VII.1

Tectonostratigraphy of western block of Porto-Tomar Shear zone; the Finisterra Terrane

Index

VII.1.1. Introduction and General Framework

The Iberian Massif has a well-developed arcuate pattern, induced by the genesis of the Variscan Ibero-Armorican Arc (Fig. 1A; Dias *et al*., 2016). Its internal domains with a WNW-ESE to NW-SE general trend (e.g. Dias *et al*., 2013; Moreira *et al*., 2014) are interrupted to the west by one of the most important Iberian Variscan structures, the Porto-Tomar-Ferreira do Alentejo shear zone (Fig. 1B; PTFSZ). The geodynamic evolution of this shear zone is controversial, presenting a polyphasic evolution not only during the Variscan orogeny, but also during Meso-

Cenozoic times (Ribeiro *et al*., 2013). The superposition of different tectonic events difficult its geodynamical interpretation.

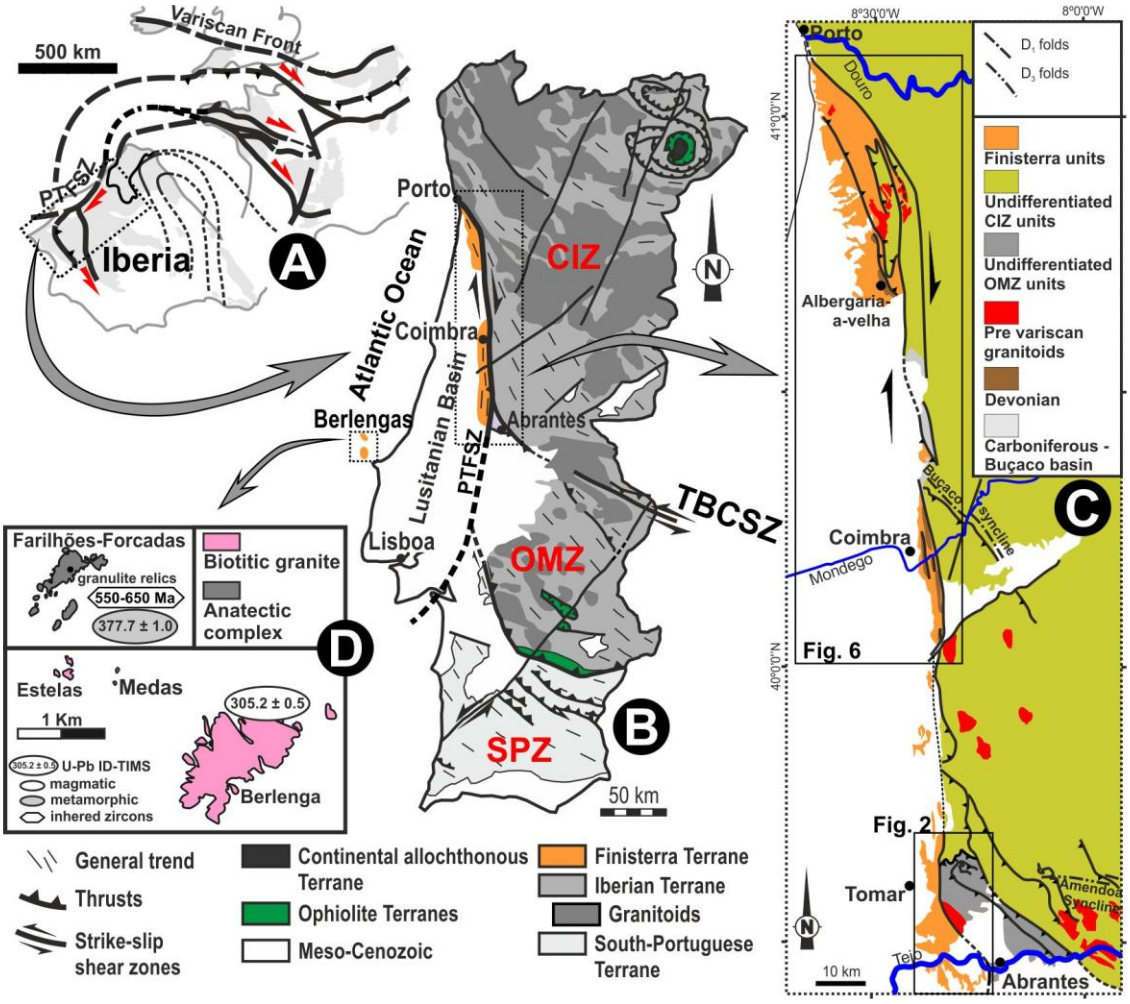


Figure 1 –The Finisterra Terrane in the context of the Iberian Variscides Massif:

A – The Ibero-Armorican Arc general pattern (adapted from Dias *et al*., 2016);

B – General overview of Finisterra Terrane (adapted from LNEG, 2010; Ribeiro *et al*., 2013);

C – The Finisterra outcrops in the PTFSZ vicinity (adapted from Chaminé *et al*., 2003a; LNEG, 2010; Romão *et al*, 2014);

D – The Berlengas islands main geological features (adapted from Bento dos Santos *et al*., in press).

In fat, the complexity of its evolution led to strongly different interpretations. Some authors consider the PTFSZ as the eastern boundary of Finisterra Terrane, a new tectonostratigraphic zone in Iberia (Ribeiro *et al*., 2007; 2009; 2013; Romão *et al*., 2013; 2014; Moreira *et al*., 2016a; 2016b). In such approach the PTFSZ is interpreted as a lithospheric transform fault (Ribeiro *et al*., 2007; 2013) active since, at least, the early Devonian stages of the Variscan Cycle (Dias and Ribeiro, 1993), but possibly reactivating an Early Variscan (or even Cadomian) structure. A different interpretation is sometimes considered and the PTFSZ is considered a major high-strain Carboniferous dextral shear zone that dismembered the western edge of Tomar-Badajoz-Córdoba Shear Zone (TBCSZ; Pereira *et al*., 2010). Similar chronological interpretation is taken by Gutiérrez-Alonso *et al*. (2015), which relate this first order structure to the Iberian Orocline genesis.

Recent geological mapping in Abrantes-Tomar sector (Fig. 1C) emphasize a N-S High Temperature (HT) tectonostratigraphic succession without similarities with well-known units of the Iberian Massif (Romão *et al*., 2013; 2014; Moreira *et al*., 2016a; 2016b). The strong disparities with the adjacent Central Iberian (CIZ) and Ossa-Morena (OMZ) Zones successions, indicates a distinct geodynamical evolution, at least, during the Palaeozoic times.

Along the western block of the PTFSZ, similar HT tectono-metamorphic units were also described in Porto-Espinho-Albergaria-a-Velha and Berlengas Islands sectors (Fig. 1C and 1D) (Chaminé, 2000; Chaminé *et al*., 2003a; Pereira *et al*., 2007; Ribeiro *et al*., 2009; 2013; Bento dos Santos *et al*., 2010; in press). In Porto-Espinho-Albergaria-a-Velha as well as in Coimbra sectors, low-grade tectono-metamorphic units are also found, which were correlated with Proterozoic units of OMZ ("Série Negra" (Black Series); Chaminé *et al*., 2003a; 2003b; Ferreira Soares *et al*., 2005). Nevertheless some authors considerer a distinct geodynamic significance, origin and age to this unit (Chaminé *et al*., 2003b; Machado *et al*., 2008; 2011).

This paper presents a general overview of the tectonostratigraphic organization for the westernmost sectors of PTFSZ, describing and discussing the distinctive features that lead to individualize the Finisterra as a new tectonostratigraphic Terrane in the Iberian Massif.

VII.1.2. The Abrantes-Tomar Finisterra Sector

VII.1.2.1. Tectonostratigraphy

Near the PTFSZ, in the westernmost sectors of Abrantes-Tomar region, a N-S to NNW-SSE elongate HT tectonostratigraphic succession was defined (Fig. 2). These metamorphic units, with a probable Neoproterozoic-Lower Cambrian age, displays a complex polyphasic deformation and cannot be correlated with neighbouring variscan zones (i.e. OMZ and CIZ). Recent geological mapping (Fig. 2) enhanced the knowledge of these units.

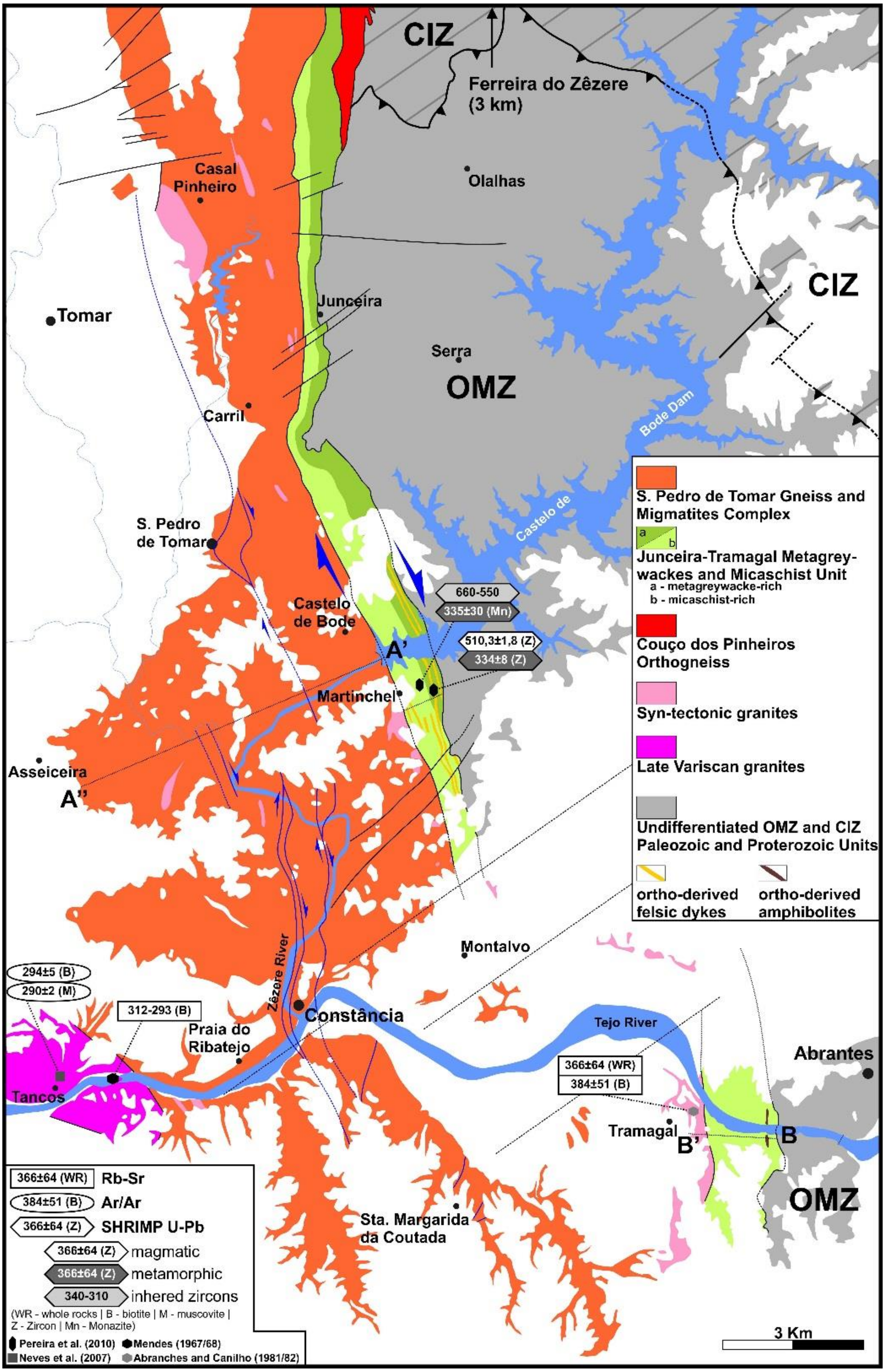


Figure 2 – Simplified geological map of the Abrantes-Tomar sector (marked cross section on figure 4B).

VII.1.2.1.1. S. Pedro de Tomar Complex

The western limit of this basal tectonostratigraphic unit is covered by the Meso-Cenozoic formations, while to the East it contacts with the Junceira-Tramagal Metagreywackes and Micaschists Unit, through a shear zone. This complex is characterized by medium to fine-grained strongly deformed gneisses, interlayered with micaschists and migmatites (Fig. 3A and 3B).

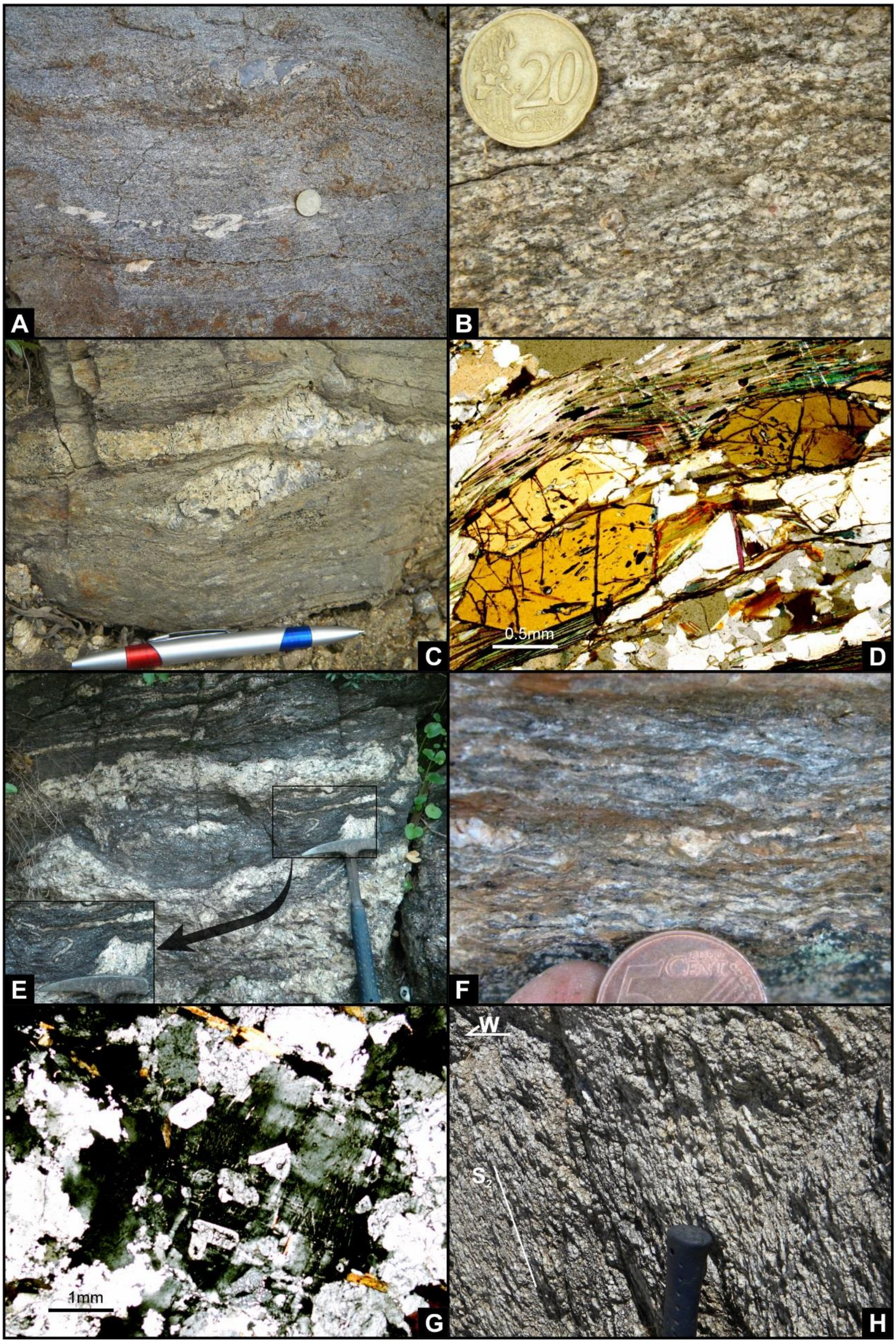
Two distinct macroscopic facies could be recognized in gneiss lithotypes. The most common are the para-derived gneisses dominated by a mineral paragenesis of plagioclase + quartz + biotite + sillimanite + muscovite + opaque minerals ± garnet ± cordierite. The other facies consists of orto-gneiss bodies, some of them less deformed and clearly related to the anatexis and melting of para-derived rocks (Fig. 3B). Its mineral paragenesis includes plagioclase + quartz + K-feldspar + biotite + sillimanite + muscovite ± cordierite. Plagioclase porphyroblasts are sometimes developed in the orthogneisses. In thin-section, the feldspar *s.l.* crystals has undulose extinction and some recrystallization, which coupled with the presence of sillimanite suggests temperatures around 500º-600ºC (Passchier and Trouw, 2005; Bucher and Graper, 2011; Singh and Gururajan, 2011). In both gneisses, the presence of chlorite intergrowth with silimanite crystals is common, being related to retrograde metamorphism.

