen of SW Quebec, Canada: Insights from gravity and aeromagnetic data. Precambrian Research, 241, 61-84.

Du Toit, A. (1937). Our wandering continents: an hypothesis of continental drifting. Oliver and Boyd, Edinburgh, 366 p.

Eichelberger, N., McQuarrie, N. (2014). Three-dimensional (3-D) finite strain at the central Andean orocline and implications for grain-scale shortening in orogens, Geological Society of America Bulletin, 127(1-2), 87-112.

Escuder Viruete, J. (1998). Relationships between structural units in the Tormes gneiss dome (NW Iberian massif, Spain): geometry, structure and kinematics of contractional and extensional Variscan deformation. Geol. Rundschau, 87, 165-170.

Escuder Viruete, J., Arenas, R., Martinez Catalán, J. (1994). Tectonothermal evolution associated with Variscan crustal extension in the Tormes Gneiss Dome (NW Salamanca, Iberian Massif, Spain). Tectonophysics, 238, 117-138.

Escuder Viruete, J., Diez Balda, M., Rubio Pascual, F., González Casado, J., Barbero, L., Martínez Poyatos, D., Villar, P., Martínez Catalán, J. (2004). La extensión varisca tardiorogénica y las deformaciones tardías. In: Vera, J. (Ed.), Geología de España. SGE-IGME, Madrid, 87-92.

Expósito, I., Simancas, J., González Lodeiro, F., Azor, A., Martínez Poyatos D. (2002). Estructura de la mitad septentrional de la zona de Ossa-Morena: Deformación en el bloque inferior de un cabalgamiento cortical de evolución compleja, Rev. Soc. Geol. Esp., 15, 3–14.

Farias, P., Gallastegui, G., Lodeiro, F., Marquinez, J., Parra, L., Martínez Catalán, J., Macia, J., Fernandez, L. (1987). Aportaciones al conocimiento de la litoestratigrafia y estructura de Galicia Central. Mem. Mus. Labor. Miner. Geol. Fac. Ciênc. Univ. Porto, 1, 411-431.

Franke, W., Matte, P., Tait, J. (2005). Europe: Variscan orogeny. Encyclopedia of Geology (vol. 2). Elsevier, Oxford, 75–85.

Fernández, F., Aller, J., Bastida, F. (2007). Kinematics of a kilometric recumbent fold: The Courel syncline (Iberian massif, NW Spain). J. Struct. Geol., 29, 1650–1664.

Fernández-Suárez, J., Dunning, G., Jenner, G., Gutiérrez-Alonso, G. (2000). Variscan collisional magmatism and deformation in NW Iberia: constraints from U-Pb geochronology of granitoids. Journal of the Geological Society, London, 157, 565-576.

Fonseca, P., Munhá, J., Pedro, J., Rosas, F., Moita, P., Araújo, A., Leal, N. (1999). Variscan Ophiolites and High-Pressure Metamorphism in Southern Ibéria. Ophioliti, 24(2), 259-268.

Gibbons, W., Moreno, T. (Eds.), 2002. The Geology of Spain. Geological Society, London, 649 p.

Gómez Pugnaire, M., Azor, A., Fernández Soler, J., López Sánchez-Vizcaíno, V. (2003). The amphibolites from the Ossa-Morena/Central Iberian Variscan Suture (Southwestern Iberian Massif): geochemistry and tectonic interpretation. Lithos, 68, 23-42.

Gutiérrez-Alonso, G., Fernández-Suárez, J., Weil, A. (2004). Orocline triggered lithospheric delamination. Geological Society of America Special Paper, 383, 121–131.

Gutiérrez-Alonso, G., Fernandez-Suarez, J., Weil, A., Murphy, J.B., Nance, R.D., Corfu, F., Johnston, S. (2008). Self-subduction of the Pangaean global plate. Nature Geoscience, 1, 549–553.

Gutiérrez-Alonso, G., Murphy, B., Fernández-Suárez, J., Weil, A., Franco, M., Gonzalo, J. (2011). Lithospheric delamination in the core of Pangea: Sm–Nd insights from the Iberian mantle. Geology, 39(2), 155–158.

Gutiérrez-Alonso, G., Johnston, S., Weil, A., Pastor-Galán, D., Fernández-Suárez, J. (2012). Buckling an orogen: the Cantabrian orocline. GSA Today, 22, 4–9.

Gutiérrez Marco, J., San Jose, M., Pieren, A. (1990). Post-Cambrian Palaeozoic Stratigraphy of Autochthonous sequence, Central-Iberian Zone. In: Dallmeyer, R. D., Martínez Garcia, E. (Eds.), Pre- Mesozoic Geology of Iberia, Springer-Verlag, New York, 160 – 171.

Hall, L., Lamb, S., Mac Niocail, C. (2004). Cenozoic distributed rotational deformation, South Island, New Zealand. Tectonics, 23, TC2002, 1–16.

Harris, R. (2011). The Nature of the Banda Arc-Continent collision in the Timor region. In: Brown, D., Ryan, p. (Eds.), Arc-Continent Collision, Frontiers in Earth Sciences, Springer-Verlag, Berlin, 163-211.

