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Abstract

This chapter, characterize the Finisterra Terrane, enhancing its differences from the neighbouring Iberian Terrane. The contact between these terranes is the Porto-Tomar-Ferreira do Alentejo Shear Zone, a major lithospheric structure whose complex Variscan evolution remains debatable. The lithostratigraphic, tectono-metamorphic and magmatic features observed in the Finisterra Terrane show that it was as an independent terrane during the Devonian. This situation changed during the Mississippian, when the main features of the Finisterra and the Iberian Terranes became similar, which indicates that both terranes evolved together since the Carboniferous times. The similarities of the Finisterra Terrane with the Central European Variscan domains, namely the Léon Block and the Mid-German Crystalline Rise, enable us to propose a new tectono-stratigraphic terrane (Finisterra-León-MGCR Terrane), which defines an arcuate pattern compatible with the Ibero-Armorican Arc.

7

The Finisterra-Léon-Mid German Cristalline Rise Domain; Proposal of a New Terrane in the Variscan Chain

AQ1

N. Moreira, J. Romão, R. Dias, A. Ribeiro, and J. Pedro

Abstract

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AQ2

7.1 Introduction

The Iberian Massif presents a well-developed arcuate pattern, in close relationship with the genesis of the Ibero-Armorican Arc (Fig. 7.1a; 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 westerly interrupted by one of the most important Iberian Variscan structures, the Porto-Tomar-Ferreira do Alentejo shear zone (Fig. 7.1b; PTFSZ). The geodynamic interpretation of this shear zone, with polyphase tectonic deformation, is controversial. Indeed, it has been interpreted as an active lithospheric-scale shear zone since the early Devonian (Dias and Ribeiro 1993), possibly reactivating an older structure (Cadomian?; Ribeiro et al. 2007, 2013). However, an alternative interpretation suggests that the PTFSZ has been active only during the Mississippian as a dextral transcurrent shear zone (Pereira et al. 2010; Martínez Catalán 2011; Gutiérrez-Alonso et al. 2015).

AQ3

Whatever the meaning of the PTFSZ, it is clear that PTFSZ marks a major boundary between a western crustal block—Finisterra Block—and the adjacent Central Iberian (CIZ) and Ossa-Morena (OMZ) Zones, both from Iberian Terrane (Fig. 7.1b; Ribeiro et al. 2007), each one with distinct geological features and geodynamical evolution, at least, during the Palaeozoic. This work presents a geological overview of the western block of PTFSZ, which has been used as the base to discuss and propose the Finisterra Block as a new terrane in the Iberian Variscides. The geological affinities between this block, the Léon Block and Mid German Crystalline Rise seems to indicate an independent terrane within the Variscan Chain.

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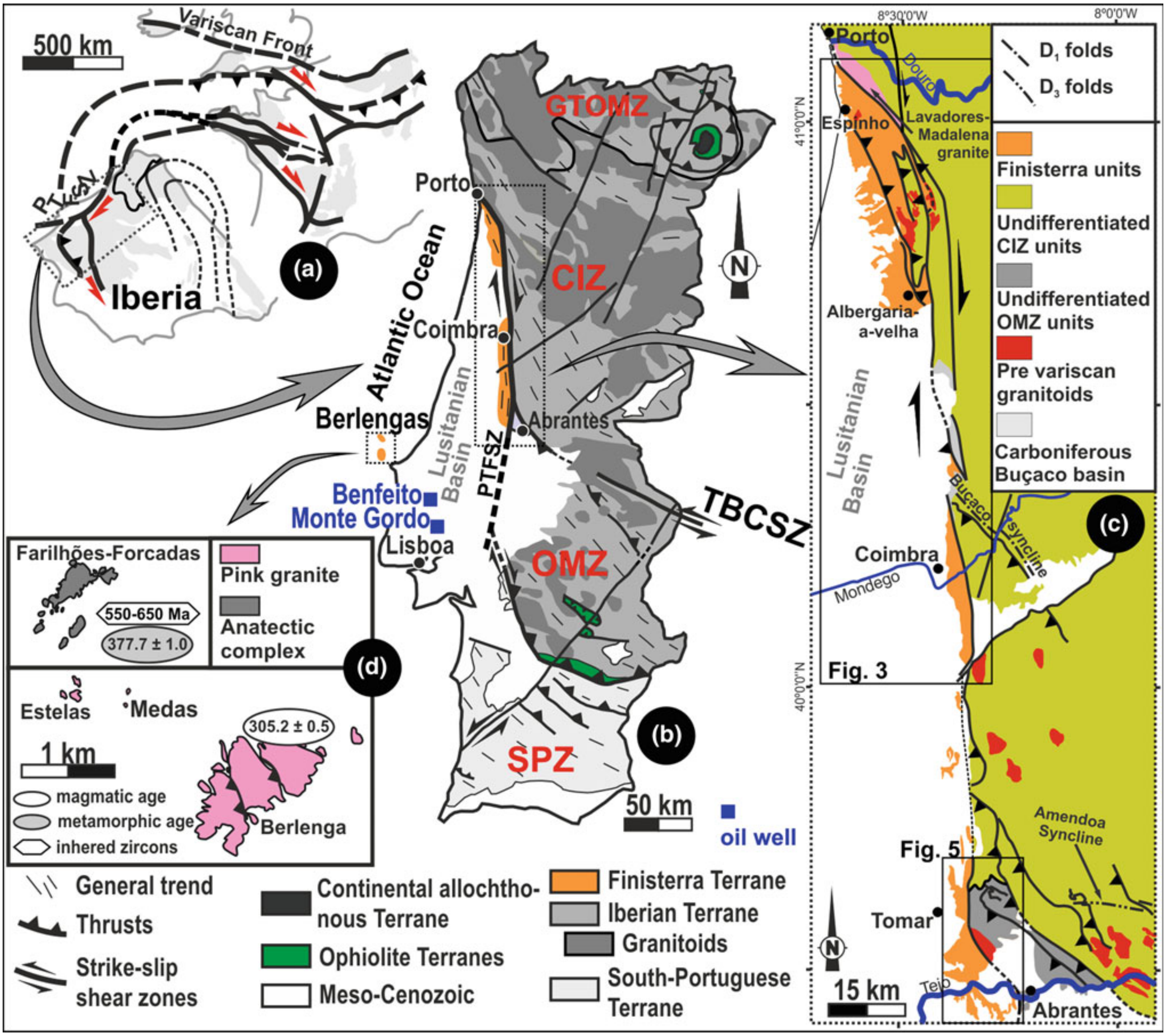

Fig. 7.1 The Finisterra block in the context of the Iberian Variscides: **a** The Ibero-Armorican arc (adapted from Dias et al. 2016); **b** General overview of Finisterra block (adapted from Ribeiro et al. 2013); **c** The Finisterra outcrops in the vicinity of PTFSZ (adapted from Chaminé et al. 2003a; Romão et al. 2014, 2016; Moreira 2017); **d** The Berlengas archipelago geological features (adapted from Bento dos Santos et al. this volume)

7.2 Tectonostratigraphy of the Finisterra Bock

West of the PTFSZ, low and high-grade tectonostratigraphic units are defined in four sectors (Porto-Espinho-Albergaria-a-Velha, Coimbra, Abrantes-Tomar and Berlengas Archipelago; Fig. 7.1c, d; Chaminé et al. 2003a, b; Ferreira Soares et al. 2007; Ribeiro et al. 2013; Romão et al. 2013, 2016; Moreira et al. 2016a, b; Bento dos Santos et al. this volume). The continuity between these sectors is not observable due to the overlying Meso-Cenozoic sedimentary cover (Fig. 7.1b). An overview of the tectonostratigraphic succession of these sectors is shown in Fig. 7.2.

7.2.1 The Porto-Espinho-Albergaria-a-Velha and Coimbra Sectors

Four pre-Mesozoic tectonostratigraphic units were defined between Porto, Albergaria-a-Velha and Coimbra (Figs. 7.2 and 7.3; Chaminé 2000; Chaminé et al. 2003a, b;