The migmatites, abundant at south of S. Pedro de Tomar village, present a banded structure with alternations of melanosomes and thinner leucosome layers. The leucosomes are mostly composed of quartz + feldspar *s.l.* + muscovite, while the melanosome presents similar paragenesis with para-derived gneisses. Indeed, some gneisses are in fact migmatite lithotypes strongly deformation by the high-strain dextral shearing which develops a gneissic foliation (Fig. 3A). This suggest that the HT metamorphism and related migmatization are previous to the dextral shearing. Sometimes, quartz-mylonites are present in the vicinity of the Junceira-Tramagal Metagreywackes and Micaschists contact. These mylonitic bands are parallel to the main foliation produced by the regional dextral shearing.

There are no available geochronological data that constrain the photolith and metamorphic ages of these gneisses and migmatites. However, the para-derived lithotypes are considered Neoproterozoic to Lower Palaeozoic in age (see discussion below).

VII.1.2.1.2. Junceira-Tramagal Metagreywackes and Micaschist Unit

This unit crops out along a narrow and elongated 40 km N-S to NNW-SSE band, from Ferreira do Zêzere, at north, to Tramagal, at south.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1mm
G
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Figure 3 – Main features of the lithotypes of the Abrantes-Tomar tectono-stratigraphic units:

A – Deformed migmatites from S. Pedro de Tomar Complex;

B – (Ortho-)Gneisse from S. Pedro de Tomar Complex;

C – Felsic coarse-grained deformed ortho-derived dyke (pegmatite) intruded in the Junceira-Tramagal Unit;

D – Prismatic staurollite + biotite + muscovite + quartz + opaque minerals paragenesis contained in micaschiscts of the Junceira-Tramagal Unit;

E – Evidences of migmatization in the Junceira-Tramagal Unit, showing syn-migmatization folds;

F – Couço dos Pinheiros Orthogneiss, showing the highly deformed fabric, with feldspar sigmoidal crystal;

G – Deformed plagioclase crystal contained in the Tramagal Granite;

H – Foliated structure emphasized by biotite alignment in the Casal Pinheiro Granite.

In the Martinchel-Ferreira do Zêzere section, the succession displays an apparent thickness of circa 100 m and seems to correspond to a "coarsening upward" sequence expressed by the increase in the thick and number of metagreywackes beds towards the top of the unit. In this section, two different members could be individualized in this unit. The Lower Member is mainly composed of micaschists, sometimes interlayered with garnet-micaschists and subordinate thin metagreywacke beds, cropping out in the western domains and becomes thinner to northwards. The Upper Member is mainly composed of medium to fine-grained centimetric to metric metagreywackes and metaquartzwackes layers, interbedded with thin layers of micaschists and, sometimes, garnet-micaschists. In the Martinchel region quartz-feldspatic orthogneisses are sometimes recognized and could be considered the result of deformation and metamorphism in felsic-rich rocks (pegmatitic dykes?; Fig. 3C). Some of these bodies present mylonitic textures with quartz ribbons due to intense dynamic recrystallization of quartz. Some late less deformed granitic dykes are also described.

In Tramagal section, this unit appears to be thicker, being mostly composed of micaschists, with subordinate metaquartzwackes, metagreywackes and black schists. The coarse-grained lithotypes predominate in the westernmost domains of this section, where the centimetric metaquartzwackes beds could reach decimetric thickness. The micaschists paragenesis is dominated by biotite + muscovite + quartz + plagioclase + opaque minerals ± K-feldspar. A distinctive feature is the presence of millimetric to centimetric garnet and staurolite porphyroblasts, which could represents the metamorphic peak conditions (Fig. 3D). Such paragenesis are ascribable to amphibolite facies conditions in staurolite zone. The edges of the staurolite porphyroblasts are sometimes corroded, being associated to biotite, possibly related with the retrograde metamorphic episode. Also in Tramagal section, evidences of local migmatization is found to western domains (Fig. 3E), near the contact with Tramagal Granite. Contained in Junceira-Tramagal Metagreywackes and Micaschist Unit, some ortho-derived dykes with a monotonous paragenesis of green amphibole + plagioclase + opaque minerals ± quartz, typical of amphibolite facies conditions, are described.

Several ages have been recently established for the Junceira-Tramagal unit in Martinchel sector (Fig. 2; Pereira *et al*., 2010). The para-derived lithotypes are considered Neoproterozoic, based on the Ediacarian age (550-660 Ma) of the most recent group of inhered zircons population. The quartz-feldspatic orthogneiss cutting previous metasedimentary lithotypes shows a Lower Cambrian age (510.3±2.0 Ma; LA-ICP-MS, U-Pb in zircons). Monazites from the para-derived lithotypes displays Carboniferous metamorphic age (ca. 335 Ma), which is similar to the ones obtained in the quartz-feldspar gneiss (334.1±8.0 Ma for the discordia age anchored at intrusion age). A late granitic dyke intruded in pelitic lithotypes provide a 318.7±1.2 Ma for the emplacement age, which could be interpret as related to the last pulses of HT metamorphic episode.

The inhered zircon populations show the presence of Neoproterozoic (690-780, 820-840 and 920 Ma), Mesoproterozoic (1050-1150 and ca. 1300 Ma), Paleoproterozoic (ca. 1650 and 1880-2200 Ma) and Paleoproterozoic-Archean (2350-2800 and ca. 2900 Ma) populations (Pereira *et al*., 2010). Some of the Mesoproterozoic zircons are slightly discordant (Fig. 4A; Pereira *et al*., 2010).

However, Silurian (ca. 440 Ma) and some Devonian-Carboniferous (350-400 Ma) zircons are also obtained in the Carboniferous granitic dyke intruded in the Junceira-Tramagal Unit (Fig. 4A; Pereira *et al*., 2010). Their origin and significance are poorly constrained: the Devonian-Carboniferous ages could represent relics of a metamorphic zircon overgrowth event (Devonian metamorphism?) and the Silurian age could represent an inhered zircon. However, this Silurian age have a high error (ca. ±40 Ma) and thus a Devonian age could not be excluded.

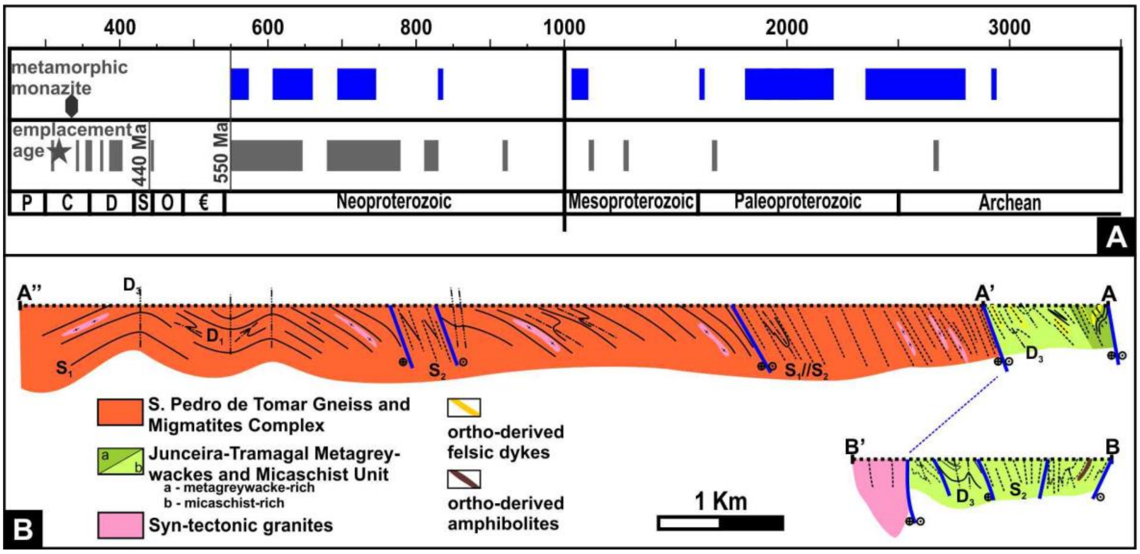


Figure 4 – Finisterra Terrane features in Tomar-Abrantes sector:

A – Zircon patterns of samples collected in Junceira-Tramagal Unit ($^{207}Pb/^{206}Pb$ ages obtained by Pereira *et al*., 2010);

B – Simplified geological cross sections.

VII.1.2.1.3. Couço dos Pinheiros Orthogneiss

This unit corresponds to a strongly stretched N-S sigmoidal ortogneiss body, with typical gneissic texture composed of millimetric felsic-rich layers (quartz and feldspars *s.l.*) and iron-magnesium silicates rich ones. The presence of sigma shaped K-feldspar porphyroblasts and strongly stretched quartz ribbons have been interpreted as the result of intense ductile non-coaxial deformation (Fig. 3F). These gneisses are intruded by less deformed felsic coarse-grained layers.

The origin and age of the Couço dos Pinheiros orthogneisses are unknown. However, its petrographic and structural similarities with those observed in HT tectono-metamorphic units, mainly the S.Pedro de Tomar Complex, suggest it is a (early- or) pre- orogenic magmatic body, with possible Neoproterozoic-Lower Cambrian age. However, an Ordovician to Devonian age for emplacement of this orthogneiss could not be excluded.

VII.1.2.1.4. Syn-orogenic Variscan Granites

Some N-S elongated granitic bodies intrude the Abrantes-Tomar HT tectonostratigraphic units. The most representatives are the Tramagal and Casal Pinheiro two mica granites, being mainly composed of quartz + K-feldspar + plagioclase (with abundant perthites) + muscovite + biotite ± tourmaline, sometimes with K-feldspar phenocryst. The presence of in the Tramagal Granite indicate an anatectic origin, probably related with the HT metamorphism.

Both granites show evidences of deformation coeval of their crystallization, developing preferential elongation of mica crystals, sometimes leading to the development of a weak foliation. This plastic deformation is also visible in micro-anisotropies like undulose extinction and kinking in feldspar crystals and quartz sub-grains recrystallization (Fig. 3G). In the Casal dos Pinheiros Granite, the deformation also give rise to biotite alignment (Fig. 3H) and sigmoidal shape of feldspar crystals, compatible with dextral shearing along foliation planes.

There is no geochronological data allowing a detailed constraining of the emplacement of these granites. However, using structural data (see 2.2 section), a Carboniferous age could be assumed to the emplacement of these felsic bodies. This assumption is compatible with the geochronological data obtained for the Tramagal Granite, although its error margin is too high to confirm (Fig. 2; Rb/Sr method in whole rock 366±64 Ma and in biotite 384±51 Ma; Abranches and Canilho, 1981/82).

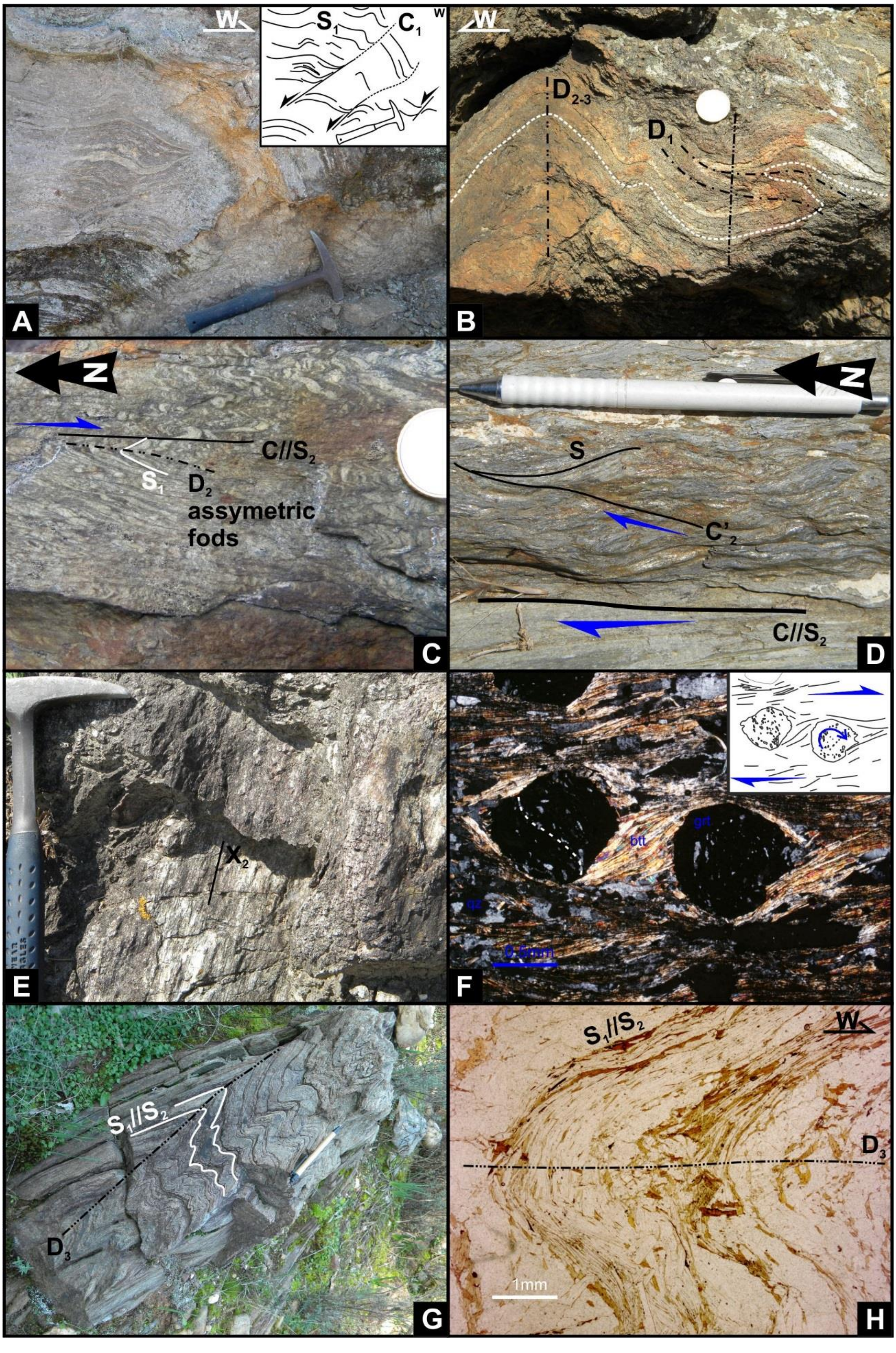
VII.1.2.2. General Structure

The Abrantes-Tomar tectonostratigraphic units, as well the ortho-derived bodies, are elongated in N-S to NNW-SSE direction, following the general trend of PTFSZ (Fig. 2). The N-S trend predominate from Ferreira do Zêzere to Martinchel, while in southward section, between Martinchel and Tramagal, the general trend present a SE deflection becoming NNW-SSE. This inflection seems to be a local effect, related to the interaction between PTFSZ and TBCSZ. The presence in the Abrantes-Tomar region of a kilometric sheath fold related to the WNW termination of TBCSZ (Ribeiro, *et al.*, 2009; Moreira *et al.*, 2011; 2013; Moreira, 2012) could generate the inflection of PTFSZ, which is compatible with the early activity of the PTFSZ, at least since the beginning of compressional stages of Variscan orogeny, as previously proposed (Dias and Ribeiro, 1993; Ribeiro *et al*., 2007; 2013). The structural data of Junceira-Tramagal Unit and S. Pedro de Tomar Complex show, at least, the presence of three deformation episodes.