Hirt, A., Lowrie, W., Julivert, M., Arboleya, M. (1992). Paleomagnetic results in support of a model for the origin of the Asturian arc. Tectonophysics, 213, 321-339.

Hobbs, B., Means, W., Williams, P. (1976). An outline of structural geology. John Wiley & Sons, Inc., New York. 571 p.

Iglésias, M., Ribeiro, A. (1981). La zone de cisaillement ductile de Juzbado (Salamanca) - Penalva do Castelo (Viseu); un linéament ancien réactivé pendant l'orogénèse hercynienne?. Comum. Serv. Geol. Portugal, 67(1), 89-93.

Johnston, S., Gutiérrez-Alonso, G. (2010). The North American Cordillera and West European Variscides: Contrasting interpretations of similar mountain systems. Gondwana Research, 17(2–3), 516–525.

Johnston, S., Weil, A., Gutiérrez-Alonso, G. (2013). Oroclines: thick and thin. Geol. Soc. Am. Bull. ,125(5-6), 643-663.

Julivert, M., Marcos, A. (1973). Superimposed folding and flexural conditions in the Cantabrian Zone (Hercynian Cordillera, NW Spain). American Journal of Science, 273, 353–375.

Julivert, M., Arboleya, M.L. (1984). Curvature increase and structural evolution of the core (Cantabrian Zone) of the Ibero-Armorican Arc. Sci. Géol. Bull., 37, 5-11.

Julivert, M., Martínez, F. (1983). Estructura de conjunto e visión global de la Cordillera Herciniana. In: Comba, J. A. (Coord.), Geologia de España, Libro Jubilar de J.M. Rios, IGME, 612-630.

Julivert, M., Fonteboté, J., Ribeiro, A., Conde, L. (1972). Mapa Tectónico de la Península Ibérica Y Baleares. 1:1 000 000. Instituto Geológico Y Minero de España. 113 p.

Julivert, M., Marcos, A., Pérez-Estaún, A. (1977). La structure de la Chaîne Hercynienne dans le secteur Ibérique de l'Arc Ibéro-Armoricain. La chaîne varisque d'Europe moyenne et occidentale, Colloque Int. CNRS, Rennes, 243, 429-440.

Julivert, M., Vegas, R., Rotz, J. Martínez Rius, A. (1983). La estructura de la extension SE de la Zona Centroiberica com metamorfismo de Bajo Grado. In: Comba, J. A. (Coord.), Geologia de España, Libro Jubilar de J.M. Rios, IGME, 477-490.

Keller, M., Bahlburg, H., Reuther, C. (2008). The transition from passive to active margin sedimentation in the Cantabrian Mountains, Northern Spain: Devonian or Carboniferous? Tectonophysics, 461, 414–427.

Lefort, J. (1979). Iberian–Armorican arc and Hercynian orogeny in western Europe. Geology, 7, 384–388.

Lefort, J. (1989). Basement correlation across the north Atlantic, Springer-Verlag, Berlin, 148 p.

Lefort, J., Van der Voo, R. (1981). A Kinematic Model for the Collision and Complete Suturing between Gondwanaland and Laurussia in the Carboniferous. The Journal of Geology, 89(5), 537-550.

Linnemann, U., Pereira, F., Jeffries, E., Drost, K., Gerdes, A. (2008). The Cadomian Orogeny and the opening of the Rheic Ocean: the diacrony of geotectonic processes constrained by LA-ICP-MS U/Pb zircon dating (Ossa Morena and Saxo-Thuringian Zones, Iberian and Bohemian Massifs). Tectonophysics, 461, 21-43.

Llana-Fúnez, S., Marcos, A. (2001). The Malpica-Lamego line: A major crustal-scale shear zone in the Variscan Belt of Iberia. J. Struct. Geol., 23, 1299-1312.

Llana-Fúnez, S., Marcos, A. (2002). Structural record during exhumation and emplacement of high-pressure-low to intermediate-temperature rocks in the Malpica-Tui (Variscan Belt of Iberia). In: Martínez Catalán, J.R., Hatcher. R., Arenas. D.R., Díaz Garcia, F. (Eds.) Variscan-Appalachian dynamics: the building of the late Palaeozoic basement. Geol. Soc. Am., Boulder, Colorado, Special Paper, 364, 125-142.

Lorenz, V. (1976). Formation of Hercynian subplates. Possible cause and consequences. Nature, 262, 374–377.

Lorenz, V., Nicholls, I. (1984). Plate and intraplate processes of Hercynian Europe during the Late Paleozoic. Tectonophysics, 107, 25-26.

Lotze, F. (1945). Zur Gliederung der Varisziden der Iberischen Meseta. Geotektonische Forschungen, 6, 78-92.

Lotze, F. (1956). Das Präkambrium Spaniens. Neues Jahrb Geol Paläontol Monatsh, 8, 599-612.

Lotze, F. (1963). Vie Varischischen Gebirgszusammenhägne im Westlichen Europe. Giornale di Geol., sér. 2, 31, 393-412.

Maas, K. (1974). The geology of Liebana, Cantabrian Mountains, Spain: deposition and deformation in a flysch area. Leidse Geol. Meded., 49, 379-465.

Macaya, J. (1981). Estudio geológico estructural de la Sierra de Francia (Provincia de Salamanca y Cáceres). Cuadernos Geología Ibérica, 7, 567-576.