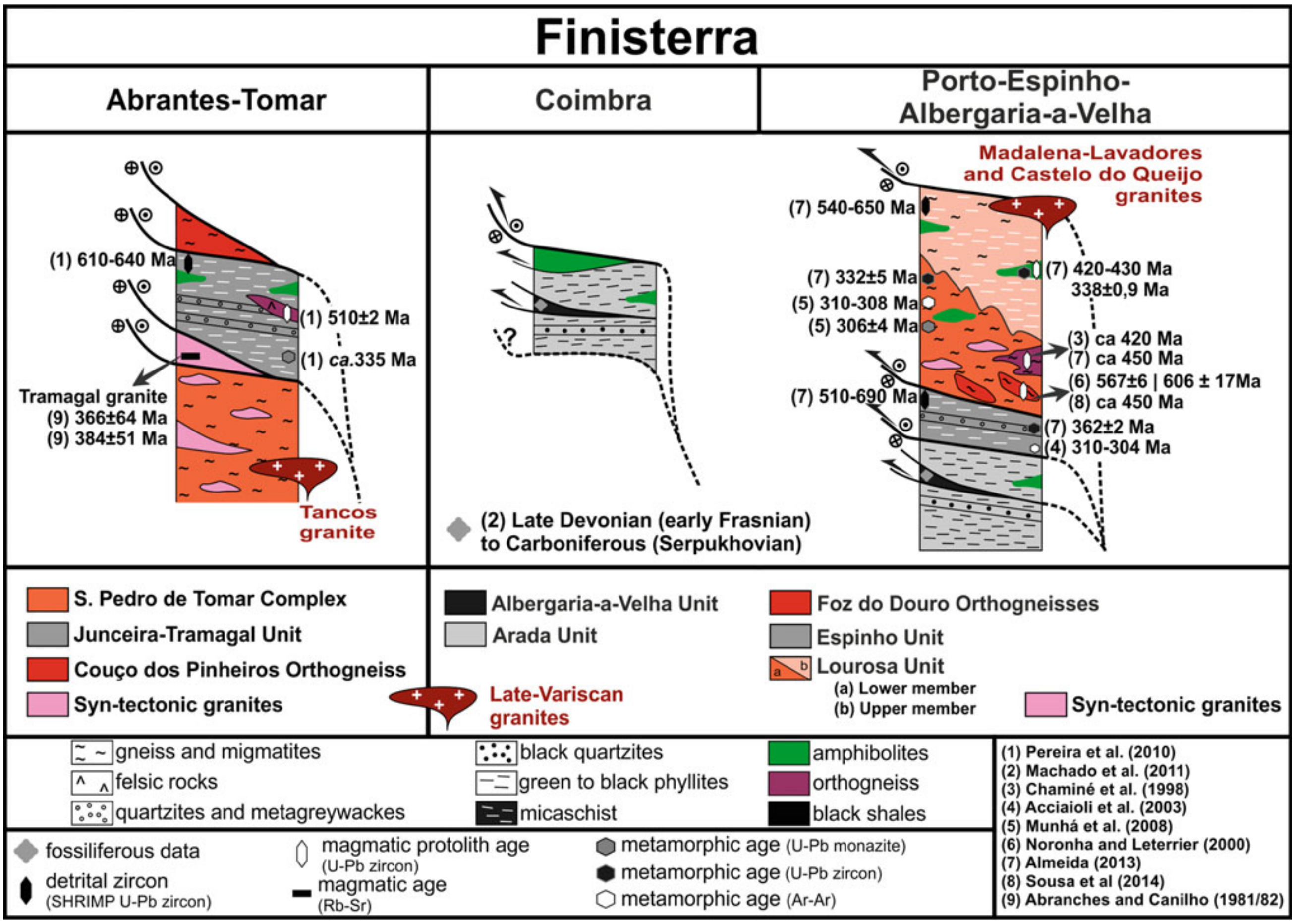

Fig. 7.2 Simplified tectonostratigraphic successions of Finisterra sectors (see references in the text)

Pereira et al. 2007; Machado et al. 2008, 2011; Ribeiro et al. 2013). The boundaries between these units are always Variscan shear zones.

7.2.1.1 Lourosa Unit

Two members were individualized in the Lourosa Unit (Fig. 7.2): the lower member mostly composed of migmatites, ortho- and paragneisses and the upper member dominated by (garnet-)biotite-micaschists (Chaminé 2000; Chaminé et al. 2003a). This high-grade unit is considered of Neoproterozoic in age (Chaminé 2000), but this age appears to be doubtful according to more recent data. Indeed, detrital zircon population obtained in a granite and a paragneiss from this unit provided a Lower Cambrian to Ediacarian para-derived protolith age (540–650 Ma; U–Pb in zircon, LA–ICP–MS), however some Upper Cambrian-Ordovician to Devonian zircons were also recognized (Fig. 7.4a; Almeida 2013; Almeida et al. 2014). The younger ages may result from analytical problems, U–Pb re-equilibrium during high temperature (HT) metamorphism or, alternatively, may indicate Palaeozoic ages of some of the para-derived lithotypes. Furthermore, the granite and paragneiss inhered zircon populations are distinct (Almeida 2013): while the paragneiss contains Mesoproterozoic populations, in the granite such ages are absent (Fig. 7.4a). This difference has paleogeographic importance and will be discussed.

Both members have (olivine-)amphibolites and amphibolic schists with geochemical signature similar to within-plate to MORB basalts (Montenegro de Andrade 1977; Silva 2007; Aires and Noronha 2010) and some orthogneisses. Lower Devonian protolith ages were obtained for mafic amphibolite (392 ± 2 Ma; U–Pb, LA–ICP–MS in zircons; Almeida et al. 2014), although older concordant ages were also obtained in these ortho-derived rocks (ca. 420–430 Ma; Almeida 2013). Therefore, Silurian-Devonian ages could be ascribed to these amphibolites or at least part of them. Concerning the orthogneisses, several ages were obtained for their protolith: Ordovician (459 ± 7 Ma; U–Pb, LA–ICP–MS in zircon; Almeida 2013), Silurian (420 ± 4 Ma in Lourosela and 419 ± 4 Ma in Souto Redondo; U–Pb, TIMS in zircon; Chaminé et al. 1998) and Upper Devonian-Mississippian (353 ± 10 Ma; U–Pb, LA–

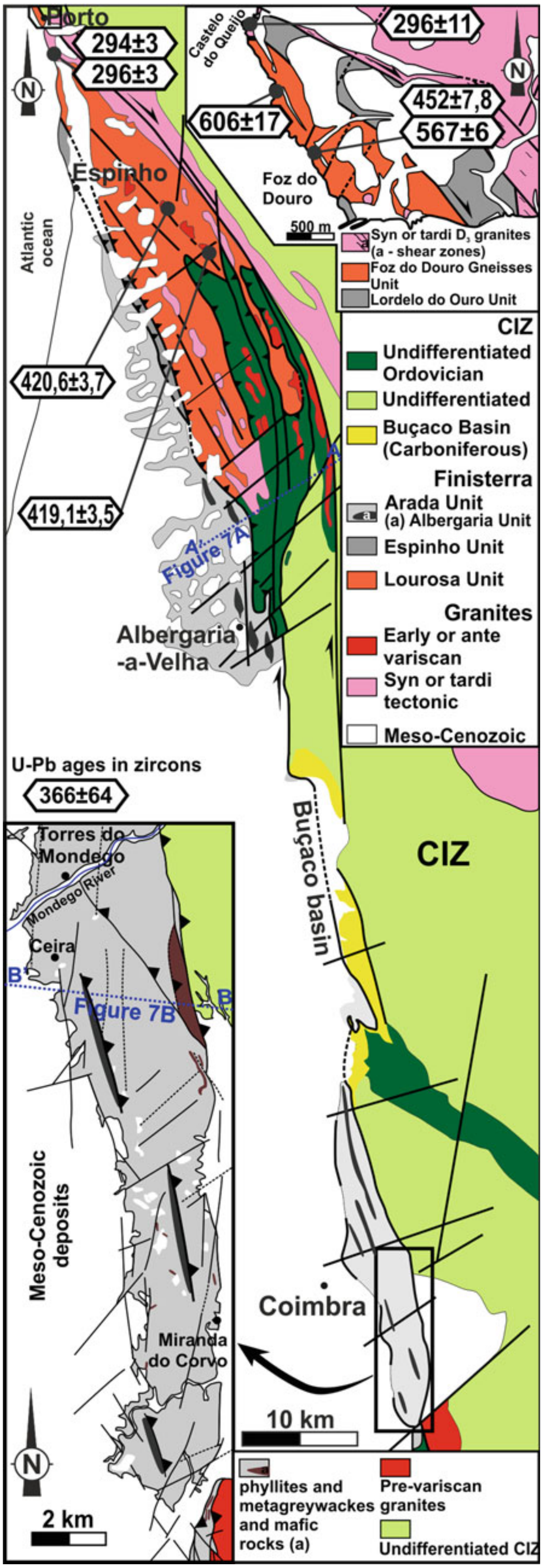

Fig. 7.3 Simplified geological map and geochronological data for the Porto-Espinho-Albergaria-a-Velha and Coimbra sectors (blue lines correspond to the cross sections of 7A, 7B; adapted from Chaminé et al. 2003a; Ferreira Soares et al. 2005; Pereira et al. 2007; LNEG 2010; Machado et al. 2011; Dinis et al. 2012)

ICP–MS in zircons; Almeida 2013). Mississippian metamorphic ages were obtained in a gneiss and an amphibolite (332 ± 5 Ma and 339 ± 1 Ma; U–Pb in zircons—SHRIMP and LA-ICP–MS respectively; Almeida 2013).

7.2.1.2 Foz Do Douro Gneissic Unit

The Foz do Douro Gneissic Unit comprises tonalitic and granitic ortogneisses with intercalations of mylonites, paragneisses, micaschists and amphibolites. The amphibolites have tholeiitic MORB geochemical affinity (Noronha and Leterrier 1995, 2000) and their Sm–Nd isotopic fingerprint suggest a Mesoproterozoic model age (TDM; ca. 1050 Ma; Noronha and Leterrier 2000). This unit is considered a geological equivalent of the Lourosa Unit described above, based on its lithological, geochemical and structural features (Chaminé et al. 2003a).

The oldest record of magmatism in the Finisterra Block has been reported in the orthogneisses of this unit (Fig. 7.2), namely an Ediacarian age for its protoliths (567 ± 6 Ma in biotitic orthogneiss and 606 ± 17 Ma in augen felsic gneisses; U–Pb, isotopic dilution in zircons; Noronha and Leterrier 2000). However, more recently, the protolith of the biotitic orthogneiss was re-evaluate, yielding an Upper Ordovician age (452 ± 8 Ma; U–Pb, SHRIMP in zircons; Sousa et al. 2014), leaving room to protolith age uncertainties.

The eastern boundary of Foz do Douro Gneissic Unit is underlined by a contact with a narrow band of micaschists and quartz-tectonites (locally named Lordelo do Ouro Unit; Fig. 7.3), which is affected by a pervasive dextral shearing, being considered as the local expression of the PTFSZ (Ribeiro et al. 2009). The strong similarities between the Lordelo do Ouro Unit and the micaschists interlayered in the Foz do Douro Gneisses Unit indicate that both units are part of the Finisterra Block.

NW-SE trending late-tectonic Variscan granites (Castelo do Queijo and Lavadores-Madalena) intrude the northernmost boundary of the Lourosa Unit and the Foz do Douro Gneissic Unit (Fig. 7.3; Chaminé et al. 2003a; LNEG 2010). This magmatism is Late Carboniferous—Permian in age: 296 ± 11 Ma for the Castelo do Queijo granite (U–Pb, LA–ICP–MS in zircons; Martins et al. 2014) and 296 ± 3 Ma (U–Pb, LA–ICP–MS in zircons), 294 ± 3 Ma (U–Pb, SHRIMP in zircons) and 298 ± 11 Ma (U–Pb, isotopic dilution in zircons) for the Lavadores-Madalena granite (Martins et al. 2011, 2014).