The gneissic foliation in S. Pedro de Tomar Complex is the most common structure related to the first deformation episode (D_1), being often associated with quartz-veins. The geometry and kinematics of the D_1 is poorly constrained, mainly due to the overprinting of younger deformations. As the intensity of the main D_2 deformation episode increase eastwards towards the PTFSZ, the D_1 structures are usually transposed near this main shear zone (Fig. 4B). However, in the west domains of S. Pedro de Tomar Complex, far from PTFSZ where the D_2 is less pervasive, a low dipping S_1 gneissic foliation is developed (Fig. 4B). The S_1 foliation is sometimes associated to asymmetric ptygmatic recumbent folds with sub-horizontal to low dipping hinges and, although highly dispersed, a NNW-SSE trend (Fig. 5A). The sense of transport is debatable, nevertheless the relation between D_1 structures seems to show transport with top-to-West.

Also in the Junceira-Tramagal Unit, some D_1 recumbent folds, with a westward geometric vergence are found (Fig. 5B). These folds, which are refolded by D_2-D_3 event, are associated to a S_1 HT foliation, which is compatible with the evidences of migmatization during the D_1 tectonic event (Fig. 3E). The leucosome and melanosomes are clearly deformed by D_2 dextral shearing episode being synchronous with the D_1 event. Although, the D_1 geometry is poorly constrained, being only preserved where the D_2 event is not pervasive.

The D_2 event develops a pervasive ductile S_2, parallel to the general N-S to NNW-SSE PTFSZ trend. Usually S_2 strongly dips to East (Fig. 4B), being associated to sub-horizontal to low SE plunging D_2 stretching mineral lineation. In the vicinity of PTFSZ, the S_2 foliation transposes previous structures ($S_1//S_2$), becoming less penetrative towards West. This foliation is associated to a non-coaxial dextral regime induced by PTFSZ. In the more deformed sectors, S_2 behave as C planes (Fig. 5C) with coeval development of several structures emphasizing the dextral kinematics: C-S and C-C' structures (Fig. 5D) or asymmetric folds with steeply plunged hinges and sigmoidal shape structures. Sometimes, the S_2 foliation is associated with an intense mylonitization, also showing dextral kinematics. The C' structures displays NNE-SSW to NE-SW strike, dipping to E and contain a down-dip to SE plunging oblique stretching lineation, showing top-to-SE criteria (Fig. 5E). This general dextral kinematic with a slightly oblique transport with top to East is compatible with the presence of the high temperature unit in the westernmost domain. It should be stressed that, not only the boundaries between all the Abrantes-Tomar tectonostratigraphic units but also the contact between the Finisterra and the Iberian Autochthon Terrane (*i.e.* the PTFSZ), are always D_2 1^{st} order dextral shear zones.



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S1
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assymetric
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S
C'2
C//S2
D
X2
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btt
grt
qz
0.5mm
F
S1//S2
D3
G
S1//S2
W
D3
1mm
H

Figure 5 – Main structural features of Abrantes-Tomar sector:

A – Syn-metamorphic D_1 recumbent folds and shear-bands in the S. Pedro de Tomar Complex;

B – Refolded D_1 recumbent folds in Junceira-Tramagal Unit;

C – D_2 shear bands affecting the previous D_1 HT foliation generated;

D – C-C' fabric related to dextral shearing during D_2 affecting the Junceira-Tramagal Unit;

E – Macroscale C' band showing down-dip stretching lineation associated to mylonitic S_2;

F – Poikilitic garnet, showing dextral synthetic spinning;

G – D_3 low grade overturned tight folds in the Junceira-Tamagal Unit;

H – Biotite foliation in a quartz-rich lithotype folded by low-grade D_3 folds.

The D_2 deformation episode is associated to a HT and high strain fabric. In Junceira-Tramagal Unit, the garnet grow and have a synthetic rotation during the D_2 dextral shearing, presenting generally poikilitic structures overlapping a previous foliation (S_1; Fig. 5F).

The Tramagal and Casal Pinheiro granites have also been deformed by D_2. They present a poorly developed N-S to NNW-SSE S_2 cleavage demarked by biotite alignment, generally dipping to E, which often bound sigmoidal shape structures compatible with dextral shear (Fig. 3H). This geometry and kinematics of S_2 shows similarities with D_2 features previously described in both HT tectonostratigraphic units. Evidences of hot-plastic deformation of feldspar crystal seems to show that the deformation is active during granite emplacement. Thus, it is considered that the emplacement of these granite bodies are contemporaneous from D_2 deformation episode.

The S_2 foliation is folded by the third deformation phase (D_3). Near the main Eastern D_2 shear zones, the D_3 folds are tight with a West facing related to moderately to steeply inclined axial surfaces (Fig. 5G and 5H). Towards the West domains, they progressively open, while the axial surfaces become subvertical (Fig. 4B). Whatever their geometry, the D_3 folds are always characterized by low dipping hinges with a N to NNW plunges, associated to poorly developed low-grade axial planar cleavage, sometimes crenulation cleavage. The influence of the D_2 main shear zones in the D_3 folds pattern, show that they also control the D_3 folding.

Concerning the deformation age, all these episodes are interpreted as Variscan. The D_1 event, related to the HT gneissic foliation, could be Late Silurian to Devonian in age (ca. 420?-350 Ma), being associated to the zircon growth emphasized in the Junceira-Tramagal Unit. This early deformation was followed by two Carboniferous episodes: the HT and high-strain D_2 event could be Mississippian in age (ca. 340-320 Ma), being associated to dextral shearing of PTFSZ, which almost completely transposes the D_1 structures, and the D_3 event Pennsylvanian-Early Permian in age (ca. 310-295 Ma), showing low-grade conditions which could be related to the exhumation and cooling related to the retrogressed metamorphism.

The proposal chronological relation was based in Pereira *et al*. (2010) ages that show the presence of metamorphic ages around 335 Ma associated with high strain deformation. The D_2 high strain deformation do not affect prevasively the granitic dyke intruded in pelitic rocks (318.7±1.2 Ma), constraining the maximum age of high strain deformation. Finally, the Tancos granite (Fig. 2), a post-tectonic porphyritic two-mica granite not affected by any of the previous deformation events, constrain the maximum ages of D_3 event; Ar/Ar geochronological data in micas (Neves *et al*., 2007) shows an Early Permian age to its emplacement (Biotite - 294±5Ma and Muscovite – 290±2 Ma). This data which is compatible with ages obtained by Mendes (1967/68) with Rb/Sr method in biotites (ca. 312-293 Ma).

Finally, it should be noted that all the study area is affected by intense subvertical fracturing with two main trends: N35ºE and N70-80ºE. The fracture pattern could be related either to the Late Variscan event (Dias *et al*., 2017), or with the Meso-Cenozoic processes associated to the opening of Lusitanian Basin and consequently the Atlantic Ocean.

VII.1.3. The North and Central Finisterra Domains

The northern continuity of the Abrantes-Tomar Finisterra domain is disrupted due to the complex interaction with the Meso-Cenozoic formations strongly deformed by the PTFSA fracture system. Two main independent domains are found in the vicinity of this first order shear zone: the Porto-Espinho-Albergaria-a-Velha and the Coimbra sectors (Fig. 6). Both are characterized by high-temperature metamorphism and/or low-grade tectonostratigraphic units not possible to correlate either to the CIZ or to the OMZ.

VII.1.3.1. Porto-Espinho-Albergaria-a-Velha Sector

West of the PTFSZ four pre-Mesozoic tectonostratigraphic units have been defined between Espinho and Albergaria-a-Velha (Fig. 6; Chaminé, 2000, Chaminé *et al*., 2003a, Pereira *et al*., 2007; Ribeiro *et al*., 2013). From the bottom to the top:

- Lourosa Unit – Composed of gneisses, migmatites, micaschists and garnet-micaschists, this unit could be divided in two members: the lower member is mostly composed of migmatites, ortho- and paragneisses while the upper member is dominated by the biotite-micaschists, sometimes with garnet. Both members present amphibolites and amphibolic schists, withgeochemical affinities similar to within-plate basalts to MORB, being derived from mafic (and ultra-mafic – Engenho Novo Olivine amphibolites) rocks (Montenegro Andrade, 1977; Aires and Noronha, 2010; Silva, 2007). This Unit has been correlated with the Pindelo Unit previously defined in the same sector (Chaminé, 2000; Chaminé *et al*., 2003a).
- Espinho Unit – West of the previous unit outcrops a narrow band of staurolite-garnet-biotite micaschists, sometimes with mylonitic garnet-quartzites and, occasionally, impure quartzites. The metamorphic paragenesis shows the presence of two distinct HT metamorphic events: the first one reaches the sillimanite zone and second one the staurolite zone (Chaminé, 2000; Fernandez *et al*., 2003).
- Arada Unit – A low-grade metamorphic unit composed of a monotonous sequence of black to green phyllites, derived mafic rocks with tholeiitic geochemical features (Silva, 2007) and black quartzites. The similarities and correlation between this succession and

the Ediacarian “Série Negra” of OMZ was initially proposed (Chaminé, 2000; Pereira *et al*., 2007).

- Albergaria Unit – Within Arada Unit, imbricate very low-grade rocks, mainly composed of black shales, with Frasnian-Serpukhovian age were described (Chaminé *et al*., 2003b; Machado *et al*., 2011).

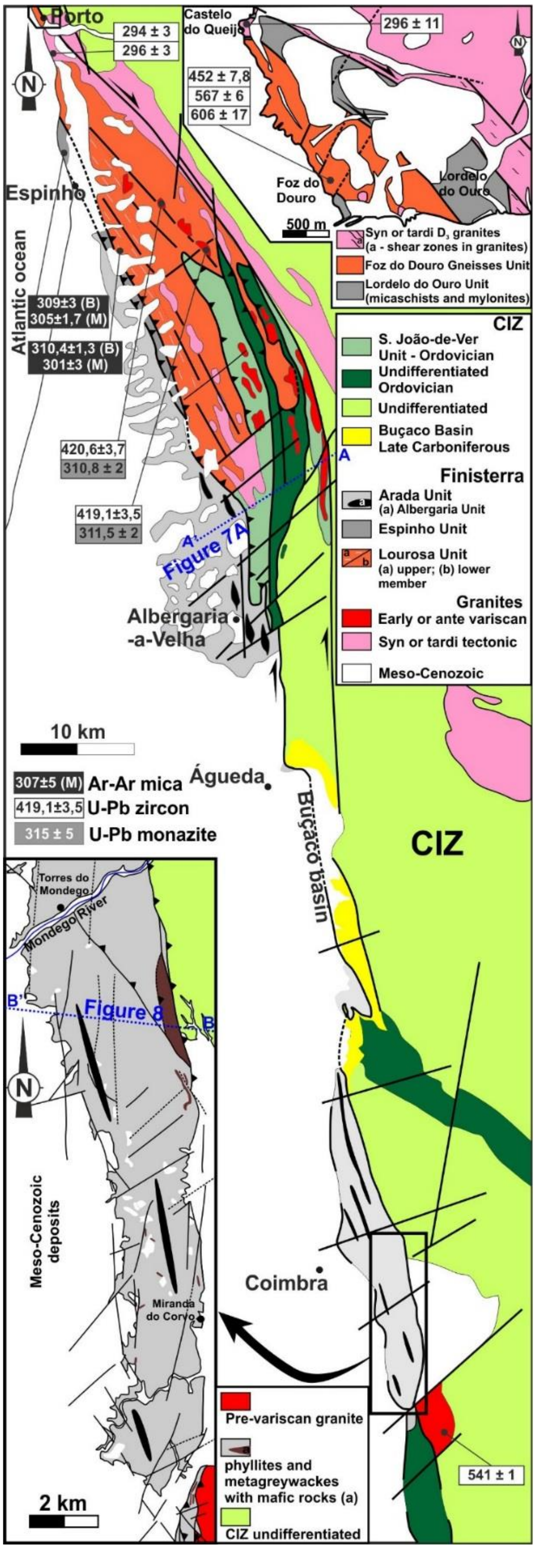

Figure 6 – Simplified geological map and geochronological data for the Porto-Coimbra Finisterra sectors (references in the text; adapted from Chaminé *et al*., 2003a; Ferreira Soares *et al*., 2005; Pereira *et al*., 2007; LNEG, 2010; Machado *et al*., 2011; Dinis *et al*., 2012).

Although the Arada, Espinho and Lourosa units have been usually considered as Neoproterozoic (Beetsma, 1995; Chaminé, 2000; Chaminé *et al*., 2003a; Ribeiro *et al*., 2009; 2013), such assumption has been debated in recent studies. The Upper Devonian-Carboniferous imbrications within the Arada Unit show, that at least in part, this unit is more recent (Albergaria Unit; Chaminé *et al*., 2003b; Machado *et al*., 2011). Moreover, the presence of acritarch assemblages in Albergaria Unit with affinities with the Late Devonian Laurassia marine acritarchs, contrasts with the Late Devonian assemblages from South Portuguese Zone which show clear affinities with North Gondwana margin (Machado *et al*., 2008).