Macedo, J., Marshak, S. (1999). Controls on the geometry of fold-thrust belt salients. Geological Society of America Bulletin, 111, 1808–1822.

Machado, G., Hladil, J., Koptíková, L., Fonseca, P., Rocha, F., Galle, A. (2009). The Odivelas Limestone: Evidence for a Middle Devonian reef system in western Ossa-Morena Zone. Geologica Carpathica, 60(2), 121-137.

Machado, G., Hladil, J., Koptikova, L., Slavik, L., Moreira, N., Fonseca, M., Fonseca, P. (2010). An Emsian-Eifelian Carbonate-Volcaniclastic Sequence and the possible Record of the basal choteč event in western Ossa-Morena Zone, Portugal (Odivelas Limestone), Geologica Belgica, 13(4), 431-446.

Marcos, A. (2004). Zona Asturoccidental -Leonesa. In: Vera, J. (Ed.), Geologia de España. SGE-IGME, Madrid, 49.

Marcos, A., Pulgar, J. (1982). An Approach to the tectonostratigraphic evolution of the Cantabrian foreland thrust and fold belt, Hercynian Cordillera of NW Spain. Neues Jahrb. Geol. Palaeontol., Abh., 163(2), 256-260.

Marcos, A., Martínez Catalán, J., Gutiérrez-Marco, J., Pérez-Estaún, A. (2004). Estratigrafia y paleogeografia da Zona Asturoccidental-Leonesa. In: Vera, J. (Ed.). Geologia de España. SGE-IGME, Madrid, 49-52.

Marshak, S. (1988). Kinematics of orocline and arc formation in thin-skinned orogens. Tectonics, 7, 73–86.

Marshak, S. (2004). Salients, recesses, arcs, oroclines, and syntaxes - A review of ideas concerning the formation of map-view curves in fold-thrust belts. In K. R. McClay (Ed.), Thrust tectonics and hydrocarbon systems, AAPG Memoir, 82, 131 – 156.

Marquínez, J. (1978). Estudio geológico del sector SE de los Picos de Europa (Cordillera Cantábrica, NW de España). Trab. Geol. Univ. Oviedo, 10, 295-315.

Martínez Catalán, J. (1985). Estratigrafía y estructura del Domo de Lugo (sector W da zona Asturoccidental-Leonesa). Corpus Geol. Gallaeceae (2ª serie), 2, 1-291.

Martínez Catalán, J. (1990). West Asturian-Leonese Zone: Introduction. In: Dallmeyer, D., Martinez Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia. Springer-Verlag, 91.

Martínez Catalán, J. (2011a). The Central Iberian arc: implications for the Iberian Massif. Geogaceta, 50/1, 7-10.

Martínez Catalán, J. (2011b). The Central Iberian arc, an orocline centered in the Iberian Massif and some implications for the Variscan belt. Int. J. Earth Sc., 101, 1299–1314.

Martínez Catalán, J. (2011c). Are the oroclines of the Variscan belt related to late Variscan strike-slip tectonics? Terra Nova, 23, 241–247.

Martínez Catalán, J., Pérez Estaún, A., Bastida, F., Pulgar, J., Marcos, A. (1990). West Asturian-Leonese Zone; structure. In: Dallmeyer, R.D., Martínez García, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 103–114.

Martínez Catalán, J., Hacar Rodrguez, M., Villar Alonzo, P., Pérez Estaún, A., González Lodeiro, F. (1992). Lower Paleozoic extensional tectonics in the limit between the West Asturian-Leonese and Central-Iberian Zones of the Variscan Fold Belt in NW Spain. Geol. Rundschau, 85, 545-560.

Martínez Catalán, J., Martínez Poyatos, D., Bea, F. (2004). Introducción to Zona Centroibérica. in: Vera, J., (Ed.), Geología de España. SGE-IGME, Madrid, 68-69.

Martínez Catalán, J., Fernández-Suárez, J., Meireles, C., González Clavijo, E., Belousova, E., Saeed, A. (2008). U-Pb detrital zircon ages in synorogenic deposits of the NW Iberian Massif: interplay of syntectonic sedimentation and thrust tectonics. Journal of the Geological Society of London, 165, 687-698.

Martínez Catalán, J., Rubio Pascual, F., Díez Montes, A., Díez Fernández, R., Barreiro, J., Dias da Silva, I., Clavijo, E., Ayarza., P., Alcock, J. (2014). The late Variscan HT-LP metamorphic event in NW and Central Iberia: relationships to crustal thickening, extension, orocline development and crustal evolution. In: Schulmann, K., Martínez Catalán, J., Lardeaux, J., Janousek, V., Oggiano, G. (Eds.), The Variscan Orogeny: Extent, Timescale and the Formation of the European Crust. Geol. Soc. London, Sp. Publ., 405, 225-247.

Martínez Poyatos, D. (2002). Estructura del borde meridional de la Zona Centroibérica y su relación con el contacto entre las zonas Centroibérica y de Ossa-Morena. Laboratorio Xeologico de Laxe, Serie Nova Terra, 18, 295 p.