7.2.1.3 Espinho Unit

The Espinho Unit outcrops to the West of the Lourosa Unit (Fig. 7.3) and it is composed of a narrow band of staurolite-garnet-biotite micaschists, locally with intercalations of (mylonitic garnet-)quartzites (Fig. 7.2; Chaminé 2000; Chaminé et al. 2003a). Two HT metamorphic events

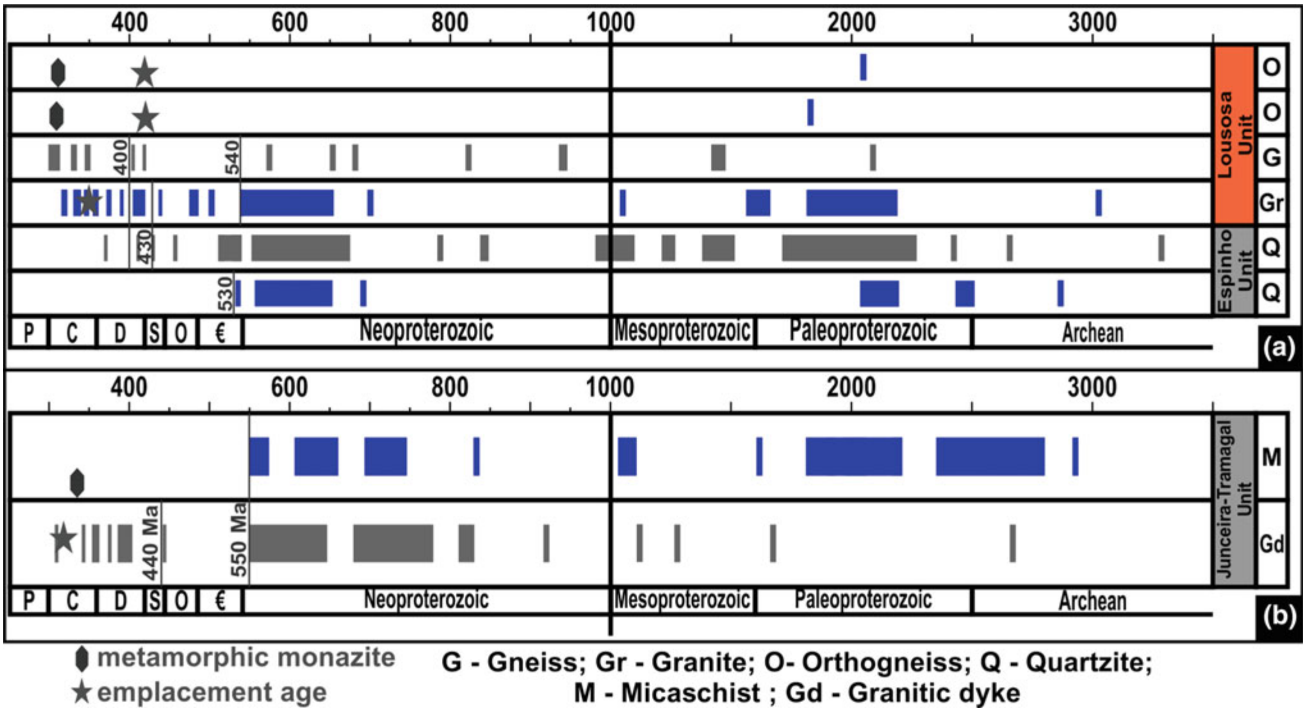

Fig. 7.4 Simplified pattern of the zircon populations of Finisterra sectors (the grey colours outline the samples with Mesoproterozoic populations): **a** Lourosa and Espinho Units (geochronological data from Chaminé et al. 1998; Almeida 2013; Almeida et al. 2014); **b** Junceira-Tramagal Unit (^{207}Pb/^{206}Pb ages obtained by Pereira et al. 2010)

are recorded in the paragenesis of garnet quartzites: the first reaches the sillimanite zone while in the second one the staurolite zone was attained (Fernández et al. 2003).

Geochronological data recovered from the quartzites (U–Pb, LA–ICP–MS in zircons; Almeida 2013; Almeida et al. 2014) indicate a Lower Cambrian protolith age (510–690 Ma is the youngest population of inhered zircons). However, as in the Lourosa Unit, Ordovician and Silurian-Devonian ages were also obtained in zircons displaying detrital morphologies (Fig. 7.4a; Almeida 2013; Almeida et al. 2014). These data may be biased by the same reasons as those described for the Lourosa Unit. Some quartzites do not present Mesoproterozoic zircon populations, while in others such populations are representative (Fig. 7.4a), as it was also emphasized in Lourosa Unit.

An Upper Devonian metamorphic event (362 ± 2 Ma; U–Pb LA–ICP–MS in zircon) is recorded in the mentioned quartzite layers (Almeida 2013; Almeida et al. 2014).

7.2.1.4 Arada Unit

This unit (Fig. 7.3) is composed of black to green phyllites, metagreywackes, black quartzites and mafic rocks with a tholeiitic geochemical fingerprint (Silva 2007), which are affected by chlorite-biotite zone metamorphism (Ferreira Soares et al. 2007). The lithological resemblances of this unit with the Ediacarian “Série Negra” of the OMZ have been emphasised by some authors (Beetsma 1995; Chaminé 2000; Chaminé et al. 2003a; Ferreira Soares et al. 2007; Pereira et al. 2007). However, the absence of the black chert (flint) horizons, typical of the “Série Negra”, is assumed to represent a distinct feature of the Arada Unit. The age of this lithological succession is open to debate, although it is considered as Neoproterozoic (Beetsma 1995; Chaminé 2000; Ferreira Soares et al. 2007).

7.2.1.5 Albergaria Unit

The Albergaria Unit crops out as narrow bands within the Arada unit (Figs. 7.2 and 7.3; Chaminé et al. 2003b). It is composed of very low-grade (low anchizone; Chaminé et al. 2003b) black shales and siltstones, which yielded Laurussia-akin acritarch assemblages of Frasnian-Serpukhovian age (Chaminé et al. 2003b; Machado et al. 2008, 2011). This unit is tectonically deformed by a single deformation episode while the older Arada Unit is deformed by two episodes. This fact combined with the distinct metamorphism shown by these units indicates the existence of an unconformity between them. Both units were tectonically imbricated during Pennsylvanian.

7.2.2 The Abrantes-Tomar Sector

In the Abrantes-Tomar sector, a N-S to NNW-SSE elongate high-grade tectonostratigraphic succession was recently defined (Fig. 7.5; Romão et al. 2013, 2016; Moreira et al. 2016a, b; Moreira 2017). The contact between the tectonostratigraphic units is always underlined by Variscan shear zones.

7.2.2.1 Pedro de Tomar Complex

The S. Pedro de Tomar Complex represents the basal unit of the Abrantes-Tomar sector. To the East this complex contacts with the Junceira-Tramagal Unit, while to the West it is covered by the Meso-Cenozoic formations (Fig. 7.5). This complex is characterized by medium to fine-grained strongly deformed para- and ortho-gneisses, interlayered with micaschists, mylonites and migmatites. The most representative lithotypes are paragneisses with sillimanite zone metamorphism (Fig. 7.6a). The orthogneisses are generally less deformed and clearly related to the anatexis and melting of para-derived rocks. The feldspars present undulose extinction and dynamic recrystallization which, coupled with the presence of sillimanite, suggests minimum temperatures around 500–600 °C (Passchier and Trouw 2005; Bucher and Graper 2011). Some gneisses result from migmatitic processes superimposed by a strong high-strain dextral shearing, giving rise to the gneissic foliation.

The protolith and metamorphic ages of these gneisses and migmatites are uncertain, being considered respectively of Neoproterozoic and Mississippian in age when compared with the overlying Tramagal-Junceira Unit (see below).

The high-grade tectonostratigraphic units are intruded by the syn-tectonic N-S elongated Tramagal and Casal Pinheiro granites (Romão et al. 2013, 2016; Moreira 2017). These are two mica granites with tourmaline and sillimanite, which indicates their peraluminous character and anatetic nature (e.g. Clarke 1981; Pesquera et al. 2012). A Mississippian emplacement age is assumed to these granites, because they are controlled by the second deformation episode, showing hot-plastic dextral shearing coeval of their crystallization (Fig. 7.6c). The field data are in accordance with inaccurate geochronological data of Tramagal granite (366 ± 64 Ma and 384 ± 51 Ma; Rb/Sr method, respectively in whole rock and in biotite; Abranches and Canilho 1981/82).

A post-tectonic porphyritic two-mica granite, not affected by ductile deformation, intrudes the S. Pedro de Tomar Complex (Fig. 7.5; Tancos Granite). Geochronological data shows an Early Permian age to its cooling based on K–Ar (294 ± 5 Ma, biotite and 290 ± 2 Ma, muscovite; Neves et al. 2007) and Rb–Sr (312–293 Ma, biotite; Mendes 1967/68) methods.

7.2.2.2 Junceira-Tramagal Unit

The Junceira-Tramagal Unit crops out in a narrow N-S to NNW-SSE 40 km long band from Ferreira do Zêzere to Tramagal (Fig. 7.5). This unit is composed of garnet and staurolite-garnet micaschists, subordinate metagreywackes, metaquartzwackes and black schists. Early HT (Variscan?) migmatization occurs near the Tramagal Granite and this migmatization could derive from the palingenesis of older deformed (Cadomian?) migmatites, also displayed in the Neoproterozoic units of the OMZ East of Abrantes (Henriques et al. 2015). The micaschists paragenesis is dominated by biotite + muscovite + quartz + plagioclase + opaque minerals ± K-feldspar. Millimetric to centimetric garnet and staurolite porphyroblasts were generated during the metamorphic peak conditions, being ascribed to the amphibolite facies (staurolite zone; Fig. 7.6b).