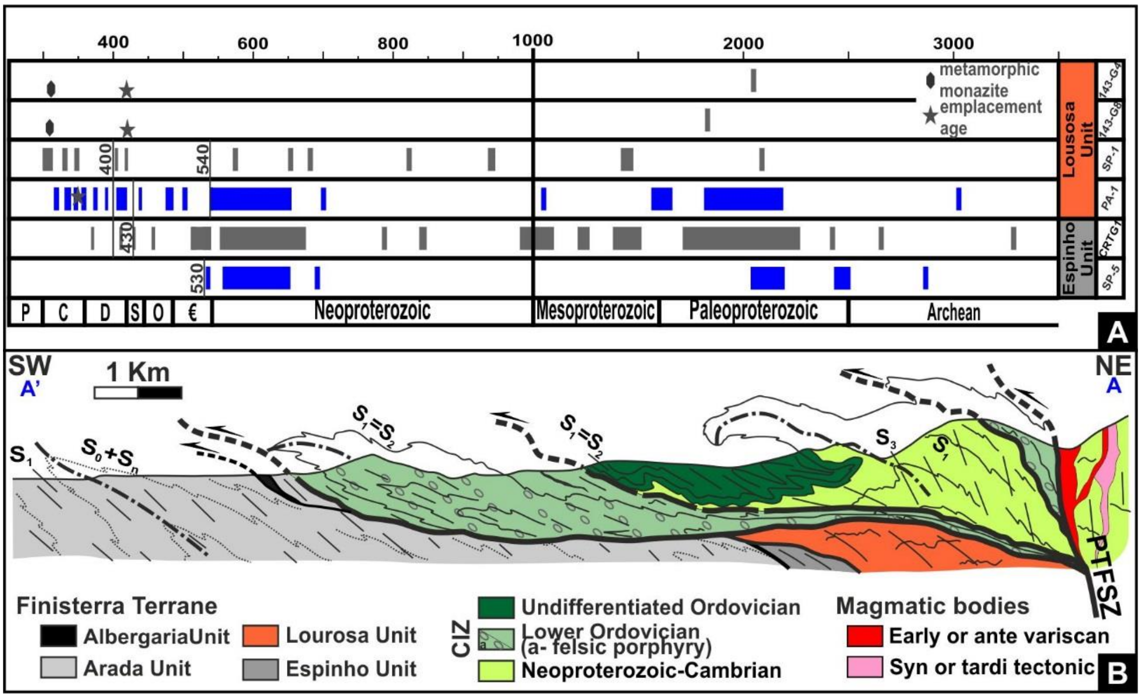

Figure 7 – Finisterra Terrane features in Porto-Espinho-Albergaria sector:

A – Zircon patterns of samples collected in Lourosa and Espinho Units (geochronological data from Chaminé *et al*., 1998; Almeida, 2013; Almeida *et al*., 2014);

B – Simplified geological cross section (adapted from Pereira *et al*., 2007).

Concerning the age of the Espinho and Lourosa Units, although the most representative detrital zircon population are Lower Cambrian to Neoproterozoic (510-680 Ma in the Espinho Unit and 540-650 Ma in the Lourosa Unit; Almeida, 2013; Almeida *et al*., 2014), they also present Upper Cambrian-Ordovician to Silurian zircons (Fig. 7A; Almeida, 2013; Almeida *et al*., 2014). Thus, at least part of these tectonostratigraphic units are Palaeozoic (Silurian?), once some of these zircons presents detrital morphology (Almeida *et al*., 2014). As expected, older inhered zircon populations are found in both units (Fig. 7A; Almeida, 2013):

- Neoproterozoic (ca. 690, ca. 790 and ca 840 Ma), Mesoproterozoic (980-1250 and 1400-1500 Ma), Paleoproterozoic (1700-2250 and 2400-2500 Ma) and Archean (ca. 2650, ca. 2850 and ca. 3280 Ma) in the Espinho Unit;
- Neoproterozoic (ca. 680, ca. 700, ca. 820 and ca. 940 Ma), Mesoproterozoic (ca. 1050 and 1400-1500 Ma), Paleoproterozoic (1550-1650 and 1800-2200 Ma) and Archean (ca. 3050 Ma) in the Lourosa Unit.

The zircon patterns from Lourosa and Espinho Units are not equal for all samples (Almeida, 2013; Almeida *et al*., 2014). Some samples do not present Mesoproterozoic populations, while in other samples the Mesoproterozoic population is representative (Fig. 7A).

Upper Silurian-Devonian overgrowths in inhered zircon are found in both units (370-410 Ma in the Espinho Unit and 350-420 Ma in the Lourosa Unit; Almeida, 2013; Almeida *et al*., 2014). This suggests the presence of a metamorphic and/or magmatic event with Silurian-Devonian age. The presence of Silurian felsic magmatism in Lourosa Unit (Lourosela and Souto Redondo Orthogneisses; ca. 420 Ma; Chaminé *et al*., 1998), is compatible with such interpretation. Also in this unit, a Lower Devonian age was obtained for the protolith of a mafic amphibolite (392±2 Ma; LA-ICP-MS in zircon; Almeida, 2013; Almeida *et al*., 2014). However, older concordant ages are Silurian (ca. 420-430 Ma) and which could represent the protolith age, being the Devonian ages obtained in some zircons (ca. 390-350 Ma) resulted from the metamorphic and/or magmatic events previously referred.

However, the amphibolites age is not consensual. Indeed, a Mesoproterozoic Sm-Nd model age (TDM; ca. 1050 Ma; Noronha and Leterrier, 2000) was obtained to the mantle protolith of similar amphibolites in the Porto region. This Precambrian age were considered similar to the emplacement of mafic magmatism with tholeiitic MORB affinities in the Foz do Douro gneiss unit (Noronha and Leterrier, 1995; 2000).

These amphibolites are contained in the Foz do Douro Gneissic Unit which is equivalent of the Lourosa unit (Fig. 6; Noronha and Leterrier, 1995; 2000; Chaminé *et al*. 2003a; Ribeiro *et al*., 2009). This unit is mainly composed of orthogneisses, sometimes with mylonitic bands, paragneisses and amphibolites. The orthogneisses have tonalitic to granitic composition, being subdivided in four distinct lithotypes (Noronha and Leterrier, 2000): (i) biotite gneisses, (ii) felsic gneisses, (iii) augen-gneisses and (iv) highly deformed felsic gneisses. The eastern boundary of the Foz do Douro Gneissic Unit is a major dextral shear zone, considered the contact between the CIZ and the Finisterra Terrane (Ribeiro *et al*., 2009). However, this contact is done with is a narrow band of micaschists and quartz-tectonites (Lordelo do Ouro Unit; Chaminé *et al*., 2003a). The strong similarities of this unit with the micaschists interlayered in Foz do Douro Gneisses Unit, seems to indicate that the Lordelo do Ouro Unit could also be part of Finisterra Terrane.

The age of Foz do Douro Gneisses Unit is debatable (Fig. 6). Ediacarian ages are sometimes assumed (Ribeiro *et al*., 2009) based in geochronological data from the biotitic orthogneisses (567± 6 Ma) and the augen felsic gneisses (606 ± 17 Ma; U/Pb – isotope dilution in zircon; Noronha and Leterrier, 2000). However, an Ordovician age have been recently obtained to the protolith of biotitic orthogneisses from this unit (452±8 Ma; SHRIMP U-Pb in zircon; Sousa *et al*., 2014). Similar Ordovician age was also obtained in a granite orthogneiss emplaced in Lourosa Unit (459±7 Ma, U/Pb LA-ICP-MS in zircon; Almeida, 2013).

Regarding the metamorphism in Porto-Espinho-Albergaria-a-Velha Sector two distinct Carboniferous clusters were identified:

- ca. 340-325 Ma, evidenced by zircon overgrowths (Almeida *et al*., 2014);
- ca. 310-300 Ma, obtained by Ar-Ar in micas and amphibole (Acciaioli *et al*., 2003; Munhá *et al*., 2008; Gutiérrez-Alonso *et al*., 2015), but also U-Pb in monazites (Chaminé *et al*., 1998; Munhá *et al*., 2008) and Rb-Sr using whole rock-feldspar-biotite-muscovite isochron (Santos *et al*., 2012).

Concerning the structure, all units have a polyphased deformation (Chaminé, 2000; Ribeiro *et al*., 2013) giving rise to a N-S general trend, deflecting to NNW-SSE in the northernmost domains (Fig. 6). Not only the boundaries between all the units are major thrusts, but also both Lourosa and Arada Units are thrusted by CIZ Neoproterozoic-Ordovician successions always with top to SW transport (Chaminé, 2000; Pereira *et al*., 1880; 2007; Ribeiro *et al*., 1880; 2013; Fig. 7B).

This sector is characterized by two ductile deformation episodes (D_1 and D_2), followed by a third brittle one (D_3). During D_1 major recumbent folds with low dipping N-S hinges, subparallel to a X_1 stretching mineral lineation, and a western vergence were developed (Pereira *et al*., 1980; Ribeiro *et al*. 1980; 1995; 2013). These folds are coeval of a penetrative axial planar penetrative foliation (S_1) cleavage generally transposed by the second foliation (S_2). The D_1 event is probably related to an early HT fabric with flattened garnet and sillimanite observed in the Espinho Unit (P = 4-5 kbar; T=700±50ºC; Fernandez *et al*., 2003).

The D_2 deformation event develops a penetrative NW-SE to N-S axial planar S_2 cleavage associated to subvertical folds with low dipping SE-plunge hinges (Ribeiro *et al*., 1980; Chaminé, 2000). The N-S trends are pervasive near the PTFSZ, deflecting to NW-SE in the western domains far from this 1st order shear zone (Ribeiro *et al*., 1980) The Espinho and Lourosa Units have been strongly deformed during D_2, giving rise to the pervasive development of S_2. This foliation, often with mylonitic and gneissic features, is associated to subhorizontal to low plunging to SE stretching mineral lineation, subparallel to fold hinges. The D_2 deformation was the result of a non-coaxial dextral shear regime, often completely transposing D_1 structures. In the Arada Unit, the S_2 foliation presents low-grade features, although also usually to a high strain deformation pattern.

The non-coaxial dextral D_2 movement along the N-S sector of PTFSZ induces space problems when approaching its NNW-SSE segment between Albergaria-a-Velha and Porto (Fig. 6), that behaved has a restraining bend (Ribeiro *et al*., 2013). Such geometrical constrain give rise to an imbricated thrust complex with a SW facing, giving rise to the superposition of the Arada and Lourosa units by the CIZ sequences (Fig. 6 and 7B; Pereira *et al*., 2007; Ribeiro *et al*., 2013). This tangential deformation roots in the lithospheric subvertical or steeply dipping to East PTFSZ, which bound two distinct basement blocks (Ribeiro *et al*., 1995; 2013).

In the Espinho Unit the D_2 event generates a second HT metamorphic paragenesis with garnet overgrowth and staurolite porphyroblasts (P=3-6 kbar; T=600±30ºC), which partially reset the first HT paragenesis in sillimanite zone (P=4-5 kbar; T=685-750ºC Fernandez *et al*., 2003). Migmatites from the Lourosa Unit also show a HT metamorphic event during D_2 (P = 8 ± 0,7 kbar; T=730±25ºC), being characterized by garnet + sillimanite + K-feldspar + biotite ± muscovite (+ melt) assemblage (Munhá *et al*., 2008).

Although D_1 and D_2 are usually consider two distinct deformation events westwards of the PTFSA, they could also be considered the result of a progressive deformation (Ribeiro *et al*., 1995; Chaminé, 2000).

The third and last deformation event (D_3) is characterized by brittle structures with cataclasites related, not only to N-S shear zones, but also to the reactivation of top-to-SW thrusts often underlined by fault-gouge (Chaminé, 2000; Ribeiro *et al*., 2013).

The D_1 is considered previous to the sedimentation of Frasnian-Serpukhovian black shales while the D_2 must to be younger (Ribeiro *et al*., 2013). Therefore, previous mentioned geochronological data led to postulate a Late Silurian-Devonian age to the HT D_1 event (ca. 420-350 Ma). Concerning the D_2 HT metamorphism, it could be related to the zircon overgrowth during Mississippian (ca. 340-325 Ma). The Ar-Ar geochronological data (Acciaioli *et al*., 2003; Munhá *et al*., 2008; Gutiérrez-Alonso *et al*., 2015) could indicate that the exhumation of the HT metamorphic rocks extending up until Pennsylvanian times (310-300 Ma). The younger age is constrained by the emplacement of Castelo do Queijo and Madalena-Lavadores post-tectonic Granites (ca. 295-290 Ma; Martins *et al*., 2011; 2014). As N-S D_3 brittle structures are observed in the Madalena-Lavadores granite (Ribeiro *et al*., 2015). This younger deformation event could be considered as contemporaneous of the emplacement of post orogenic magmatism (ca. 300-290 Ma).

VII.1.3.2. Coimbra Sector

The western block of PTFSZ in Coimbra sector is characterized by one low-grade unit strongly similar to previously described to Porto-Espinho-Albergaria-a-Velha sector (Arada Unit; Chaminé *et al*., 2003a; Machado *et al*., 2008; 2011). It is a siliciclastic sequence mainly composed of black phyllites and metagreywackes, with black quartzites and ortho-derived mafic rocks (Ferreira Soares *et al*., 2005). The lithological resemblances led to consider this unit equivalent of the OMZ "Série Negra" (Ferreira Soares *et al*., 2005), with a Neoproterozoic age (Ribeiro *et al*., 2013). However, also here are described the presence of very low-grade imbricate metasedimentary rocks (black shales, with less abundant siltstones and sandstones; Fig. 6 and 8) with Frasnian-Serpukhovian age (Chaminé *et al*., 2003b; Machado *et al*., 2011), which led to consider that, at least part of the sequence could be equivalent of the Albergaria Unit.

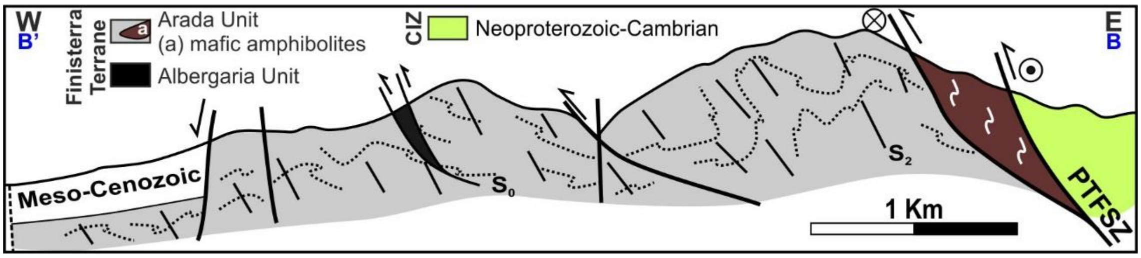

Figure 8 – Simplified geological cross section of Coimbra sectors (adapted from Ferreira Soares *et al*., 2005; Machado *et al*., 2011).

This low-grade unit has polyphased deformation, being trusted by the CIZ successions (Fig. 8; Ferreira Soares *et al*., 2005). The most important deformation are characterized by N-S dextral ductile-brittle to brittle shear zones related to the PTFSZ. These structures are subparallel to sub-vertical axial plane folds, where an axial planar cleavage is found (Ribeiro *et al*., 2013). The parallelism between both structures indicate an intense shear component related to the PTFSZ (Ribeiro *et al*., 2013). These intense deformation is superposed on an early deformation episode, poorly geometrically constrained, with recumbent folds (early Variscan or Cadomian?).

VII.1.4. The Berlengas Archipelago Finisterra Domain

The Berlengas Archipelago was considered a "suspect" terrane due the presence of deformed two mica granites, gneisses and micaschists West of Meso-Cenozoic Lusitanian Basin (Ribeiro *et al*., 1991). The similarities with the lithotypes of Tomar-Abrantes sector, which is the nearest Variscan basement, led to consider the Berlengas Islands part of the Western block of PTFSZ.