Martínez Poyatos, D., Díez Balda, M., Macaya, J., González Lodeiro, F., Martínez Catalán, J., Vegas, R. (2004). El acortamiento varisco inicial del Dominio del Complejo Esquisto-grauváquico. In: Vera, J. (Ed.), Geologia de España. SGE-IGME, Madrid, 84-87.

Matte, Ph. (1968). La structure de la virgation hercynienne de Galice (Espagne). Revue de Géologie Alpine, 44, 155-280.

Matte, Ph. (1986). Tectonics and Plate Tectonics Model for the Variscan Belt of Europe. Tectonophysics, 126, 329-374.

Matte, Ph. (1991). Accretionary history and crustal evolution of the Variscan belt in Western Europe. Tectonophysics, 196, 309-337.

Matte, Ph. (2001). The Variscan collage and orogeny (480-290 Ma) and the tectonic definition of the Armorica microplate: a review. Terra Nova, 13, 122-128.

Matte, Ph. (2002). Variscides between the Appalachians and the Urals: similarities and differences between Palaeozoic subduction and collision belts. In: Martínez Catalán, J., Hatcher, R., Arenas R., Díaz Garcia, F. (Eds.) Variscan-Appalachian dynamics: the building of the late Paleozoic basement. Geol. Soc. Am., 364, 239-251.

Matte, Ph., Burg, J. (1981). Sutures, thrusts and nappes in the Variscan arc of western Europe: Plate tectonics interpretation. In: McClay, K. R., Price, N. J. (Eds.), Thrust and Nappe Tectonics, Geol. Soc. London, Spec. Publication, 8, 353-358.

Matte, Ph., Ribeiro, A. (1975). Forme et orientation de l´ellipsoïde de déformation dans la virgation hercynienne de Galice. Relations avec le plissement et hypothèses sur la genèse de l´arc ibéro-armoricain. C. R. Académie Sciences Paris, D/280, 2825-2828.

Mendes Victor, L., Miranda, M., Matias, L. (1993). Crustal structure of western Iberia from geophysical studies. Publ. Inst. Geog. Nac. España, Série monográfica, 10, 179-196.

Metodiev, D., Romão, J., Dias, R., Ribeiro, A. (2009). Sinclinal de Vila Velha de Ródão (Zona Centro-Ibérica, Portugal): litostratigrafia, estrutura e modelo de evolução da tectónica Varisca. Comunicações Geológicas, 96, 5-18.

Miranda, M. (1990). O levantamento aeromagnético de Portugal - contribuição para o conhecimento da estrutura geológica do continente português. Ph. D. Thesis, Lisbon University, 146 p.

Miranda, M., Mendes Victor, L. (1990). Aeromagnetic Map of Portugal, 1/1 000 000. Serv. Geol. Portugal.

McKerrow, W., Ziegler, A. (1972). Paleozoic oceans. Nature (London), Phys. Sci., 240, 92-94.

Moreira, N., Búrcio, M., Dias, R., Coke, C. (2010a). Partição da deformação Varisca nos sectores de Peso da Régua e Vila Nova de Foz Côa (Autóctone da Zona Centro Ibérica). Comunicações Geológicas, 97, 147-162.

Moreira, N., Machado, G., Fonseca, P., Silva, J., Jorge, R.., Mata, J. (2010b). The Odivelas Palaeozoic volcanosedimentary sequence: Implications for the geology of the Ossa-Morena Southwestern border. Comunicações Geológicas, 97, 129-146.

Moreira, N., Araújo, A., Pedro, J., Dias, R. (2014). Evolução geodinâmica da Zona de Ossa-Morena no contexto do SW Ibérico durante o Ciclo Varisco. Comunicações geológicas, 101(I) 275-278.

Murphy, J., Gutierrez-Alonso, G., Nance, R., Keppie, J., Quesada, C., Strachan, R., Dostal, J. (2006). Origin of the Rheic Ocean: Rifting along a Neoproterozoic suture? Geology, 34, 325–328.

Nance, R., Gutierrez-Alonso, G., Keppie, J., Linnemann, U., Murphy, J., Quesada, C., Strachan, R., Woodcock, N., (2012). A brief history of the Rheic Ocean. Geoscience Frontiers, 3/2, 125-135.

Noronha, F., Ramos, J., Rebelo, J., Ribeiro, A., Ribeiro, M. L. (1981). Essai de corrélation des phases de déformation hercynienne dans le Nord-Ouest Péninsulaire. Leidse Geologische Mededelingen, 52(1), 87-91.

Oliveira, J., Oliveira, V., Piçarra, J. (1991). Traços gerais da evolução tectono-estratigráfica da Zona de Ossa Morena, em Portugal: síntese crítica do estado actual dos conhecimentos. Comun. Serv. Geol. Port., 77, 3-26.

Oliveira, J.T., Relvas, J., Pereira, Z., Matos, J., Rosa, C. Rosa, D., Munhá, J. Fernandes, P., Jorge, R., Pinto, A. (2013). Geologia Sul portuguesa, com ênfase na estratigrafia, vulcanologia física, geoquímica e mineralizações da faixa piritosa. In: Dias, R., Araújo, A., Terrinha, P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1), Escolar Editora, 673-767.

Pamplona, J., Ribeiro, A. (2013). Evolução geodinâmica da região de Viana do Castelo (Zona Centro-Ibérica, NW de Portugal). In: Dias, R., Araújo, A., Terrinha, P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1), Escolar Editora, 149-203.