Geochronological data (U–Pb, LA–ICP–MS in zircons; Pereira et al. 2010) indicate an Ediacarian protolith age for the para-derived lithotypes of the Junceira-Tramagal Unit (550–660 Ma is the most recent population of inhered zircons) and a Mississippian metamorphic episode (ca. 335–330 Ma). Neoproterozoic (700–750 and ca. 830 Ma) Mesoproterozoic (1050–1150 Ma), Paleoproterozoic (ca. 1650 and 1880–2200 Ma) and Paleoproterozoic-Archean (2350–2900 Ma) inhered zircon populations were also found (Fig. 7.4b).

Ortho-derived lithotypes are also found in this unit, namely:

- Amphibolite dykes with green amphibole + plagioclase + opaque minerals ± quartz, typical of the amphibolite facies, and with unknown age;
- Quartz-feldspatic orthogneisses, sometimes with mylonitic textures, interpreted as the result of the tectonometamorphism affecting felsic-rich rocks (pegmatitic dykes?), present Lower Cambrian protolith ages (510.3 ± 2.0 Ma; U–Pb, LA–ICP–MS in zircons; Fig. 7.5; Pereira et al. 2010);
- (Micro-)granitic dykes, less deformed than the quartz-feldspatic orthogneisses and cutting the gneissic foliation, with Pennsylvanian age (318.7 ± 1.2 Ma; U–Pb, LA–ICP–MS in zircons; Pereira et al. 2010). Several zircon populations were found in this granite (Fig. 7.4a), with emphasis on the Silurian-Carboniferous (ca. 350–420 Ma) and the Mesoproterozoic (ca. 1100, 1270 Ma) ages.

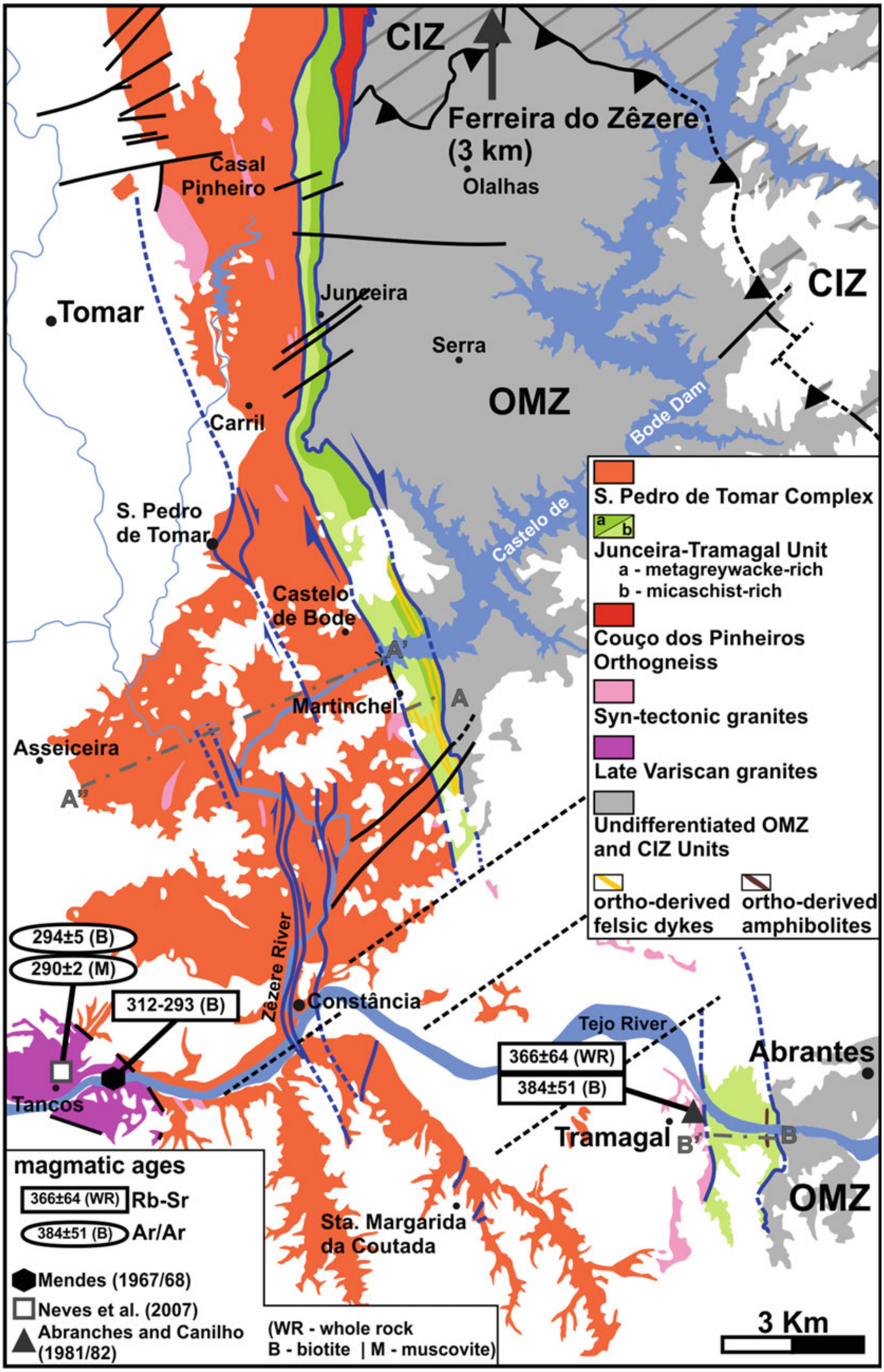

Fig. 7.5 Simplified geological map of the Abrantes-Tomar sector and published geochronological ages (grey lines show the location of the Fig. 7.7c cross sections)

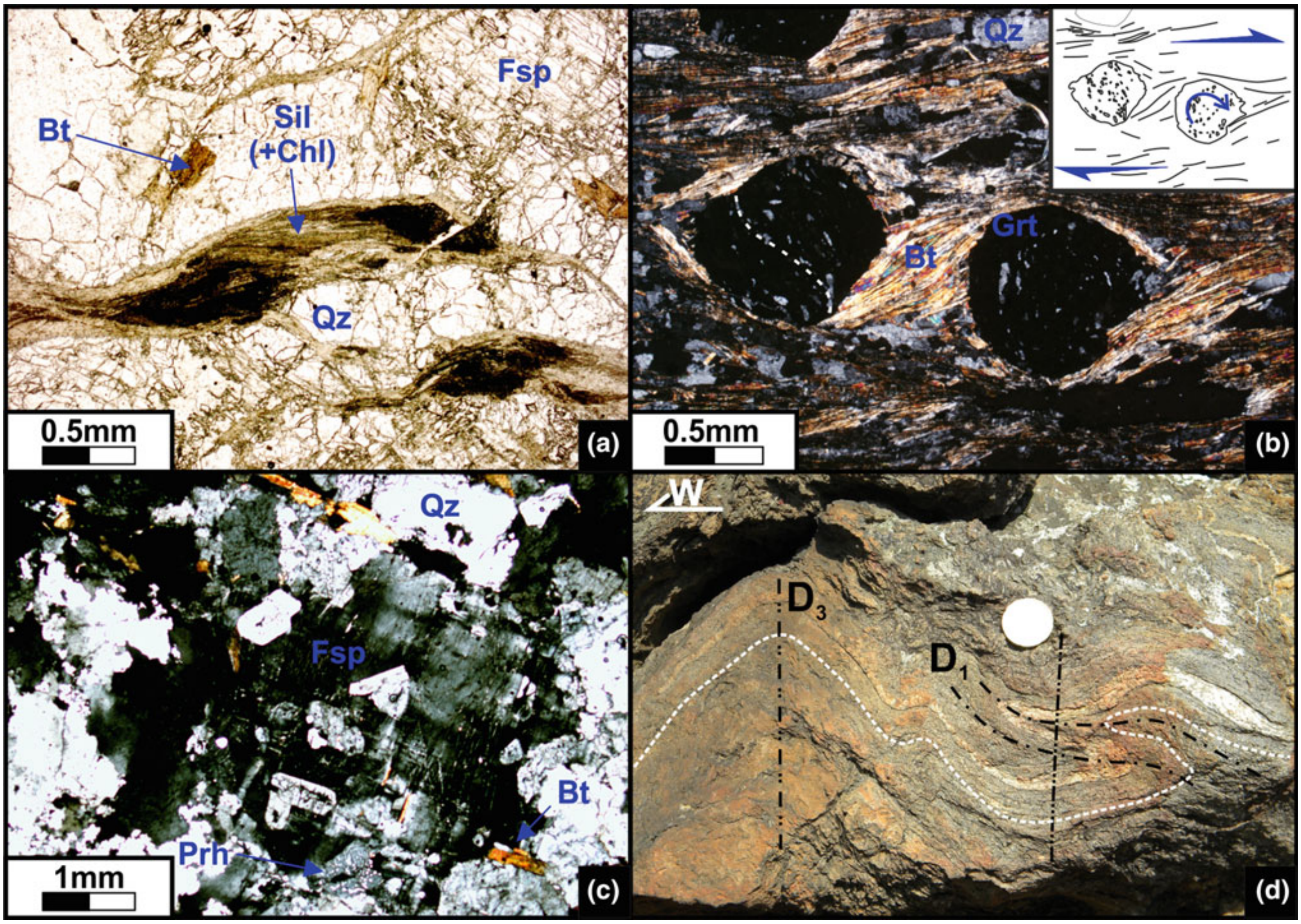

Fig. 7.6 Petrographic and structural representative features of the Abrantes-Tomar sector (Bt—biotite; Sil—sillimanite; Chl—chlorite; Fsp—feldspar; Qz—quartz; Grt—garnet; Prh—prehnite): **a** Sillimanite partially retrograded to chlorite in gneisses of the S. Pedro de Tomar Complex (parallel nicols); **b** Syn-tectonic poikilitic garnets in micaschists of the Junceira-Tramagal Unit, showing dextral synthetic spinning (crossed nicols); **c** Deformed plagioclase crystal of the Tramagal Granite (crossed nicols); **d** Refolded D_1 recumbent fold in micaschists of the Junceira-Tramagal Unit

7.2.2.3 Couço Dos Pinheiros Orthogneiss

The Couço dos Pinheiros Orthogneiss is a strongly stretched N-S body (Fig. 7.5), whose gneissic texture is composed of millimetric-thick felsic-rich layers (quartz and feldspars *s.l.*) and iron-magnesium rich silicates. The presence of sigma shaped K-feldspar porphyroblasts and strongly stretched quartz ribbons indicate an intense ductile dextral deformation. The gneiss is intruded by less deformed felsic coarse-grained dykes, possibly with similar ages to those described in the micro-granitic dykes cutting the Junceira-Tramagal Unit.