In the Farilhões and Forcadas islands outcrop an anatectic complex, mainly composed of gneisses, migmatites and micaschists, while the Berlengas, Estelas and Medas are mainly composed of a coarse-grained biotite pink granite (Fig. 1D; Valverde Vaquero *et al*., 2010a; 2010b; Bento dos Santos *et al*., in press).

The Berlenga granite is mainly composed of quartz + K-feldspar + plagioclase + biotite with no evidences of ductile deformation (Valverde Vaquero *et al*., 2010b). Initially considered as a Permian granite (280±15 Ma; $^{87}Rb/^{86}Sr$ de whole rock; Priem *et al*., 1965), recent data led to consider its emplacement as Upper Carboniferous (307,4±0,8 Ma; U/Pb concordant age in monazite and zircon, ID-TIMS; Valverde Vaquero *et al*., 2010a; 2010b). This granite was deformed by brittle-ductile N-S thrusts with low dips and an E vergence (Ribeiro, 2013).

The Farilhões metatexites are composed of quartz + plagioclase + k-feldspar + biotite ± sillimanite ± garnet ± muscovite, sometimes with granulite relics with plagioclase + quartz + biotite + amphibole ± garnet ± clinopyroxene ± ilmenite ± titanite paragenesis (Bento dos Santos *et al*., in press). Geothermobarometric studies show a prograde metamorphism reaching granulite facies conditions (P=8.6±1 kbar; T=915±50ºC; Bento dos Santos *et al*., 2010; in press), followed by retrograde metamorphism. The same authors consider these metatexistes genetically related to the diatexites composed of K-feldspar + plagioclase + sillimanite + biotite ± muscovite.

Monazites fractions from the diatexites provide Upper Devonian (377±1 Ma; U/Pb, ID-TIMS), which has been considered the age of the HT metamorphism in sillimanite zone that follows the granulite facies conditions (Valverde Vaquero *et al*., 2010a; 2010b; Bento dos Santos *et al*., in press). Some inhered zircons with Neoproterozoic ages (550-650 Ma; LA-ICP-MS, U-Pb in zircons; Bento dos Santos *et al*., in press) seems to indicate a siliciclastic Neoproterozoic protolith for the para-derived series.

VII.1.5. Distinctive Features of Finisterra Terrane

The individualization of the Finisterra Terrane from the adjacent Iberian one, must be supported by stratigraphic, tectonic, metamorphic and magmatic features, emphasizing a distinct geodynamical evolution (Coney *et al*., 1980). These terranes must to be representative at lithospheric scale, presenting distinct geodynamical evolution until its collage/accretion.

The presence of Ediacarian magmatism in Foz do Douro Gneiss Unit (Fig. 9; ca. 600-550 Ma; Noronha and Leterrier, 2000) and Ediacarian-Lower Cambrian inhered zircon populations in the Finisterra units (Pereira *et al*, 2010; Almeida, 2013; Almeida *et al*, 2014; Bento dos Santos, in press) are compatible with the Cadomian Orogeny, developed in North Gondwana margin (Linnemann *et al*., 2008). Thus, this could not be considered a distinctive feature, because in the

OMZ of the Iberian Terrane, an intense magmatism related to Cadomian event is also found (e.g. Salman, 2004; Simancas *et al*., 2004; Linnemann *et al*., 2008; Sánchez-Lorda *et al*., 2016). This should indicate a similar evolution of Finisterra and Iberian Terranes during Neoproterozoic times.

However, from Lower Cambrian until Lower Carboniferous times, several features indicate that the Finisterra was a terrane distinct from the Iberian one, being its contact outlined by the lithospheric scale PTFSZ.

The tectonostratigraphic units are clearly distinct from the Iberian Terrane successions. The Finisterra Terrane is characterized by the widespread occurrence of two HT tectonostratigraphic units: a basal gneiss and migmatite complex and an upper staurolite-garnet-micaschists unit. Between Espinho and Coimbra, there is also the development of a low-grade metamorphic unit (Arada Unit). Although the age of these HT tectonostratigraphic units are generally attributed to Proterozoic (e.g. Chaminé, 2000; Ribeiro *et al*., 2009), some Ordovician-Silurian ages are obtained in detrital zircons contained in these units (Fig. 9; Pereira *et al*., 2010; Almeida, 2013; Almeida *et al*., 2014), indicating that these units, or part of them, are possibly Silurian in age, also constraining the lower age of metamorphic event.

These HT tectonostratigraphic successions seem to be spatially (and generically?) associated to PTFSZ, presenting Silurian-Devonian zircon overgrowths (ca. 420-350 Ma; Fig. 9). The observed pre-Carboniferous HT foliation (Fernandez *et al*., 2003), could probably be related with such Silurian-Devonian event. This conclusions are compatible with the Berlengas data, where the HT early metamorphic event was attributed to Upper Devonian (ca. 380 Ma; Valverde Vaquero *et al*., 2010a; 2010b; Bento dos Santos *et al*., in press). In the Porto-Espinho-Albergaria-a-Velha sector this HT metamorphism is associated to Silurian felsic magmatism (Fig. 9; ca. 420 Ma; Chaminé *et al*., 1998). In the Iberian Terrane there are no evidences of a similar metamorphic and magmatic event, with exception to local HP metamorphic ages obtained in the southernmost domains of OMZ (ca. 370 Ma; Moita *et al*., 2005).

The low grade Arada Unit, is also considered as Neoproterozoic (Beetsma, 1995; Chaminé, 2000), but without recent geochronological constrains. In this unit occurs imbrications of very low-grade Frasnian-Serpukhovian black shales (Fig. 9; Chaminé *et al*., 2003b; Machado *et al*., 2011). Assuming the Arada Unit as Neoproterozoic, it requires the presence of an angular unconformity of a very low-grade Lower Devonian-Carboniferous cover over a Neoproterozoic low-grade metamorphic unit. Both Palaeozoic and Neoproterozoic units were tectonically imbricate by the Variscan deformation, possibly during the Mississippian event. Furthermore, the acritarchs assemblages of the very low-grade Palaeozoic unit show affinities with Laurussian

Late Devonian faunas (Machado *et al*., 2008) and not to Gondwana one as expected if an Iberian Terrane correlation exist.

Other particular feature, is the presence of a Silurian (or Proterozoic?) mafic and ultramafic magmatism with tholeiitic nature and within plate to MORB signature interlayered in all Finisterra units (Noronha and Leterrier, 2000; Silva, Almeida *et al*., 2014). This contrast with the Iberian Terrane because, although in OMZ some rocks with this nature are described, they have a Cambrian to Ordovician age (Mata and Munhá, 1990; Sánchez-García *et al*., 2008; Pedro *et al*., 2010).

The inhered zircons older than 500 Ma contained in the Finisterra units (Pereira *et al*., 2010; Almeida, 2013; Almeida *et al*., 2014) also have a distinct pattern from the observed in Neoproterozoic and Cambrian units of the Iberian Terrane from, either OMZ (e.g. Fernandez-Suárez *et al*., 2002; Linnemann *et al*., 2008; Pereira *et al*., 2008; 2011; 2012b), or CIZ (e.g. Pereira *et al*., 2012c; Talavera *et al*., 2012; Orejana *et al*., 2015).

The general patterns of OMZ and CIZ and Finisterra Terrane present similar representative populations in 500-750 Ma and 1800-2000 Ma (Fig. 10 and 11). One of distinctive feature of OMZ and CIZ patterns is the almost absence of Mesoproterozoic inhered zircons (ca. 1000-1600 Ma; Fig. 10). In some samples collected in Finisterra Terrane the absence of Mesoproterozoic ages are also denoted, although some samples presents significant populations of Mesoproterozoic ages, representing 9% of all Neoproterozoic-Archean obtained ages for all samples, suggesting dissimilar sources of clastic sediments (Fig. 10 and 11).

The peculiar features of Finisterra Terrane in the context of the Iberian Geology, is also expressed by its structural pattern which strongly contrast with the general trend of the Iberian Terrane. Indeed the oldest observed deformation episode (D_1), although highly disturbed by the Carboniferous tectono-metamorphic events, shows the presence of N-S recumbent folds with top-to-W transport and a possible root zone in the PTFSZ. Such geometry has no equivalent in the Iberian Terrane, where a NW-SE general trend prevails (Dias *et al*., 2013; Moreira *et al*., 2014). We consider this deformation episode contemporaneous of the Silurian-Devonian metamorphic event, which has no equivalent in the Iberian Terrane.

During Carboniferous, both Terranes seem to have been a similar geodynamical evolution. Indeed, similar Mississippian HT metamorphic ages (ca. 340-310 Ma) are also described in the Iberian Terrane, not only in the CIZ (e.g. Bea *et al*., 2006; Castiñeiras *et al*. 2008) but also in OMZ (e.g. Pereira *et al*., 2012a). The same happens with the Ar-Ar Pennsylvanian metamorphic ages of the NW Iberian shear zones (Gutiérrez-Alonso *et al*., 2015). Also the late magmatism have Permian age, denoting similar evolution during the Carboniferous and Permian times.

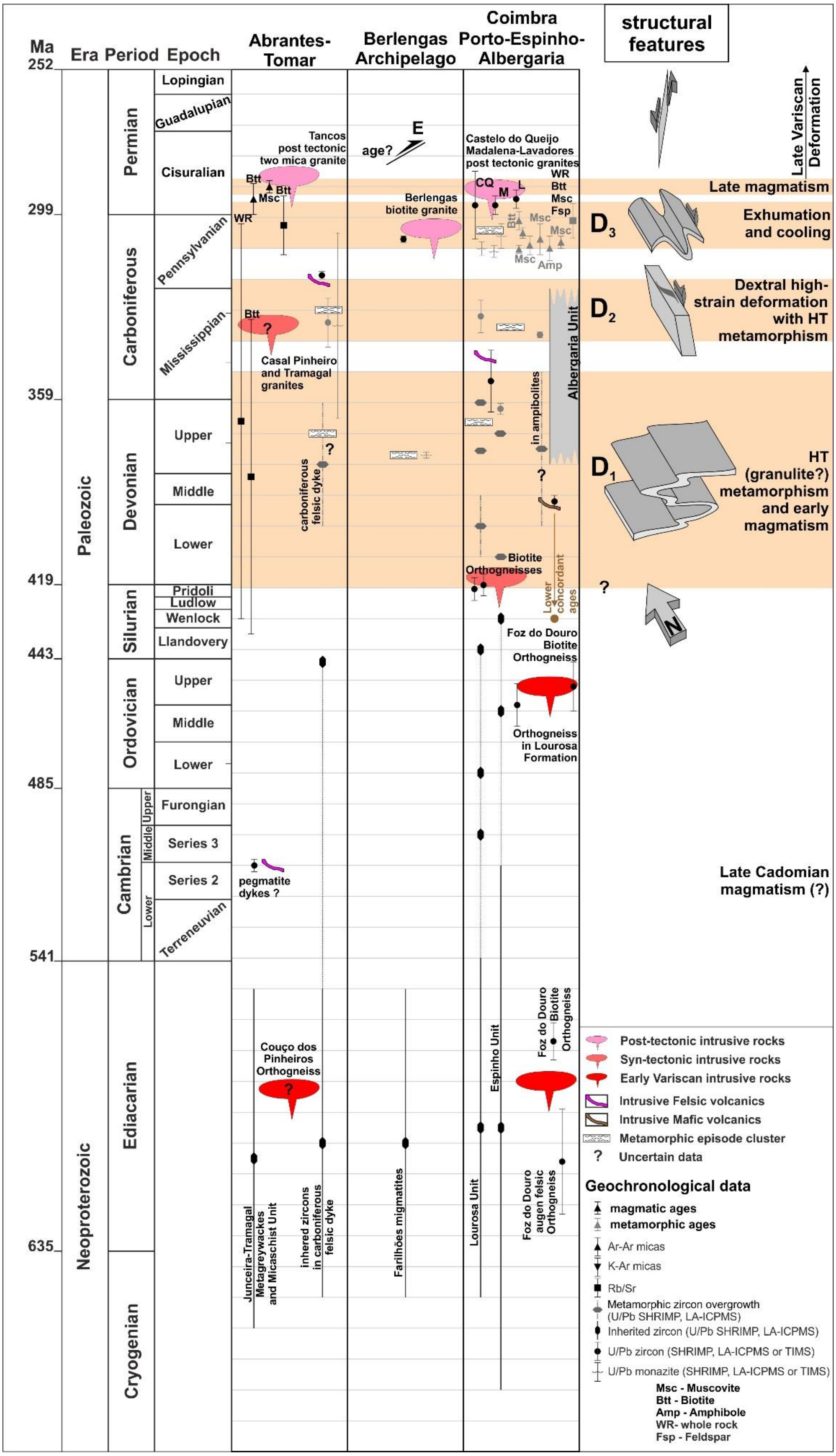

Figure 9 –Geological and geochronological features of Finisterra Terrane (data references on text).

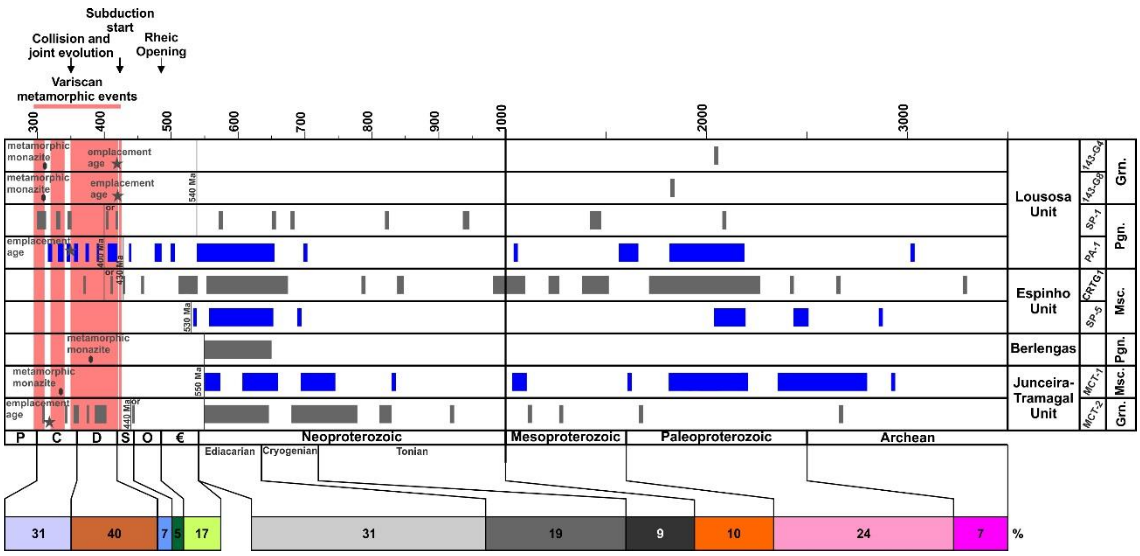

Figure 10 –Distribution of zircon and monazite ages from Finisterra Terrane units ($^{207}Pb/^{206}Pb$ ages from Pereira *et al*., 2010; Almeida, 2013; Almeida *et al*., 2014).