Pamplona, J., Rodrigues, B., Llana-Fúnez, S., Simões, P., Ferreira, N., Coke, C., Pereira, E., Castro, P., Rodrigues, J.. (2016). Structure and Variscan evolution of Malpica–Lamego ductile shear zone (NW of Iberian Peninsula). In: Mukherjee, S., Mulchrone, K. F. (Eds.), Ductile Shear Zones: From Micro- to Macro-scales, John Wiley & Sons, 206-223.

Parés J., Van der Pluijm, B. (2004). Correlating magnetic fabrics with finite strain: Comparing results from mudrocks in the Variscan and Appalachian Orogens. Geologica Acta, 2 /3, 213-220.

Parés, J., Van der Voo, R., Stamatakos, J., Pérez-Estaún, A. (1994). Remagnetization and postfoldingd oroclinal rotations in the Cantabrian/Asturian arc, northern Spain. Tectonics, 13(6), 1461-1471.

Pastor-Galán, D., Gutiérrez-Alonso, G., Weil, A. (2011). Orocline timing through joint analysis: Insights from the Ibero-Armorican Arc. Tectonophysics, 507, 31–46.

Pastor-Galán, D., Groenewegen, T., Brouwer, D., Krijgsman, W., Dekkers, M. (2015). One or two Oroclines in the Variscan orogen of Iberia? Implications for Pangea amalgamation. Geology, 43(6), 527-530.

Pereira, E. (1987). Estudo geológico-estrutural da área de Celorico de Basto e sua interpretação geodinâmica. Ph. D. Thesis, Lisbon University, 274 p.

Pereira, E. (1988). Soco Hercínico da Zona Centro-Ibérica – Evolução Geodinâmica. Geonovas, 10, 13-35.

Pereira, E. (1989). Carta Geológica de Portugal na Escala 1:50.000 e notícia explicativa da Folha 10-A (Celorico de Basto), Serv. Geol. Portugal, 53 p.

Pereira, E., Ribeiro, A. (1983). Tectónica do sector noroeste da Serra do Marão. Com. Serv. Geol. Port., 69(2), 283-290.

Pereira, E., Ribeiro, A. (1992). Paleozóico. In: Pereira, E. (Ed.), Carta Geológica de Portugal na Escala 1:200.000 e notícia explicativa da Folha 1. Serviços Geológicos de Portugal, 9-26.

Pereira, E., Ribeiro, A., Meireles, C. (1993). Cisalhamentos hercínicos e controlo das mineralizações de Sn-W, Au e U na Zona Centro-Ibérica em Portugal. Cuaderno Lab. Xeolóxico de Laxe, La Coruña, 18, 89-119.

Pereira, M., Silva, J., Ribeiro, C. (2010). The role of bedding in the formation of fault–fold structures, Portalegre - Esperanca transpressional shear zone, SW Iberia. Geol. J., 45, 521–535.

Pereira, M., Chichorro, M., Johnston, S., Gutiérrez-Alonso, G., Silva, J., Linnemann, U., Hofmann, M., Drost, K. (2012). The missing Rheic Ocean magmatics arcs: Provenance analysis of Late Paleozoic sedimentary clastic rocks of SW Iberia. Gondwana Research, 22, 882-891.

Pereira, Z., Meireles, C., Pereira, E. (1999). Upper Devonian Palinomorphs of the NE sector of Trás-os-Montes (Central Ibéria Zone). XV Reunião do Oeste Peninsular – International Meeting on Cadomian orogens, Badajoz, 201-206.

Pérez Estaún, A. (1974). Aportaciones al conocimiento del Carbonífero de San Clodio (Prov. de Lugo). Breviora Geologica Asturica, 18, 23-25.

Pérez Estaún, A. (1990). Introduction to the Cantabrian. In: Dallmeyer, R. D., Martínez-Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 7-8.

Pérez Estaún, A., Bastida, F. (1990). Structure of the Cantabrian. In: Dallmeyer, R. D., Martínez-Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 55-69.

Pérez-Estaun, A., Bea F. (2004). Macizo Iberico. in: Vera, J.A. (Ed.). Geologia de España. Sociedad Geológica de España, Geologia de España. SGE-IGME, Madrid, 19–230.

Pérez Estaún, A., Bastida, F., Alonso, J., Marquinez, J., Aller, J., Alvarez, M., Marcos, A., Pulgar, J. (1988). A thin-skinned tectonics model for an arcuate fold and thrust belt, the Cantabrian Zone (Variscan Ibero-Armorican Arc), Tectonics, 7, 517-537.

Pérez Estaún, A., Bastida, F., Martínez Catalán, J., Gutierrez Marco, J., Marcos, A., Pulgar, J. (1990). Stratigraphy of the West Asturian-Leonese Zone. In: Dallmeyer, R. D., Martínez-Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 92-102.

Pérez Estaún, A., Martínez Catalán, J., Bastida, F. (1991). Crustal thickening and deformation sequence in the footwall to the suture of the Variscan belt of northwest Spain, Tectonophysics, 191, 243-253.

Perroud, H., Bonhommet, N. (1981). Palaeomagnetism of the Ibero-Armorican arc and the Hercynian orogeny in Western Europe. Nature, 292, 445–448.