The origin and age of the Couço dos Pinheiros Orthogneiss is unknown. The petrographic and structural similarities with the S. Pedro de Tomar Complex suggest a pre-orogenic origin for this orthogneiss and a Neoproterozoic-Lower Cambrian age could be considered. However, an Ordovician to Devonian age should not be excluded, because similar ages were obtained in the pre-orogenic magmatism of northern Finisterra sectors previously described.

7.2.3 The Berlengas Archipelago Sector

The Berlengas Archipelago is composed of granites, gneisses and micaschists. It was considered a “suspect” terrane due its position W of the Lusitanian Basin (Fig. 7.1b; Ribeiro et al. 1991). The similarities with the lithotypes of Abrantes-Tomar sector led us to consider this archipelago as part of the Finisterra Block. In the Farilhões and Forcadas islands outcrops an anatectic complex with gneisses, migmatites and micaschists, while in the Berlengas, Estelas and

Medas islands a pink granite is the most representative lithotype (Fig. 7.1d; Valverde Vaquero et al. 2010a, b; Bento dos Santos et al. this volume).

The Farilhões metatexites highlight a HT metamorphism (sillimanite zone) with Upper Devonian age (377 ± 1 Ma; U–Pb, TIMS in monazites; Valverde Vaquero et al. 2010a, b; Bento dos Santos et al. this volume). Some relics in these metatexites show previous prograde metamorphism reaching granulite facies (P = 8.6 ± 1 kbar; T = 915 ± 50 °C; Bento dos Santos et al. 2010; this volume). Inhered zircon populations indicate a Neoproterozoic para-derived protolith (Bento dos Santos et al. this volume).

The Berlengas granite (Fig. 7.1d) was initially considered of Permian age (280 ± 15 Ma; ^{87}Rb/^{86}Sr in whole rock; Priem et al. 1965), but recent geochronological data indicates a Pennsylvanian age (307.4 ± 0.8 Ma; U–Pb, ID–TIMS in monazite and zircon; Valverde Vaquero et al. 2010a, b), only affected by Tardi-Variscan deformation (Fig. 7.1d; Ribeiro et al. 1991; Dias et al. 2017a).

7.3 Structure and Metamorphism

The sectors described above share a common structural framework characterized by a predominant N-S Variscan trend parallel to the PTFSZ, with NNW-SSE deflections in the vicinity of Porto and Abrantes (Fig. 7.1c). The Abrantes inflection, between Martinchel and Tramagal (Fig. 7.5), is related to a deca kilometric-scale sheath fold that resulted from the interaction between the Tomar-Badajoz-Cordoba Shear Zone (TBCSZ) and the PTFSZ (Ribeiro et al. 2009; Moreira et al. 2011, 2013). The Porto inflection is ascribed to the strike irregularities of the PTFSZ, which generated a restraining band (Ribeiro et al. 2013). Both inflections are compatible with an early activity of the PTFSZ, at least since the beginning of the Variscan Orogeny (Dias and Ribeiro 1993).

Two ductile Variscan deformation episodes (D_1 and D_2) are interpreted as a progressive tectonic process (Ribeiro et al. 1995; Chaminé 2000; Ribeiro et al. 2013) that affects all sectors and units, with the exception of the younger Albergaria Unit, which does not show the D_1 episode (Ribeiro et al. 2013). Frequently, the D_1 and D_2 ductile structures are overprinted by a brittle to brittle-ductile deformation event (D_3).

The D_1 episode consists of recumbent West quadrant facing folds, with low dipping hinges and a pervasive S_1 foliation, being expressed in all the Finisterra sectors (Figs. 7.6d and 7.7; Pereira et al. 1980, 2007; Ribeiro et al. 1980, 1995, 2013; Chaminé 2000; Ferreira Soares et al. 2007; Moreira et al. 2016a; Moreira 2017). Several features are considered coeval with the D_1 tectonic episode:

- The extremely flattened garnets in the Espinho Unit developed in the HT sillimanite zone (P = 4–5 kbar; T = 700 ± 50 °C; Fernández et al. 2003);
- The early metamorphic ages in the same unit (ca. 360 Ma; Almeida et al. 2014);
- The early HT migmatites in the Abrantes-Tomar sector (Moreira 2017);
- The sillimanite zone metamorphism of the Farilhões migmatites (ca. 380 Ma; Bento dos Santos et al. this volume);
- The Upper Silurian-Devonian magmatism of Lourosa Unit (ca. 420 Ma; Chaminé et al. 1998);
- The Late Silurian-Devonian metamorphic overgrowths in inherited zircon (Fig. 7.4; Pereira et al. 2010; Almeida et al. 2014).

This event took place before the deposition of the Frasnian-Serpukhovian black shales of the Albergaria Unit where D_1 structures are absent (Ribeiro et al. 2013). However, a previous Cadomian episode cannot be excluded (Ferreira Soares et al. 2007; Ribeiro et al. 2013).

The D_2 episode is marked by folds with an associated East dipping pervasive S_2 cleavage (sometimes mylonitic), subparallel to the PTFSZ (Fig. 7.7). The presence of a sub-horizontal to low plunging X_2 stretching mineral lineation highlights the dominant dextral transcurrent component (Ribeiro et al. 1980, 2013; Chaminé 2000; Moreira et al. 2016a). The intensity of the D_2 deformation increases eastward towards the PTFSZ where the D_1 structures are often transposed (Chaminé 2000; Moreira et al. 2016a). The Finisterra Block units are always bounded by D_2 shear zones.

The D_2 episode generated a staurolite zone HT metamorphic paragenesis in the Espinho Unit (with garnet overgrowth and staurolite porphyroblasts—P = 3–6 kbar; T = 600 ± 30 °C; Fernández et al. 2003), in the Lourosa Unit migmatites (garnet + sillimanite + K-feldspar + biotite ± muscovite + melt assemblage—P = 8 ± 0.7 kbar; T = 730 ± 25 °C, Acciaioli et al. 2003; Munhá et al. 2008) and in the micaschists of the Junceira-Tramagal Unit (syn-kinematic growth of garnet, with poikilitic structures, and staurolite Fig. 7.6b; Moreira 2017). The D_2 metamorphic event partially resets the previous D_1 HT metamorphic event (Fernández et al. 2003; Moreira et al. 2016a; Moreira 2017). The D_2 tectono-metamorphic event is considered Mississippian in age (ca. 340–315; Pereira et al. 2010; Almeida et al. 2014). The syn-tectonic Carboniferous Tramagal granite (Abranches and Canilho 1981/82) is coeval with the D_2 event (Fig. 7.6c; Romão et al. 2013, 2016; Moreira et al. 2016a). However, this deformation episode do not affect the granitic dykes (318.7 ± 1.2 Ma; Pereira et al. 2010) that are intrusive in the Junceira-Tramagal Unit.