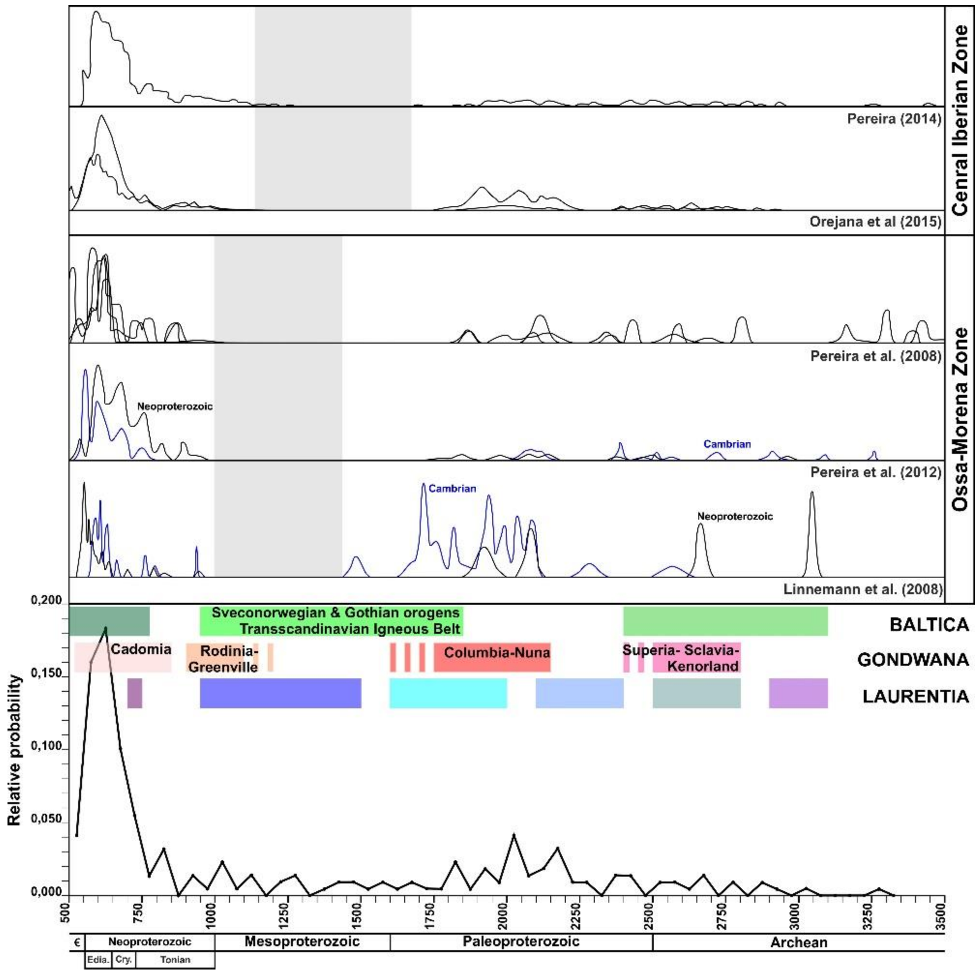

Figure 11 – Synthesis of Lower Cambrian-Neoproterozoic published zircon ages from Finisterra Terrane as probability plots, comparing with OMZ and CZI patterns and with possible zircon sources (adapted from Murphy *et al*., 2004; Pollock *et al*., 2007; Kuznetsov *et al*. 2014; Petterson *et al*., 2015).

Inhered zircons studies are performed in the Late Pennsylvanian Buçaco Basin (LA-ICP-MS; U-Pb in zircon; Dinis *et al*., 2012), located in the western border of CIZ near the Finisterra Terrane (Fig. 6), being identified some Silurian-Devonian (ca. 440-360 Ma) and also Mesoproterozoic (ca. 1350-1600 Ma) populations. This fact also emphasize the joint evolution of Iberian and Finisterra Terrane during Carboniferous times, being the Buçaco molasses Basin feed by both Terranes.

VII.1.6. Final Remarks and Paleogeographic Considerations

The geodynamics of the Finisterra Terrane cannot be dissociated from the European Variscides evolution. The nature of its eastern boundary (PTFSZ), its geological features and its relation with surrounding domains is crutial to understand its relation either with the northern branch of the Ibero-Armorican Arc (Dias *et al*., 2016), or with the northern Africa and the eastern America. This is not an easy task, due to Atlantic opening and the complexity of the Azores-Gibraltar plate boundary. The continuity of the narrow Finisterra Terrane is not obvious because it is isolated from the Iberian Terrane by a lithospheric shear zone and from the Central European Variscides due the Ibero-Armorican Arc. Nevertheless, its geological similarities with the Léon Domain and the Mid-German Crystalline Rise (MGCR), allow the establishment of some correlations among them.

VII.1.6.1. The Léon Domain

The Léon Domain is the small and northernmost domain of the Armorican Massif (Ballèvre *et al*., 2009; Faure *et al*., 2010; Fig. 12A). Its "exotic" features were extensively described since classical works (Bale and Brun, 1986; Le Corre *et al*., 1989). This block is composed of strongly deformed metamorphic rocks with Palaeozoic and Proterozoic age (Schulz *et al*., 2007; Ballèvre *et al*., 2009). It is interpreted as a complex stack of nappes, composed of several tectono-stratigraphic units (Faure *et al*., 2005; 2010; Ballèvre *et al*., 2009). The contact between the Léon Domain and the North and Central-Armorican Domains is debatable. While some considers the boundary in the Elorn fault, a (back)thrust with southward transport during Carboniferous times (Ballèvre *et al*., 2009), others put this limit in the Le Conquet-Penzé Shear Zone, a shear zone with tangential top-to-north transport, reactivated as a dextral strike slip fault during Carboniferous times (Faure *et al*., 2010).

VII.1.6.1.1. Main Tectonostratigraphic Units

In Léon Domain, a pile of nappes generated during the early Variscan deformation episode was emphasized (Faure *et al*., 2010), which will be summary described below.

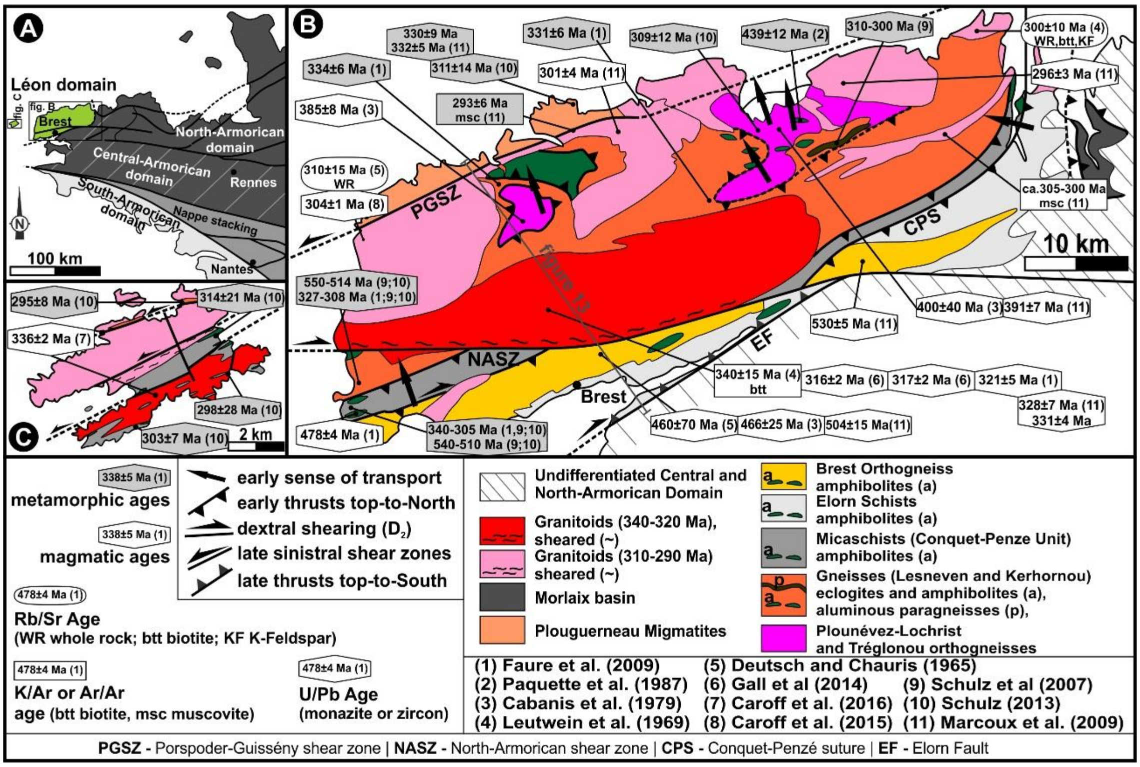

Figure 12 – The Léon Domain general framework:

A – Geographical relation wbetween the Léon Block and the Armorican Domains (adapted from Ballévre *et al*., 2009);

B – Simplified geological map of Léon Domain (adapted from Faure *et al*., 2010; Schulz, 2013);

C – Simplified geological map of Ouessant Island, the westernmost Léon outcrop (adapted from Schulz, 2013; Caroff *et al*., 2016).

- Parautochthonous unit

The lower unit (i.e. the Plounevez-Lochrist and Tréglonou Augen gneisses) is considered a parautochthonous unit composed of paragneisses intruded by K-feldspar Augen orthogneiss with biotite, garnet and sillimanite, both with intense migmatization. The emplacement age of the magmatic bodies is Lower-Middle Devonian (400±40 and 385±8 Ma, Ma, U-Pb in zircon, Cabanis *et al*., 1979; 391±7 Ma, U-Pb in zircon TIMS, Marcoux *et al*., 2009). Geochronological data on monazites, shows a HT metamorphic event during Late Carboniferous (ca. 320-310 Ma; Schulz, 2013) affecting the orthogneisses. Marcoux *et al*. (2009) notice a Lower Cambrian (ca. 530 Ma) inhered zircons in this ortho-derived body. The contact with the overlying lower nappe unit is marked a shear zone with blastomylonite rocks (Bale and Brun, 1986).

- Lower nappe

The lower nappe is a paragneiss unit (Lesneven and Kerhornou gneisses), mainly composed of biotite-garnet-sillimanite gneisses and micaschists bearing. Occasionally, they present evidences of migmatization, sometimes with generation of quartz-feldspar-biotite leucossomes related to an incipient crustal melting and anatetic granitoid dykes, as well deformed quartz-mylonites (Bale and Brun, 1986; Ballèvre *et al*., 2009; Faure *et al*., 2010). The transition from high-grade gneiss to sillimanite metatexites and diatexites is progressive (Faure *et al*., 2010).

This unit includes ortho-derived mafic tholeiites, with amphibolites, pyroxenites, serpentinites and eclogites (Bale and Brun, 1986; Faure *et al*., 2010). The eclogites presents garnet+clinopyroxene+plagioclase paragenesis, with amphiboles resulting from retrograde metamorphism. The age of the HP metamorphism is Silurian (439±12 Ma, U-Pb in zircon; Paquette *et al*., 1987), while the age of migmatization in the para-derived gneiss has been considered either Upper Mississippian (ca. 335-330 Ma; Faure *et al*., 2010), or Pennsylvanian (ca. 310-300 Ma; Schulz *et al*., 2007). The presence of Neoproterozoic-Cambrian metamorphic monazites in these gneisses (ca. 550-514 Ma) shows that the siliciclastic protolith is probably Neoproterozoic (Schulz *et al*., 2007; Schulz, 2013). Indeed, the uniform monazite chemical composition and the narrow range of Cadomian ages, exclude the detrital origin of these monazites (Schulz *et al*., 2007).

- Intermediate nappe

This nappe is mainly composed of biotite-garnet-staurolite to biotite-muscovite-oligoclase micaschists (Conquet-Penze Micaschists) with metacherts, quartzites and conglomeratic lenses (Faure *et al*., 2010). Some ortho-derived amphibolites and Early Ordovician meta-gabbros (478±4 Ma; U-Pb in zircon, LA-ICP-MS, Faure *et al*., 2010) are also present.

The staurolite zone metamorphism present a Carboniferous age (ca. 340-305 Ma; U-Ph in monazites; Schulz *et al*., 2007; Faure *et al*., 2010; Schulz, 2013). This unit also presents Neoproterozoic-Cambrian monazites, showing a Proterozoic age for its protolith (ca. 540-510 Ma; Schulz *et al*., 2007; Schulz, 2013). Also in this case the detrital origin of these monazites is excluded by Schulz *et al*. (2007).

- Upper nappe

The upper nappe consists of micaschists (Elorn Schists) intruded by ortho-gneiss (Brest orthogneiss), being considered the Armorican Massif basement (Faure *et al*., 2010).

The Elorn Schists, considered Late Proterozoic (Ballèvre at al., 2009; Faure at al., 2010), is a low-grade metamorphic unit (greenschists facies; Ballèvre *et al*., 2009), mainly composed of quartz phyllites, metagreywackes and metasandstones. Nevertheless, sometimes the metamorphic grade is slightly higher, generating low-grade micaschists (Faure *et al*., 2010).

The Brest gneiss is considered an ortho-derived gneiss with granodiorite composition (Bradshaw *et al*., 1967), showing hornfels metamorphism in Elorn metasediments (Bradshaw *et al*., 1967). The emplacement of this granodiorite is considered Cambrian-Early Ordovician in age (460±70 Ma, U-Pb in zircon – Deutsch and Chauris, 1965; 466 ± 25 Ma, U-Pb dissolution in zircon – Cabanis *et al*. 1979; 504±15 Ma and 530±5 Ma, U-Pb in zircons LA-ICP-MS and TIMS respectively – Macoux *et al*., 2009). Some inhered zircon populations are identified in Brest gneiss, being Neoproterozoic (600-700 Ma), Mesoproterozoic (ca. 1350 Ma) and Paleoproterozoic (ca. 1600, 1850 and 1900-2100 Ma) in age (Macoux *et al*., 2009).

- Migmatites of Plouguerneau

North of the Porspoder-Guissény shear zone outcrop the Migmatites of Plouguerneau, whose relation with the southern nappe pile units is not clear (Fig. 12B; Faure *et al*., 2010). These migmatites are strongly deformed, presenting monazites that indicate a Carboniferous age for the metamorphism (ca. 330-300 Ma; Ballèvre *et al*., 2009; Marcoux *et al*., 2009; Schulz, 2013), and inhered cores of Cadomian monazites (ca. 580 Ma; Ballèvre *et al*., 2009; Marcoux *et al*., 2009). Similar lithotypes and metamorphic ages were also obtained in the Ouessant Island (Fig. 12C; Caroff *et al*., 2015).

- Carboniferous magmatism

The Léon Domain was affected by an extensive Variscan magmatism comprising two main events (Fig. 12B):

- The oldest one, composed of granites and granodiorites with calco-alcaline signature (Bale and Brun, 1986), were intruded between 340 and 320 Ma (Cabanis *et al*., 1979; Faure *et al*., 2010; Marcoux *et al*., 2009; Le Gall *et al*., 2014). Some enclaves of Lesneven gneiss were identified in the Saint Renan-Kersaint massif (Bale and Brun, 1986). Inhered zircon populations with Neoproterozoic age were reported (540-650 and ca. 900 Ma; Marcoux *et al*., 2009).
- The younger assemblage, located in the northern sectors of Léon Domain, is composed of sub-alkaline granites and monzogranites (Bale and Brun, 1986), with ages ranging between 310-290 Ma (Cabanis *et al*., 1979; Marcoux *et al*., 2009; Caroff *et al*., 2015).

VII.1.6.1.2. Structural pattern

The high to medium-grade tectono-metamorphic units of the Léon Domain present a general ENE-WSW to NE-SW trend, affected by three main tectonic events.