Pohn, H. (2000). Lateral Ramps in the Folded Appalachians and in Overthrust Belts Worldwide—A Fundamental Element of Thrust-Belt Architecture. USGS Bullletin, 2163, 63 p.

Quesada, C. (1990). Ossa Morena Zone, Introdution, In: Dallmeyer, R. D., Martínez-Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 249-251.

Quesada, C., Dallmeyer, R. (1994). Tectonothermal evolution of the Badajoz-Cordóba shear zone (SW Iberia): characteristics and $^{40}Ar/^{39}Ar$ mineral age constraints. Tectonophysics, 231, 195-213.

Quesada, C., Fonseca, P., Munhá, J., Oliveira, J., Ribeiro, A. (1994). The Beja-Acebuches Ophiolite (Southern Iberia Variscan fold belt): geological characterization and geodynamic significance. Bol. Geol. Y Min., 105/1, 3-49.

Ramsay, J. (1967). Folding and Fracturing of Rocks. McGraw-Hill, New York, 568 p.

Ribeiro, A. (1974). Contribution à l'étude tectonique de Trás-os-Montes Oriental. Mem. Serv. Geol. Portugal, 24, 168 p.

Ribeiro, A. (1990). Central-Iberian Zone: Introduction. In: Dallmeyer, R. D., Martínez-Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 143-144.

Ribeiro, A., Silva, J. (1983). Structure of the South Portuguese Zone. In: Lemos de Sousa, M.J., Oliveira, J.T. (Eds.), The Carboniferous of Portugal, Serv. Geol. Portugal, 29, 83-89.

Ribeiro, A., Pereira, E., Dias, R. (1990). Allochthonous Sequences: Structure in the Northwest of the Iberian Peninsula. In: Dallmeyer, R. D., Martínez-Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 220-236.

Ribeiro, A., Dias, R., Silva, J. B. (1995). Genesis of the Ibero-Armorican Arc. Geodinamica Acta, 8(2), 173-184.

Ribeiro, A., Munhá, J., Dias, R., Mateus, A., Pereira, E., Ribeiro, L., Fonseca, P., Araújo, A., Oliveira, T., Romão, J., Chaminé, H., Coke, C., Pedro, J. (2007). Geodynamic evolution of SW Europe Variscides. Tectonics, 26, 1-24.

Ribeiro, A., Munhá, J., Mateus, A., Fonseca, P., Pereira, E., Noronha, F., Romão, J., Rodrigues, J.F., Castro, P., Meireles, C., Ferreira, N. (2009). Mechanics of thick-skinned Variscan overprinting of Cadomian basement (Iberian Variscides). Comptes Rendus Geoscience, 341(2-3), 127-139.

Ribeiro, A., Munhá, J., Fonseca, P.E., Araújo, A., Pedro, J.C., Mateus, A., Tassinari, C., Machado, G., Jesus, A. (2010). Variscan ophiolite belts in the Ossa-Morena Zone (Southwest Iberia): geological characterization and geodynamic significance. Gondwana Res, 17, 408–421.

Ribeiro, A., Romão, J., Munhá, J., Rodrigues, J., Pereira, E., Mateus, A., Araújo, A. (2013). Relações tectonostratigráficas e fronteiras entre a Zona Centro-Ibérica e a Zona Ossa-Morena do Terreno Ibérico e do Terreno Finisterra. In: Dias, R., Araújo, A., Terrinha P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1), Escolar Editora, 439-481.

Ribeiro, M. L., Ribeiro, A. (1974). Signification paleogéographique et tectonique de la présence de galets de roches métamorphiques dans un flysch d'âge dévonien supérieur du Trás-os-Montes Oriental (Nord-Est du Portugal). C. R. Acad. Sc. Paris, série D, 278, 1361- 1363.

Richard, P., Cobbold, P. (1990). Experimental insights into partitioning of fault motions in continental convergent wrench zones. Annales Tectonicae, IV/2, 35-44.

Ries, A. (1978). The opening of the Bay of Biscay - a review. Earth-Science Reviews, 14, 35-63.

Ries, A., Shackleton, R. (1976). Patterns of strain variation in arcuate fold belts. Phil. Trans. R. Soc. Lond. A283, 281-288.

Ries, A., Richardson, A., Shackleton, R. (1980). Rotation of the Iberian arc: paleomagnetic results from north Spain. Earth and Planetary Science Letters, 50, 301–310.

Robardet, M. (2002). Alternative approach to the Variscan Belt in southwestern Europe. Variscan–Appalachian Dynamics. Geol. Soc. Am., 364, 1–15.

Rodrigues, B., Chew, D., Jorge, R., Fernandes, P., Veiga-Pires, C., Oliveira, J. (2014). Idades U-Pb de zircões detríticos do Grupo do Flysch do Baixo Alentejo, Zona Sul Portuguesa, Comunicações geológicas 101/I, 301-305.

Rodrigues, J. (2008). Estrutura da Arco da Serra de Santa Comba – Serra da Garraia; Parautóctone de Trás-os-Montes. Ph.D. Thesis, Lisbon Univ., 308 p.