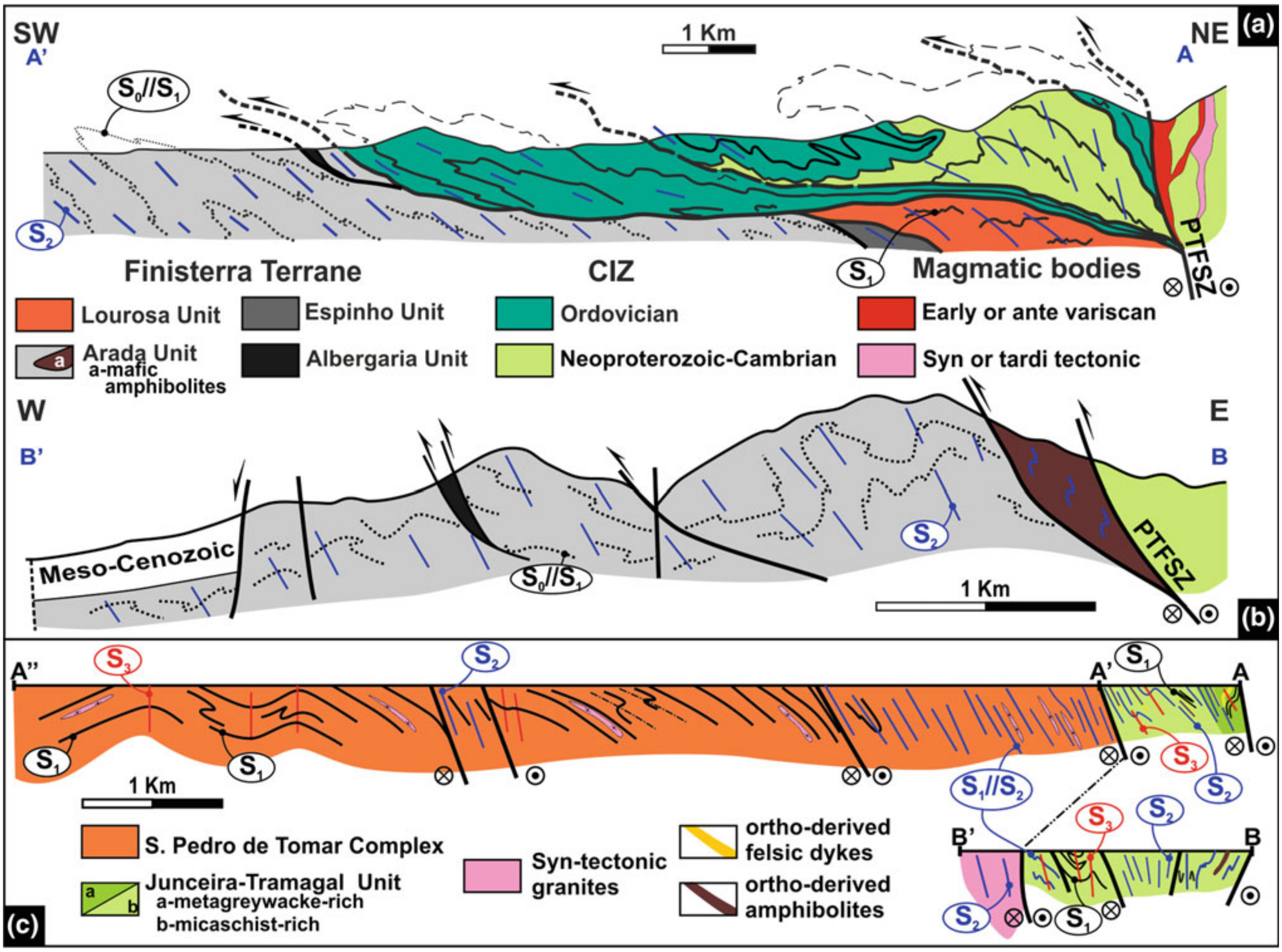

Fig. 7.7 Simplified cross-sections in the Finisterra block: **a** Main structural features of Porto-Espinho-Albergaria sector (see location in Fig. 7.3; adapted from Pereira et al. 2007); **b** Main structural features of Coimbra sector (see location in Fig. 7.3; adapted from Ferreira Soares et al. 2005; Machado et al. 2011); **c** Main structural features of Abrantes-Tomar sector (see location in Fig. 7.5)

The last Variscan deformation episode (D_3) is characterized by the development of folds subparallel to the PTFSZ and faults, generated in brittle-ductile to brittle conditions, frequently associated with the reactivation of D_2 N-S shear zones or the top-to-SW thrusts generated during D_1/D_2 (Ribeiro et al. 2013; Moreira 2017). In the Abrantes-Tomar sector (Moreira 2017) the intensity of the D_3 folds increases towards the PTFSZ, where the open D_3 folds become tight slightly W vergence and with a weak low-grade axial planar cleavage (Fig. 7.7c).

The D_3 deformation event is constrained by the 310–305 Ma Ar–Ar ages obtained in micas of the para-derived rocks of the Espinho and Lourosa Units (Acciaioli et al. 2003; Munhá et al. 2008; Gutiérrez-Alonso et al. 2015) and the 295 Ma of the late-tectonic Tancos, Castelo do Queijo and Madalena-Lavadores granites (Neves et al. 2007; Martins et al. 2011, 2014). However, the Madalena-Lavadores granite is affected by brittle N-S dextral faults (Ribeiro et al. 2015) that result from Late Variscan and/or Meso-Cenozoic tectonic deformations.

7.4 Distinctive Features of Finisterra Block

The individualization of a lithospheric terrane must be supported by stratigraphic, tectonic, metamorphic and magmatic data, emphasizing a distinct geodynamical evolution (Coney et al. 1980). In the author's opinion the Finisterra Block fulfil these conditions because (Fig. 7.8):

(i) It has its own tectonostratigraphic succession composed of:
   - Neoproterozoic-Lower Cambrian high-grade assemblage with a basal gneissic-migmatite

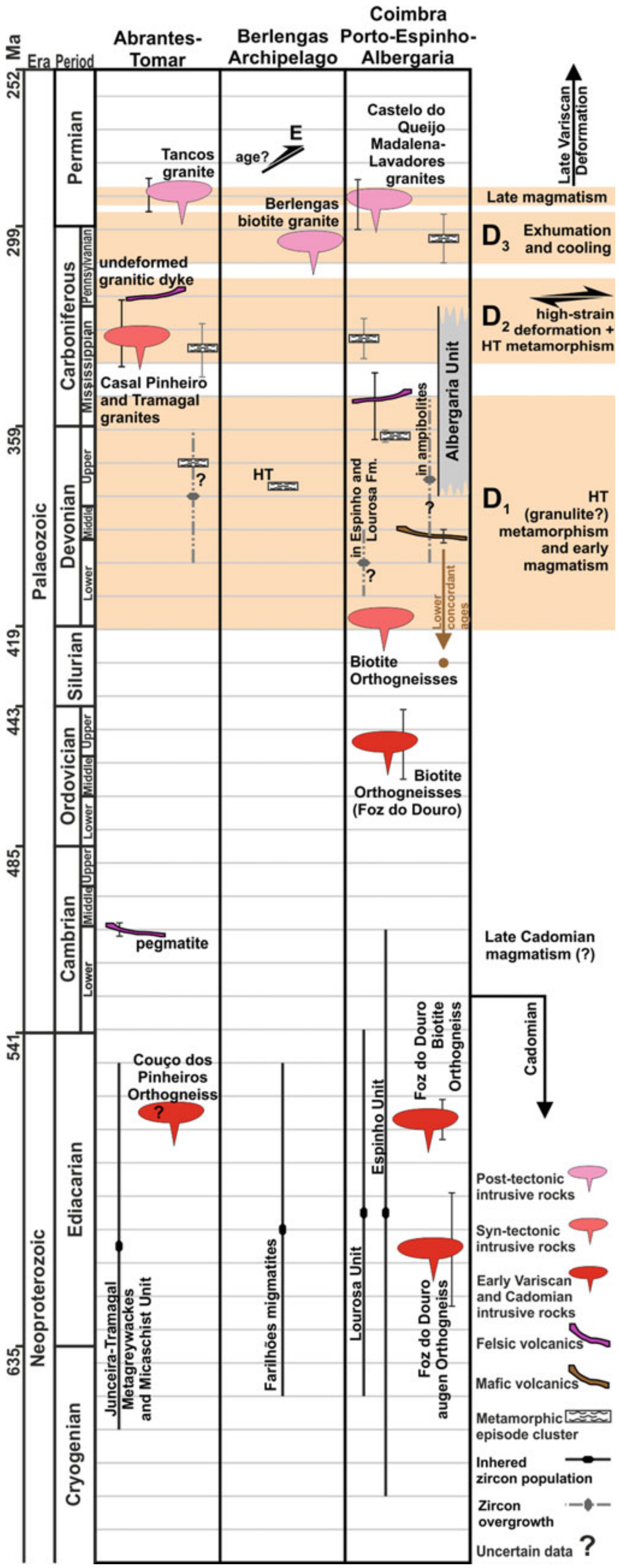

Fig. 7.8 Geological and geochronological synthesis of Finisterra block (see text for references)

complex (Foz do Douro Gneiss, Farilhões, S. Pedro de Tomar and Lourosa Units) and an upper staurolite-garnet-micaschists succession (Espinho and Junceira-Tramagal Units);
- A low-grade assemblage, where the Lower Devonian-Carboniferous Albergaria Unit is discordant over the more deformed and metamorphosed Neoproterozoic Arada Unit.

(ii) The high-grade assemblage shows predominance of Archean, Paleoproterozoic and Neoproterozoic detrital zircon populations. Some lithotypes show the lack of Mesoproterozoic ages (Fig. 7.4), which is a distinctive feature of the North Gondwana margin (Fernández-Suárez et al. 2002; Linnemann et al. 2008; Pereira et al. 2008, 2011, 2012a, b; Talavera et al. 2012; Orejana et al. 2015). However, the presence of Mesoproterozoic zircons in some of the samples (Fig. 7.4; Pereira et al. 2010; Almeida 2013; Almeida et al. 2014) indicates a more complex evolution of these units, with different sources for the clastic sediments of the Finisterra Block. Moreover, the presence of rare (and dubious?; Pereira et al. 2010) Ordovician and Silurian detrital zircons (Pereira et al. 2010; Almeida et al. 2014) could indicate that part of these units are Palaeozoic.

(iii) The Lower Cambrian carbonate sedimentation typical of the OMZ (e.g. Oliveira et al. 1991) and the Ordovician siliciclastic sedimentation recognized in the CIZ (e.g. Dias et al. 2013) are not recognized in any of the Finisterra tectonostratigraphic units.

(iv) The mafic and ultramafic Silurian/Devonian magmatism with intra-plate to MORB geochemistry interlayered in high-grade and Arada Units (Fig. 7.8; e.g. Noronha and Leterrier 2000; Silva 2007; Almeida et al. 2014) is not observed in the Iberian Terrane (e.g. Mata and Munhá 1990; Sánchez-García et al. 2008; Pedro et al. 2010).

(v) The low anchizone marine black shales and siltstones of the Albergaria Unit with Laurussia-type acritarch assemblages of Frasnian-Serpukhovian age (Chaminé et al. 2003b; Machado et al. 2008, 2011) are not recognized, neither in the Iberian Terrane nor in the South Portuguese Terrane. Indeed:
- the lack of marine sedimentation during Frasnian is one of the distinctive features of Iberian Terrane (e.g. Oliveira et al. 1991; Dias et al. 2013; Moreira and Machado this volume), although continental successions with similar ages are found in the lower parauthochton of Galiza-Trás-os-Montes Zone (GTOMZ; Martínez-Catalán et al. 2008).
- in Pulo do Lobo Domain of the South Portuguese Terrane, the marine sedimentation with Frasnian acritarch assemblages have Avalonia affinities (Oliveira et al. 2013; Pereira et al. 2018).

(vi) An Eo-Variscan HT metamorphic event (Fig. 7.8) is recognized in the high-grade tectonostratigraphic units of the Finisterra Block (ca. 420–350 Ma). This event could explain the pre-Carboniferous HT paragenesis observed in Espinho Unit (Fernández et al. 2003), with stretched garnets representative of extremely HT metamorphism (Ji and Martignole 1994), the Silurian-Devonian zircon overgrowths observed in these high-grade units (Pereira et al. 2010; Almeida 2013; Almeida et al. 2014), the metamorphic ages obtained in Farilhões metatexites (ca. 380 Ma; Valverde Vaquero et al. 2010a, b; Bento dos Santos et al. this volume) and in Espinho Unit (ca. 360 Ma; Almeida 2013; Almeida et al. 2014). This HT metamorphic event is not recognized in the Iberian Variscides, where similar ages are only found in the high pressure (HP) metamorphism in the OMZ (Moita et al. 2005) and the HP-granulitic metamorphism of the GTOMZ (e.g. Gómez Barreiro et al. 2007; Mateus et al. 2016; Puelles et al. 2017).