The early deformation episode (D_1) is related to the nappe stacking of the tectonostratigraphic units. The contact between the tectono-metamorphic units are always D_1 tangential shears with transport top-to-the-N or NNW (Fig. 13; Faure *et al*., 2010; Bale and Brun, 1986), sometimes outlined by blastomylonite rocks (Bale and Brun, 1986). As mentioned, Lesneven and Kerhornou gneisses include eclogitic rocks (mostly retrogressed into amphibolites) with Silurian metamorphic ages (Paquette *et al*., 1987). Faure *et al*. (2010) considers the HP-metamorphism previous to nappe stacking episode (D_1), that constrain the timing of the episode as Late Silurian(?)-Devonian. In the Upper Unit, there is progressive increase of the deformation towards north, in direction of the contact with the Penzé-Le Conquet Micaschists unit (Le Conquet-Penzé Shear Zone; Bradshaw *et al*., 1967; Faure *et al*. 2010). These major shear zone presents poly-phasic deformation (Fig. 13; Faure *et al*., 2010), being reactivated during the second deformation episode (D_2) as a dextral strike-slip shear zone (Balé and Brun, 1986).

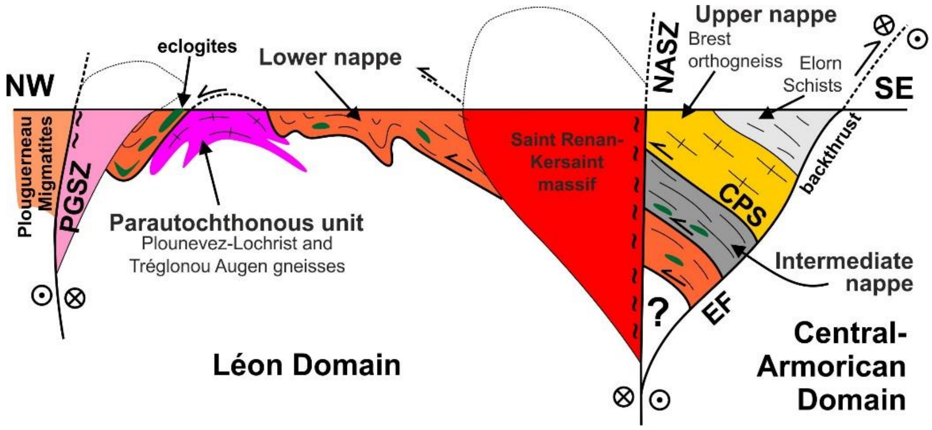

Figure 13 – Simplified cross-section from Léon Domain (adapted from Ballévre *et al*., 2009; Faure *et al*., 2010; Schulz, 2013).

All these units were deformed by a HT D_2 event (Bale and Brun, 1986) that often deeply reworked previous D_1 fabrics (Le Corre *et al*., 1989; Faure *et al*., 2005; 2010). In the Lesneven and Kerhornou gneisses, where D_2 is weaker, it is possible to show that the D_2 migmatization and crustal melting postdate the HP metamorphism (Faure *et al*., 2010). The metamorphic ages for D_2 are consistent with the Mississippian metamorphic event age (Schulz *et al*., 2007; Faure *et al*., 2010, Schulz, 2013). Thus, the HT metamorphism is temporally related with the emplacement of the previously mentioned first plutonic episode (ca. 340-320 Ma), represented by the Saint Renan-Kersaint massif, considered syn-tectonic with the D_2 E-W dextral kinematics of North-Armorican shear zone (NASZ; Bale and Brun, 1986; Schulz *et al*. 2007; Faure *et al*., 2010). Similar dextral wrench kinematics is described in Elorn Fault, which could be interpreted as a branch of the NASZ (Faure *et al*., 2005).

A late episode of deformation (D_3) is described, not only in the northern sectors of Léon Domain (Fig. 12B and 13; Le Corre *et al*., 1989; Marcoux *et al*., 2009), but also in the Ouessant Island (Fig. 12C; Caroff *et al*., 2016). This D_3 event in contemporaneous of the Plouguerneau migmatites and related to the NE-SW Porspoder-Guissény sinistral shear zone (Le Coure *et al*., 1989). The metamorphic ages obtained in the migmatites (311±14Ma, Th-U-Pb in Monazites; Schulz, 2013) and in the moscovites contained in myllonites of the Porspoder-Guissény shear zone (293±3 Ma, Ar-Ar; Marcoux *et al*., 2009), constrain this deformation episode between 310 and 290 Ma. However, some older metamorphic ages were also obtained in this migmatites (ca. 330 Ma; Marcoux *et al*., 2009), which seem to indicate that the migmatization could have begun during D_2 episode. The age for D_3 is compatible with field observations. Indeed, the NE-SW lineaments seem to have controlled the emplacement of the Late Carboniferous-Permian magmatic bodies (Fig. 12B; Ballèvre *et al*., 2009; Caroff *et al*., 2015) and the D_3 Porspoder-Guissény shear zone generates mylonites and ultramylonites that affects those granitoids (Marcoux *et al*., 2009; Le Gall *et al*., 2014; Caroff *et al*., 2016). These granites are considered posterior to D_2 dextral deformation episodes (Bale and Brun, 1986).

VII.1.6.2. The Mid-German Crystalline Rise

The Mid-German Crystalline Rise (MGCR), with a SW-NE trend, is mainly composed of medium- to high-grade gneisses, migmatites and plutonic rocks, partially covered by Permian to Quaternary deposits (Zeh and Will, 2010). This domain is located in northern sectors of the Saxo-Thuringian Domain, south of the low-grade metasediments and volcanic rocks of the Northern Phyllite Belt and the Rhenohercynian domain (Zeh and Will, 2010). It appears in small basement outcrops, generally called crystalline complexes, such as the Spessart, Kyffhäuser, Ruhla or Odenwald (e.g. Nasir *et al*., 1991; Dombrowski *et al*., 1995; Will and Schmädicke, 2003; Zeh and Will, 2010), being subdivided in several tectonostratigraphic units.

The crystalline complexes are mainly composed of gneisses and migmatites, containing amphibolite and, occasionally, marbles, calcisilicate rocks and quartzites. The gneisses, usually with para-derived nature, are interlayered with migmatites, micaschists and some ortho-derived gneisses (Dombrowski *et al*., 1995; Altherr *et al*., 1999). The metamorphism reaches high-temperature conditions, equivalent to amphibolite(-granulite) facies, with generation of cordierite+sillimanite+garnet+staurolite paragenesis (Will and Schmädicke, 2003). In the Odenwald Crystalline Complex, it was described retrogressed eclogites, now garnet amphibolites presenting tholeiitic nature and MORB to within-plate basalts geochemical signature (Scherer *et al*., 2002, Will and Schmädicke, 2001; 2003). This indicates the presence of a HP metamorphism previous to the HT one. The minimum age of this eclogites was considered to be Upper Devonian (357±6 Ma; Lu-Hf, garnet-whole rock; Scherer *et al*., 2002); according to authors, some resetting during the retrograde metamorphism may happen. Similar metamorphic ages have been also obtained in the Odenwald (375± 5 Ma; U-Pb in zircon, Todt *et al*., 1995) and Ruhla Crystalline Complexes (356.7±4.7 – zircon – and 352±8 Ma – monazite; U-Pb SHRIMP, Zeh *et al*., 2003), although in these cases the association with the HP event is not clear (Zeh and Will, 2010). The existence of an Upper Devonian metamorphic event is also supported by the presence of zircon grown in para-derived lithotypes (ca. 380-360 Ma; Reischmann and Anthes, 1996; Anthes and Reischmann, 2001). This metamorphic event is temporally associated to the emplacement of felsic (e.g. Albersweiler granitic gneiss; Reischmann and Anthes, 1996) and mafic-intermediate (e.g. Frankenstein gabbro; Kirsch *et al*., 1988; Zeh *et al*., 2005) bodies.

Concerning to the HT metamorphic event, its Mississippian age (340-320 Ma) is constrained by several geochronological studies (Nasir *et al*., 1991; Todt *et al*., 1995; Zeh *et al*., 2003; 2005), although some older ages are obtained in Odenwald (349±14 and 430±43 Ma, Th–U–Pb in Monazites; Will *et al*., 2016), emphasizing an early HT episode. The Mississippian metamorphic event is coeval with the emplacement of plutonic bodies, with granitic and granodiorite composition, such as Spessart diorite-granodiorite, Pretzsch-Prettin granite (Anthes and Reischmann, 2001), Edenkoben granite, Windstein granodiorite (Reischmann and Anthes, 1996) or Borntal intrusive complex (Anthes and Reischmann, 2001; Zeh *et al*., 2005).

However, in MGCR the plutonism is not restricted to previous events having a wide temporal range: Late Cambrian-Early Ordovician (e.g. Volkach Syenite; Anthes and Reischmann, 2001), Silurian-Devonian (ca. 420-410 Ma; e.g. Thal, Erbstrom and Silbergrund granite gneisses; Dombrowski *et al*., 1995; Brätz, 2000; Zeh *et al*., 2003) and Pennsylvanian-Early Permian bodies (310-290 Ma; e.g. Delitzsch granite; Anthes and Reischmann, 2001).

Geochronological studies in para-derived gneisses and migmatites (e.g. Zeh *et al*., 2001; 2003; 2005; Gerdes and Zeh, 2006; Zeh and Gerdes, 2010) and some ortho-derived gneisses (Anthes and Reischmann, 2001) show the presence of several populations of detrital zircons, although the general pattern of inhered zircons presents great dispersion within MGCR (Zeh and Gerdes, 2010). Indeed, the authors emphasize distinct patterns of inhered zircons, showing the presence of distinct clusters in Brotterode and Ruhla Groups (Fig. 14; both in Ruhla Crystalline Complex):

- Brotterode Group (Zeh *et al*., 2001; 2003; Gerdes and Zeh, 2006) – Lower Palaeozoic to Neoproterozoic representative populations (460-489, 500-590, 640-670 and 720-740 Ma) and clusters of Mesoproterozoic (ca. 1000 Ma), Paleoproterozoic (ca. 1800, 1950-2150 and ca. 2500 Ma) and Archean (ca. 2650 and 2830-2900 Ma) populations.

- Ruhla Group (Zeh and Gerdes, 2010) – Lower Palaeozoic to Ediacatian (435-485, 550-650Ma) and Meso- to Paleoproterozoic (900-1800 Ma with peaks at 1100, 1250, 1450, 1600 and 1800 Ma) representative populations and clusters of Paleoproterozoic (1850-1900 and ca. 2000 Ma) and Archean (ca. 2550, 2650 and 2800 Ma) populations.

The Meso- to Paleoproterozoic population, highly represented in Ruhla Group, is absence in Brotterode Group (Zeh and Gerdes, 2010). Similar Lower Paleozoic-Neoproterozoic and Mesoproterozoic populations were also described in Spessart and Kyffhäuser Crystalline Complexes (Anthes and Reischmann, 2001; Zeh *et al*., 2005).

Towards the southern boundary of the MGCR there are three (very) low-grade metamorphic units (Fig. 15), which contact with the Moldanubian Zone by the Lalaye-Lubine dextral shear zone (LLSZ; Franke, 2000; Zeh and Will, 2010). From south to north are:

- Villé unit: It is composed of late Cambrian to early Ordovician metapelitic to metapsammitic schists and quartzites;
- Steige unit: It is a monotonous metapelitic sucession deposited in a shallow marine environment during Ordovician to Silurian times. This unit thrusts the Villé unit;
- Bruche unit: It is a sedimentary and tectonic mélange comprising Frasnian black shales and Fammenian to early Carboniferous shelf and slope sediments, greywackes and conglomerates with calc-alkaline volcanic rocks.

The Steige and Villé Units are more deformed than the Bruche Unit (Skrzypek *et al*., 2014). Indeed, the Bruche unit is only affected by a Carboniferous tectono-metamorphic event (ca. 340-330) while the other units have a previous deformation episode (Skrzypek *et al*., 2014). All these very volcano-sedimentary sequences were intruded by diorites and granites during Carboniferous times.

6.3. Finisterra-Léon-MGCR Terrane; an essay of correlation

The Finisterra Terrane geological features shows clear similarities with the Léon Domain and MGCR. The correlation between the Léon Domain and the MGCR had already been established by several authors (e.g. Schulz *et al*., 2007; Faure *et al*., 2010; Ballévre *et al*., 2009; Franke, 2014; Franke and Dulce, 2016; Will *et al*., 2016). The strong similarities of these units with the geological features observed in the Finisterra Domain, led us to propose that the Léon-MGCR Domain extends up until Finisterra Terrane, defining the Finisterra-Léon-MGCR Terrane (Fig. 16).

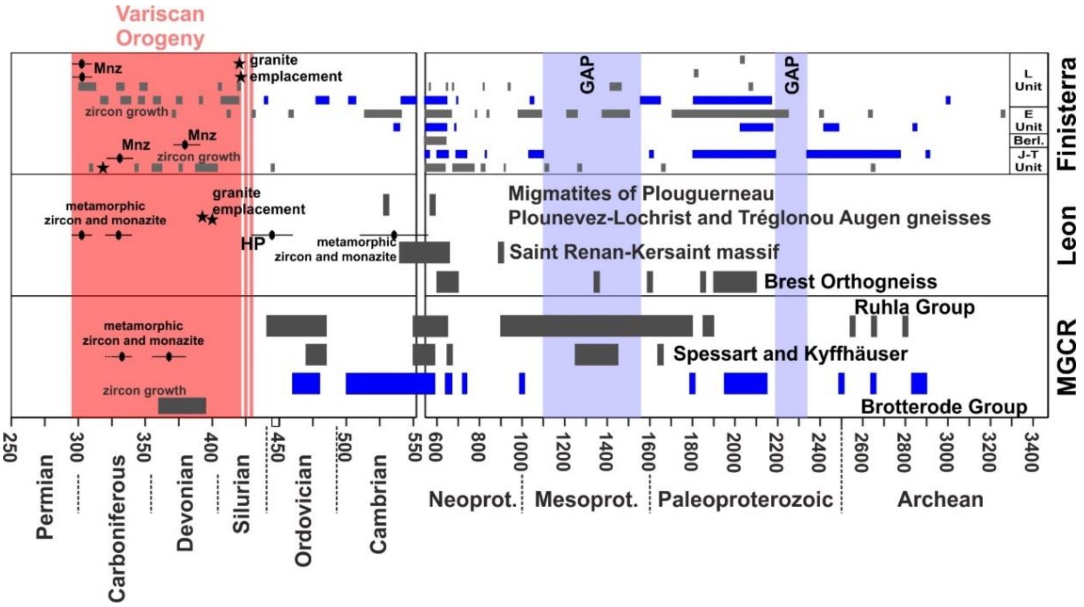

Figure 14 – Zircon patterns of Finisterra-Léon-MGCR Terrane (see references on text).

As previously described all these domains present an early magmatic event ca. (420-370 Ma), with the emplacement of several early granites (Fig. 15; Cabanis *et al*., 1979; Chaminé *et al*. 1998; Dombrowski *et al*., 1995; Brätz, 2000; Marcoux *et al*., 2009). The early magmatic event is accompanied by an early Variscan HT metamorphism (granulite facies conditions) during Devonian times (ca. 390-360 Ma; Fig. 15), however in the Léon Domain the geochronological data of such event seems to be absent (Schulz, 2013). Similar Late Silurian-Devonian metamorphic and felsic magmatic ages are rare in European Variscides. Although similar metamorphic ages are known, not only in the South and Central Armorican Domains (Balévre *et al*., 2009; Schulz, 2013), but also at OMZ (Moita *et al*., 2005), being generally associated to HP metamorphic rocks.