Rodrigues, J., Coke, C., Dias, R., Pereira, E., Ribeiro, A. (2005). Transition from autochthonous to parautochtonous deformation regimes in Murça-Marão sector (Central-Iberian Zone, northern Portugal). In: Carosi, R., Dias, R., Iacopini D., Rosebaum, G. (Eds.) The southern Variscan belt, Journal of the Virtual Explorer, Electronic Edition, ISSN 1441-8142, 19, paper 8.

Rodrigues, J., Pereira, E., Ribeiro, A. (2013). Complexo de mantos parautóctones do NE de Portugal: estrutura interna e tectonoestratigrafia. In: Dias, R., Araújo, A., Terrinha P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1), Escolar Editora, 275-331.

Rodríguez Alonso, M., Balda, D., Perejón, A., Pieren, A., Liñan, E., López Díaz, F., Moreno, F., Gámez Vintaned, J., González Lodeiro, F., Martínez Poyatos, D., Vegas, R. (2004). Estratigrafía do Dominio del Complejo Esquisto-Grauváquico. in: Vera, J. (Ed.). Geología de España. SGE-IGME, Madrid, 78-83.

Rölz, P. (1975). Beiträge zum Aufbau des junpräkambrischen und attpalaozoischen Grundgebirges in den Provinzen Salamanca und Cáceres (Sierra de Tamames, Sierra de Francia und östliche Sierra de Gata) Spanish (Auszug). Münster Forsch, Geol. Paläont. Heft, 36, 1-68.

Romão, J. (2000). Carta Geológica de Portugal na Escala 1:50.000 e notícia explicativa da Folha 28-A (Mação), Inst. Geol. Min., Lisboa, 85 p.

Romão, J., Metodiev, D., Dias, R., Ribeiro, A. (2013). Evolução geodinâmica dos sectores meridionais da Zona Centro-Ibérica. In: Dias, R., Araújo, A., Terrinha P., Kullberg, J.C. (Eds.), Geologia de Portugal (vol. 1), Escolar Editora, 206-257.

Rubio Pascual, F., Arenas, R., Martínez Catalán, J., Rodríguez Fernández, L., Wijbrans, J. (2013). Thickening and exhumation of the Variscan roots in the Iberian Central System: tectonothermal processes and $^{40}Ar/^{39}Ar$ ages. Tectonophysics, 587, 207–221.

San José, M., Pieren, P., Garcia Hidalgo, F., Vilas, L., Herranz, P., Pelaez, J., Perejon, A. (1990). Ante-Ordovician stratigraphy of Autochthonous sequence, Central-Iberian Zone. In: Dallmeyer, D., Martinez Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia. Springer-Verlag, 147-159.

Schulmann, K., Martínez Catalán, J., Lardeaux, J., Janousek, V., Oggiano. G. (Eds.) (2014). The Variscan Orogeny: extend, timescale and the formation of the European Crust. Geological Society of London, 405, 406 p.

Schultz, G. (1858). Atlas geológico y topográfico de Asturias. Lit. de G. Pfeiffer (2 maps +1 plate).

Sengör, A. (2013). The Pyrenean Hercynian Keirogen and the Cantabrian Orocline as genetically coupled structures. Journal of Geodynamics, 65, 3– 21.

Shaw, J., Johnston, S., Gutierrez-Alonso, G., Weil, A. (2012a). Oroclines of the Variscan orogen of Iberia: Paleocurrent analysis and paleogeographic implications. Earth and Planetary Science Letters, 329/330, 60–70.

Shaw, J., Johnston, S., Gutierrez-Alonso, G., Weil, A. (2012b). Oroclines of the Variscan orogen of Iberia: Paleocurrent analysis, U-Pb detrital zircon age dating, and paleogeographic implications. Geotemas, 13, 116.

Shaw, J., Gutierrez-Alonso, G., Johnston, S., Pastor Galán, D. (2014). Provenance variability along the Early Ordovician north Gondwana margin: Paleogeographic ant tectonic implications of U-Pb detrital zircon ages from the Armorican Quartzite of the Iberian belt. Geol. Soc. Am. Bull., 126, 702–719.

Shaw, J., Johnston, S., Gutierrez-Alonso, G. (2015). Orocline formation at the core of Pangea: A structural study of the Cantabrian orocline, NW Iberian Massif. Lithosphere, 7-6; 653–661. doi: 10.1130/L461.1.

Shelley, D., Bossière, G. (2000). A new model for the Hercynian orogen of Gondwanan France and Iberia. J. Struct. Geol., 22, 757-776.

Shelley, D., Bossière, G. (2002). Megadisplacements and the Hercynian orogen of Gondwanan France and Iberia. In: Martínez Catalán, J., Hatcher Jr., R., Arenas, R., Díaz García, F. (Eds.), Variscan-Appalachian dynamics: The building of the late Paleozoic basement: Boulder, Colorado. Geological Society of America, Special Paper, 364, 209–222.

Silva J., Oliveira J., Ribeiro A. (1990). South Portuguese Zone. Structural outline. In: Dallmeyer, D., Martinez Garcia, E. (Eds.), Pre-Mesozoic Geology of Iberia. Springer-Verlag, 348-362.