(vii) The Eo-Variscan Silurian magmatism recognized in the Lourosa Unit (ca. 420 Ma; Chaminé et al. 1998) is absent in the Iberian Terrane.

(viii) There is also a strong structural contrast between the Finisterra block and the Iberian Terrane. The oldest D_1 deformation of the Finisterra Block, although highly disturbed by the Carboniferous tectono-metamorphic events, shows N-S oriented recumbent folds with top-to-W transport and rooted in the PTFSZ (Fig. 7.7). Such geometry has no equivalent in the Iberian Terrane, where a NW-SE general trend prevails during early episodes of deformation (Fig. 7.1b; Dias et al. 2013, 2016; Moreira et al. 2014). This early deformation episode is considered contemporaneous of the Silurian-Devonian Finisterra metamorphic event.

Since the Carboniferous, the Finisterra Block and Iberian Terranes share a common geodynamical evolution:

- The Mississippian D_2 HT metamorphic event of Finisterra is synchronous of the HT event described in the Iberian Terrane (Bea et al. 2006; Castiñeiras et al. 2008; Pereira et al. 2012c), where a dextral shearing related to the D_2 evolution of PTFSZ is also observed (Ribeiro et al. 2014; Dias et al. 2017b; Moreira and Dias 2018);
- In the Pennsylvanian, the Finisterra and Iberian Terranes were both pervasively deformed by regional D_3 shear zones (Gutiérrez-Alonso et al. 2015);
- The Late Pennsylvanian sediments of the Buçaco Basin, located in the western border of CIZ near the Finisterra Block (Fig. 7.3), show some Silurian-Devonian and Mesoproterozoic inhered zircon populations (Dinis et al. 2012). The absence of such zircons ages in the CIZ, led to

propose a long source for such populations (Dinis et al. 2012). An alternative proposal is to consider that these sediments were fed by both Finisterra and Iberian Terranes;

- The Upper Pennsylvanian-Permian granitic magmatism is represented in the Finisterra Block (e.g. the Tancos, Castelo do Queijo and Madalena-Lavadores; Figs. 7.1c and 7.2; Neves et al. 2007; Martins et al. 2011, 2014) and in the Iberian Terrane (e.g. Pinto and Andrade 1987; Sant'Ovaia et al. 2013).

7.5 The Finisterra Block in the Context of the European Variscides

The geodynamics of the Finisterra Block cannot be dissociated from the evolution of the European Variscides. However, the continuity of the narrow Finisterra Block is not obvious because its boundary with the Iberian Terrane is marked by a lithospheric shear zone (the PTFSZ) and is separated from the Central European Variscides by the Ibero-Armorican Arc (Dias et al. 2016), which was disrupted during the opening of the Atlantic Ocean. In spite of these difficulties, the main geological features of the Finisterra Block support correlations with the Léon Domain and the Mid-German Crystalline Rise (MGCR), in a similar way to what has already been proposed (Mateus et al. 2016).

7.5.1 The Léon Domain

The Léon Domain (also called Léon-Normanian Domain; Ballèvre et al. 2009) is the northernmost domain of the Armorican Massif (Fig. 7.9a; Ballèvre et al. 2009; Faure et al. 2010), whose "exotic" nature was emphasized long ago (Balé and Brun 1986; Le Corre et al. 1989). The boundary between the Léon and the Armorican domains (Fig. 7.9b) is considered either in the Elorn fault (Ballèvre et al. 2009) or in the Le Conquet-Penzé Shear Zone (Faure et al. 2010). The highly deformed Precambrian and Palaeozoic rocks are structured in a complex stack of nappes as follows (Fig. 7.9c, d; Faure et al. 2005, 2010; Schulz et al. 2007; Ballèvre et al. 2009):

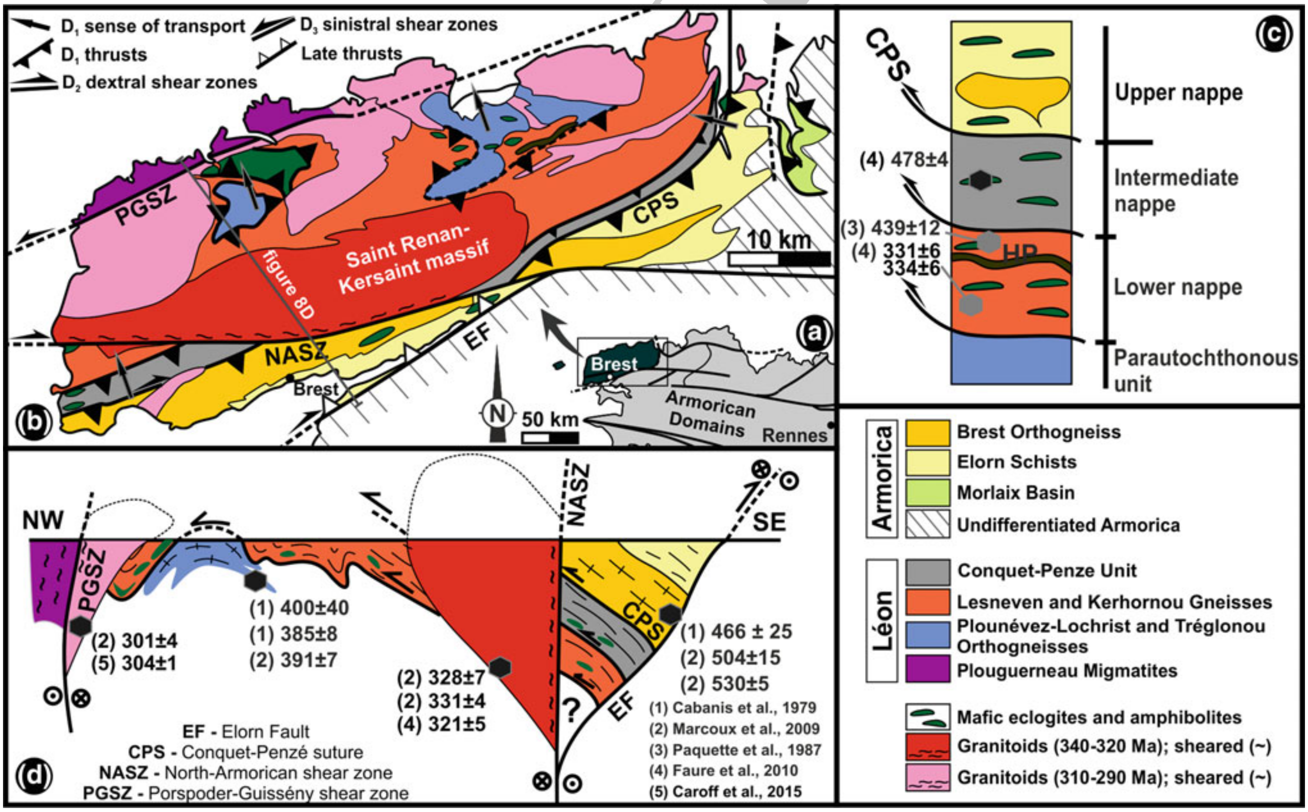

Fig. 7.9 The Léon Domain geological setting: **a** The Léon block and its geological relationship with the Armorican Domain (adapted from Ballèvre et al. 2009); **b** Simplified geological map (adapted from Faure et al. 2010; Schulz 2013); **c** Simplified tectonostratigraphic nappe stack organization (see text for geochronological references); **d** Simplified cross-section (see text for geochronological references; adapted from Ballèvre et al. 2009; Faure et al. 2010; Schulz 2013)

- A parautochthonous unit of paragneisses intruded by the Lower-Middle Devonian Plounevez-Lochrist and Tréglonou Augen orthogneisses (ca. 400–380 Ma; Fig. 7.9 d; Cabanis et al. 1979; Marcoux et al. 2009), affected by intense migmatization during the late Carboniferous (ca. 320–310 Ma; Schulz 2013).
- A lower nappe consisting of garnet-sillimanite gneisses and micaschists (Lesneven and Kerhornou gneisses) with a Proterozoic para-derived protolith (Schulz et al. 2007; Schulz 2013), as well as mafic tholeiites (amphibolites, pyroxenites, serpentinites and eclogites; Balé and Brun 1986; Faure et al. 2010). Eclogite metamorphism of Silurian age (439 ± 12 Ma; Fig. 7.9c; Paquette et al. 1987), Upper Mississippian HT migmatization (ca. 335–330 Ma; Faure et al. 2010) and/or Pennsylvanian (ca. 310–300 Ma; Schulz et al. 2007) ages have been described.
- An intermediate nappe, where biotite-garnet-staurolite micaschists (Conquet-Penzé Micaschists) with a Neoproterozoic protolith and Carboniferous metamorphism (ca. 340–305 Ma; Schulz et al. 2007; Faure et al. 2010; Schulz 2013) predominates. Metacherts, quartzites, conglomeratic lenses and Ordovician amphibolites and Early Ordovician meta-gabbros (Fig. 7.9c; Faure et al. 2010) are also present.
- An upper nappe represented by the Late Proterozoic Elorn Schists (greenschists facies; Ballèvre et al. 2009; Faure et al. 2010), which were intruded by the Cambrian-Early Ordovician Brest orthogneiss with granodiorite composition (Fig. 7.9c, d; Deutsch and Chauris 1965; Cabanis et al. 1979; Marcoux et al. 2009). The Elorn Schists are ascribed to the Armorican Massif basement (Faure et al. 2010).