Also concerning the structural features, the three domains of the Finisterra-Léon-MGCR share a complex and polyphased deformation. The early tectono-metamorphic episodes are characterized by fold and thrust structures facing to N-NW in Léon and MGCR Domains (Faure *et al*., 2010; Zeh and Will, 2010) and to W in the Finisterra, probably rooting in its south (Le Conquet-Penzé Suture) and Eastern (PTFSZ) boundaries respectively.

Devonian HP metamorphic episode with eclogites is present both in Léon and MGCR Domains. These eclogites, which have been retrograded during Carboniferous events, have distinct ages (ca. 420 Ma in Léon Domain and ca. 360 Ma in MGCR; Paquette *et al*., 1987; Scherer *et al*., 2002). This seems to indicate a diachronic Variscan subduction during Devonian times that controls the early tectono-metamorphic stages of the Finisterra-Léon-MGCR Terrane, being developed at North of the proposed Terrane (Rheic Ocean subduction?). Although, until now, the eclogitic rocks are not described in the Finisterra Domain, this can be the result of the absence of detailed metamorphic studies in these area.

Abundant mafic and ultramafic magmatism associated to HT metamorphic units is described in all domains, but also to low-grade ones in Finisterra Terrane. The geochemical signature of this magmatism is compatible with within-plate to MORB basalts. Nevertheless, the meaning of this magmatism is not well constrained, it could be the result of the Cambrian-Ordovician times (or even during Silurian?) extensional processes related to the Variscan Ocean opening. More geochronological and isotopic studies should help to constrain the age and melting sources of such mafic magmatism.

Also the zircon patterns of MGCR and Finisterra Terrane show clear similarities. In both domains, two distinct patterns are present (Pereira *et al*., 2010; Zeh and Gerdes, 2010; Almeida *et al*., 2014; Fig. 14): samples with abundant Mesoproterozoic zircons and samples without Mesoproterozoic zircons. In both groups, Ordovician and Silurian zircons are sometimes found, showing that, at least, some of the metasedimentary protoliths of these rocks are Palaeozoic (Fig. 14). The existence of two distinct patterns of inhered zircons shows contrasting sources during sedimentation. As the Mesoproterozoic gap is one of the most distinctive feature of Gondwana-derived metasediments (e.g. Zeh and Gerdes, 2010; Pereira, 2014), the presence of significant populations of Mesoproterozoic zircons in some samples seems to preclude a Gondwana source for some samples. This shows that the Finisterra-Léon-MGCR represents a composite terrane, containing affinities with Peri-Gondwana terranes, but also with Laurussia-Baltica (?) continents (Zeh and Gerdes, 2010), where significant Mesoproterozoic ages are described (e.g. Murphy *et al*., 2004; Pollock *et al*., 2007; Kuznetsov *et al*. 2014; Petterson *et al*., 2015).

The low-grade metamorphic units described in Finisterra Terrane, namely as the black shales of Albergaria Unit, contains acritarchs with Laurussian affinities, imbricated in the previously deformed Arada Unit. The acritarchs assemblage is quite similar to those described in Rhenish massif (northern of MGCR; Fig. 16; Machado *et al*., 2008), which also includes Devonian clastic sediments derived from Caledonian Laurussia sources (Franke, 2000). Such data reinforce the zircon pattern previously described. The Albergaria Unit could also be a lateral equivalent of the Bruche unit (MGCR), presenting similar ages and lithotypes during the Upper Devonian, although the Carboniferous succession is slightly distinct. In both cases, the Devonian-Carboniferous units are spatially associated to older and mostly deformed low-grade units. The proximity of the Finisterra-Léon-MGCR Terrane to Laurussia in the Upper Devonian, contrasts to the Iberia and Armorica paleogeography that were still close to Gondwana.

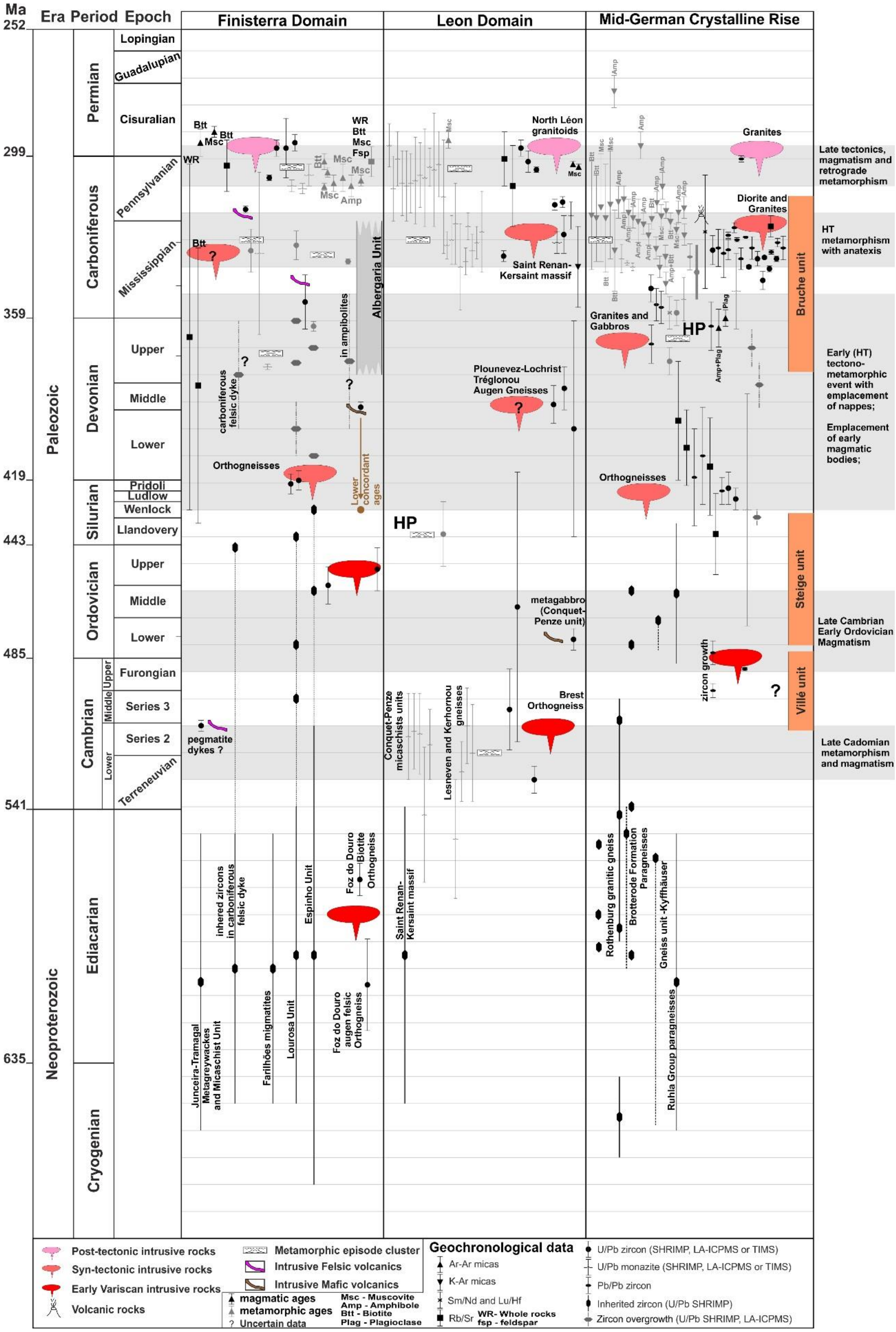

Figure 15 – Main tectono-metamorphic and magmatic features of Finisterra-Léon-MGCR Terrane (based on references presented on text).

The similarity of the MGCR zircon pattern with those exhibited in Pulo do Lobo Domain (*i.e*. Alájar Melange, Ribeira de Limas – Braid *et al*., 2011 – and Horta da Torre Formations – Perez-Cáceres *et al*., 2016) led to consider their possible correlation (Franke and Dulce, 2016; Fig. 16). Nevertheless, although the inhered zircon patterns are similar, the early magmatic event is absent in the Pulo do Lobo Domain and such correlation must to be strengthen.

Concerning the geodynamical meaning of the Finisterra-Léon-MGCR boundaries, distinct interpretation are possible for these lithospheric-scale shear zones:

- The Eastern Boundary of the Finisterra Terrane (PTFSZ) was considered a Paleo-transform fault with polyphasic deformation, at least, since early Variscan Cycle (Ribeiro *et al*., 2007). The western Finisterra boundary is unknown;
- The southern boundary of Léon (Le Conquet-Penzé Shear Zone) was be interpreted as an oceanic suture, separating this domain from the Armorican Massif (Faure *et al*., 2010). Nevertheless, the same authors do not exclude that this suture could represent only the closure of a thinned continental crust basin. Its northern boundary is presently hidden below the sea, being interpreted as the Rheic suture zone (Faure *et al*., 2010).
- The MGCR boundaries are generally covered by Permian to Quaternary sediments (Zeh and Will, 2010). Nevertheless, the contact with the southern Moldanubian Zone, is sometimes exposed, where it corresponds to the dextral LLSZ. The dextral shearing is Carboniferous, being superimposed on previous deformation (Skrzypek *et al*., 2014). However, the geodynamical significance of this major shear zone is not consensual, being interpreted either as a suture or an early Variscan detachment reactivated during Carboniferous (Skrzypek *et al*., 2014 to a discussion). The non-exposed northern limit of MGCR is considered a Variscan Suture (e.g. Skrzypek *et al*., 2014; Franke and Dulce, 2016).

According to previous data and interpretations the northernmost boundary of Finisterra-León-MGCR Terrane appears be consensual, representing a Variscan Oceanic suture (Rheic and/or Rheno-Hercynian Oceanic Suture?; Franke, 2000; Faure *et al*., 2010; Franke and Dulce, 2016). However, its southernmost boundary with Gondwana derived Terranes (Armorica and Iberia) is debatable and two distinct interpretations are possible:

- An active transform margin expressed by the PTFSZ, which connects the SW Iberian suture with the northern European suture(s), mainly the Le Conquet-Penzé Suture (and/or Paleotethys suture);
- All this boundary represents the suture zone of a secondary Palaeozoic Ocean (or a stretched continental crust basin) opened during Palaeozoic times.

The second hypothesis could explain the abundant Ordocivian to Silurian mafic and ultramafic rocks (Faure *et al*., 2010; Almeida *et al*., 2014), interspersed in the clastic rocks, which subsequently undergo HP metamorphism during Upper Silurian to Devonian times (Paquette *et al.*, 1987; Scherer *et al.*, 2002). The obtained Mesoproterozoic Sm-Nd model age for the amphibolites in Finisterra Terrane (Noronha and Leterrier, 2000) seems to be incongruent with the remaining data. Indeed, the mafic and ultramafic rock could represent, either mafic dykes intruded in Lower Palaeozoic siliciclastic sequences related to the Palaeozoic continental stretching or even oceanic floor rocks obducted during collisional processes.

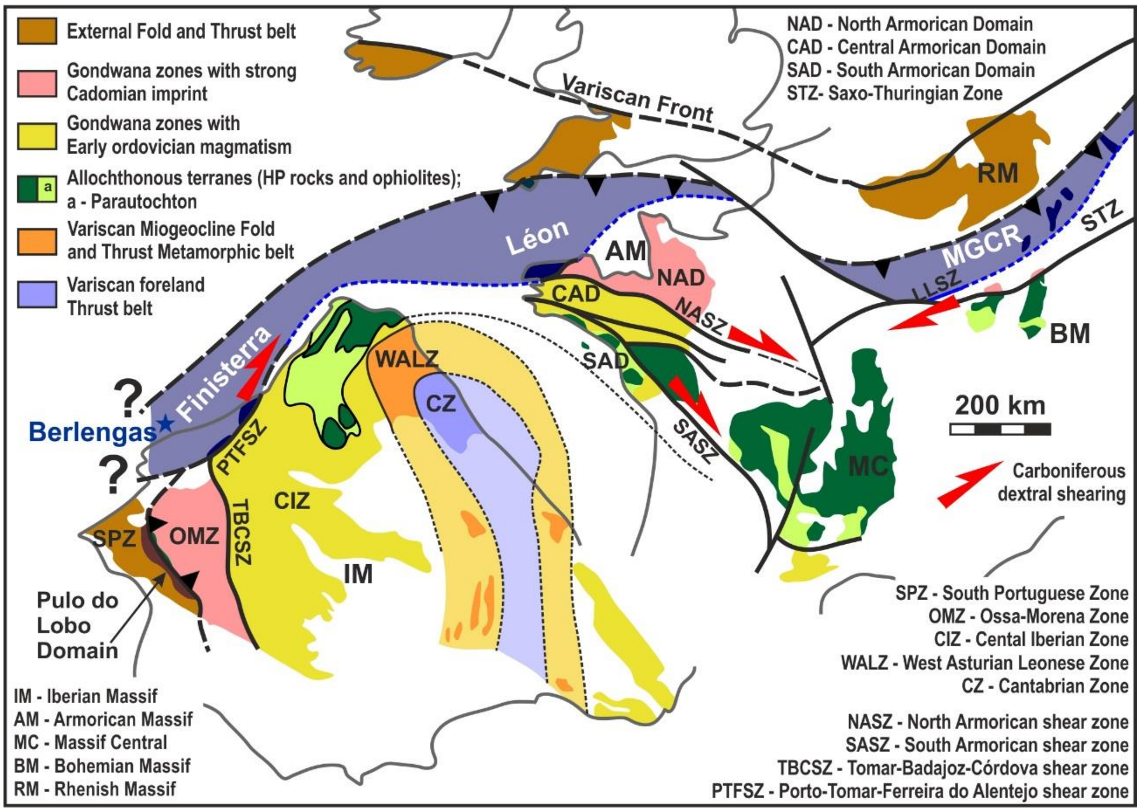

Figure 16 – The Finisterra-Leon-MGCR Terrane in the context of the European Variscides (adapted from Dias *et al*., 2016; Franke and Dulce, 2016).

Since Mississippian (ca. 340-330 Ma), the described Terrane and the other Peri-Gondwana Terranes shows similar metamorphic and magmatic ages (Fig. 15), which suggest they began to evolve together. Thus, such could represent the beginning of collision between Gondwana and Laurentia as often considered (Ribeiro *et al*., 2007; Moreira *et al*. 2014; Dias *et al*., 2016). In Mississippian, not only the Finisterra-León-MGCR Terrane but also the Iberian and Armorican Terranes, where affected by a pervasive dextral kinematics (e.g. PTFSZ, NASZ and LLSZ). The strong HT metamorphism related to the collisional process with melting generation was superimposed on previous events (Fig. 15), almost obliterating the early Variscan events in the Finisterra-Léon-MGCR Terrane.

The Neoproterozoic magmatism and metamorphism of Finisterra and Léon Domains (ascribable to the Cadomian event) and the presence of Late Cambrian-Early Ordovician magmatism (Fig. 15), also seems to indicate their North Gondwana affinities. Thus, the Finisterra-Léon-MGCR Terrane have a distict evolution of North Peri-Gondwana realm during Early Palaeozoic times, sharing, not only a common origin and evolution during Cadomian Cycle, but also a similar evolution after the Variscan collision during Carboniferous times (ca. 340-330 Ma).

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