Silva, J., Mata, J., Moreira, N., Fonseca, P., Jorge, R., Machado, G. (2011). Evidência para o funcionamento, desde o Devónico inferior, de subducção no bordo meridional da zona de Ossa-Morena, VIII Congresso Ibérico de Geoquímica - XVII Semana de Geoquímica, Instituto Politécnico de Castelo Branco.

Simancas, J., Martínez Poyatos, D., Expósito, I., Azor, A., González Lodeiro, F. (2001). The structure of a major suture in the SW Iberian Massif: the Ossa-Morena/Central Iberian contact. Tectonophysics, 332, 295–308.

Simancas, J. F., *et al*. (2003). Crustal structure of the transpressional Variscan orogen of SW Iberia: SW Iberia deep seismic reflection profile (IBERSEIS), Tectonics, 22(6), 1062.

Simancas, J., Ayarza, P., Azor, A., Carbonell, R., Martínez Poyatos, D., Pérez-Estaún, A., González Lodeiro, F. (2013). A seismic geotraverse across the Iberian Variscides: orogenic shortening, collisional magmatism, and orocline development. Tectonics, 32, 417–432.

Stampfli, G. (1996). The intra-alpine terrain: a paleotethyan remnant in the Alpine Variscides. Eclogae Geologicae Helvetiae, 89, 13–42.

Stampfli, G., Borel, G. (2002). A plate tectonic model for the Paleozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrones, Earth Planet. Sci. Lett., 196, 17-33.

Staub, R. (1927). Ideas sobre la tectónica de España. Spanish version and prologue: Carbonell, T. Real Academia de Ciencias, Bellas Letras y Nobles Artes de Córdoba, 1–83.

Stille, H. (1924). Grundfragen der vergleichenden tektonic. Brontraeger, 443 p.

Suárez, O., Corretge, L., Martínez, F. (1990). West Asturian-Leonese Zone; distibution an characteristics of Hercynian metamorphism. In: Dallmeyer, R.D., Martínez García, E. (Eds.), Pre-Mesozoic Geology of Iberia, Springer-Verlag, Berlin, 129-133.

Sutherland, R. (1999). Cenozoic bending of New Zealand basement terranes and Alpine Fault displacement: a brief review. New Zealand Journal of Geology and Geophysics, 42, 295-301.

Suess, E. (1888). Das Antlitz der Erde, vol. II. F. Tempsky, Prag and Wien, and G. Freytag, Leipzig, IV, 704 p.

Tapponier, P., Molnar, P. (1976). Slip-line field theory and large scale continental tectonics. Nature, 264, 319-324.

Teixeira, C., Pais, J. (1973). Sobre a presença do Devónico na região de Bragança (Guadramil e Mofreita) e de Alcañices (Zamora). Bol. Soc. Geol. Portugal, 18(2-3), 199-202.

Tena, M. (1980=. Mapa Geológico Nacional, E. 1/50 000, 704/11-28, Cáceres, Inst. Geol. Min. España, Madrid.

Tex, E., Floor, P. (1971). A synopsis of the geology of the western Galicia. Histoire structurale du Golfe de Gascogne. Technip., 1, 3-13.

Treloar, P., Coward, M., Chambers, A., Izatt, C., Jackson, K. (1992). Thrust geometries, interferences and rotations in the northwest Himalaya. In: McClay, K. (Ed.), Thrust Tectonics. Chapman and Hall, London, 25-342.

Twiss, R., Moores, E. (1992). Structural Geology. W. H. Freeman and Company, 532 p.

Valle Aguado, B. (1992). Geología estructural de la Zona de Cisalla de Porto-Tomar en la región de Oliveira de Azeméis-Serra de Arada (Norte de Portugal). Ph. D. Thesis, Salamanca Univ., 254 p.

Van der Voo, R. (2004). Paleomagnetism, oroclines, and growth of the continental crust. GSA Today, 14, 4–9.

Vera, J. (Ed.) (2004). Geología de España. SGE-IGME, Madrid, 890 p.

Vroon, P., van Bergen, M., Klaver, G., White, W. (1995). Strontium, neodymium, and lead isotopic and trace-element signatures of the East Indonesian sediments: provenance and implications for Banda arc magma genesis. Geochimica et Cosmochimica Acta, 59/12, 2573-2598.

Weil, A. (2006). Kinematics of orocline tightening in the core of an arc: paleomagnetic analysis of the Ponga Unit, Cantabrian Arc, northern Spain. Tectonics, 25, 1–23, TC3012.

Weil, A., Sussman, A. (2004). Classifying curved orogens based on timing relationships between structural development and vertical-axis rotations. In: Sussman, A., Weil, A. (Eds.), Orogenic curvature: Integrating paleomagnetic and structural analyses. Geological Society of America Special Paper, 383, 1–15.

Weil, A., Gutiérrez-Alonso, G., Conan, J. (2010). New time constraints on lithospheric scale oroclinal buckling of the Ibero-Armorican arc: a palaeomagnetic study of earliest Permian rocks from Iberia. Journal of the Geological Society of London, 167, 127-143.

Weil, A., Gutiérrez-Alonso, G., Johnston, S., Pastor-Galán, D. (2013). Kinematic constraints on buckling in a lithospheric-scale orocline along the northern margin of Gondwana: a geologic synthesis. Tectonophysics, 582, 25-49.