Two magmatic events took place during the Carboniferous:

- The oldest (340–320 Ma; Cabanis et al. 1979; Faure et al. 2010; Marcoux et al. 2009; Le Gall et al. 2014) composed of calc-alcaline granites and granodiorites (Balé and Brun 1986);
- The youngest (310–290 Ma; Cabanis et al. 1979; Marcoux et al. 2009; Caroff et al. 2015), located in the northern sectors, consisting of sub-alkaline granitoids (Balé and Brun 1986).

Three main tectono-metamorphic events affect the Léon Domain, generating an ENE-WSW to NE-SW global trend (Fig. 7.9b). The early event (D_1) is linked to the emplacement to NNW of nappes (Fig. 7.9b, d; Faure et al. 2010; Balé and Brun 1986). The HP metamorphism registered in the lower nappe is considered previous to the D_1 episode (Bradshaw et al. 1967; Faure et al. 2010), so constraining the timing of this episode to Late Silurian (?)-Devonian.

The HT D_2 episode, which deeply reworks the D_1 fabrics (Balé and Brun 1986; Le Corre et al. 1989; Faure et al. 2005; 2010), is associated to the E-W dextral North-Armorican shear zone (NASZ; Fig. 7.9b; Balé and Brun 1986; Schulz et al. 2007; Faure et al. 2010) and reactivate the Elorn Fault (Faure et al. 2005). This episode is coeval of the Mississippian HT metamorphic event (Schulz et al. 2007; Faure et al. 2010, Schulz 2013) and the first plutonic intrusion (ca. 340–320 Ma). In the lower nappe, where the D_2 is weaker, the D_2 migmatization and melting postdates the eclogite metamorphism (Faure et al. 2010).

The D_3 episode is restricted to the northern sectors (Fig. 7.9b; Le Corre et al. 1989; Marcoux et al. 2009; Caroff et al. 2016). It is closely linked to the NE-SW Porspoder-Guissény sinistral shear zone (Fig. 7.9b, c; Le Corre et al. 1989), which controls the second episode of magmatism and the Plouguerneau migmatites (Fig. 7.9; Ballèvre et al. 2009; Caroff et al. 2015). The metamorphic ages obtained in the migmatites (ca. 330 Ma—U–Pb in monazites, Marcoux et al. 2009; 311 ± 14 Ma; Schulz 2013) and in the mylonites of the Porspoder-Guissény shear zone (293 ± 3 Ma—Ar/Ar in muscovites, Marcoux et al. 2009) constrain this deformation episode between 330 and 290 Ma, which seems to indicate that the migmatization was initiated during D_2 episode.

7.5.1.1 The Mid-German Crystalline Rise

The Mid-German Cristaline Rise (MGCR; Fig. 7.10a; sometimes also called Mid-German Cristaline High) forms the northern sector of the Saxo-Thuringian Domain. It is mostly composed of medium- to high-grade gneisses, migmatites and plutonic rocks, exposed in small basement outcrops with general NE-SW trend (Ruhla—Fig. 7.10b, Kyffhäuser—Fig. 7.10c, Spessart, or Odenwald Crystalline Complexes; Fig. 7.10d; Zeh and Will 2010).

The metamorphism reaches HT conditions (amphibolite-granulite facies) during the Mississippian (340–320 Ma; Nasir et al. 1991; Todt et al. 1995; Will and Schmädicke 2003; Zeh et al. 2003, 2005). This event is coeval with the emplacement of several plutonic bodies (Reischmann and Anthes 1996; Anthes and Reischmann 2001; Zeh et al. 2005). Older ages were obtained in the Odenwald Crystalline Complex (349 ± 14 and 430 ± 43 Ma; Will et al. 2017), suggesting, at least, one early HT episode associated to magmatism. This complex also contains retrograde eclogites derived from within-plate to MORB basalts geochemical signature (Scherer et al. 2002, Will and Schmädicke 2001, 2003) and a Silurian/Lower Devonian protolith age (Fig. 7.10; 410–400 Ma; Zeh and Will 2010). The HP metamorphism of these eclogites is dated of Upper Devonian

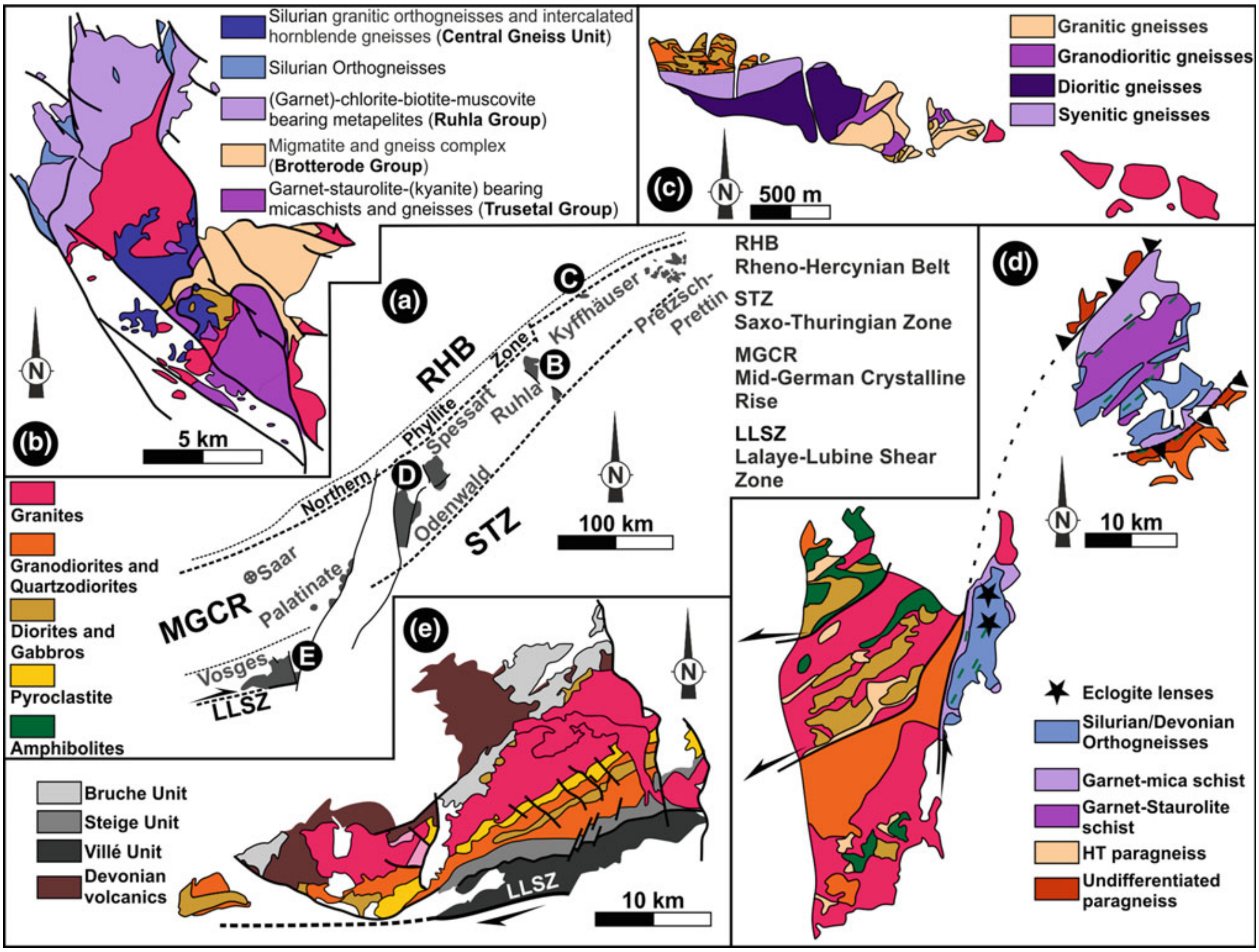

Fig. 7.10 Simplified geological maps of the MGCR (adapted from Zeh and Will 2010): **a** General relation between the crystalline complexes; **b** The Ruhla Complex; **c** The Kyffhäuser Complex; **d** The Spessart and Odenwald Complexes; **e** The Vosges Complex

(357 ± 6 Ma; Scherer et al. 2002), although some resetting could have occurred during the Mississippian retrograde metamorphism (Scherer et al. 2002). Similar metamorphic ages were obtained, not only in the Odenwald Complex (375 ± 5 Ma; Todt et al. 1995), but also in the Ruhla one (357 ± 5 and 352 ± 8 Ma; Zeh et al. 2003), but in these cases the association with the HP metamorphic event is not identified (Zeh and Will 2010). The Upper Devonian metamorphism is coeval with the felsic and mafic-intermediate plutonism (Kirsch et al. 1988; Reischmann and Anthes 1996; Zeh et al. 2005).

The MGCR plutonism is not restricted to the above mentioned events having a wider temporal range: Late Cambrian-Early Ordovician (Anthes and Reischmann 2001), Silurian-Devonian (ca. 420–410 Ma; Dombrowski et al. 1995; Zeh et al. 2003) and Pennsylvanian-Early Permian (310–290 Ma; Anthes and Reischmann 2001). The geological meaning of this plutonism is not treated in the present work.

Detrital zircon populations in the para-derived gneisses and migmatites (Zeh et al. 2001, 2003, 2005; Gerdes and Zeh 2006; Zeh and Gerdes 2010) and some ortho-derived gneisses (Anthes and Reischmann 2001) show two distinct patterns in the Ruhla Crystalline Complex (Fig. 7.10b): samples where Mesoproterozoic populations are absent (Brotterode Group) and samples where the Mesoproterozoic populations are significant (Ruhla Group).

The Vosges complex has a distinct geological history because low-grade metamorphic units are dominant, namely (Fig. 7.10e; Franke 2000; Zeh and Will 2010):

- The Villé Unit, composed of late Cambrian to early Ordovician metapelitic to meta-psammitic schists and quartzites;
- The Steige Unit, a monotonous Ordovician to Silurian shallow marine metapelitic succession, which thrust the Villé Unit;

- The Bruche Unit, a sedimentary and tectonic mélange comprising Frasnian black shales and Fammenian to early Carboniferous shelf and slope sediments, greywackes and conglomerates, as well as calc-alkaline volcanic rocks.

The Bruche Unit is only affected by a Carboniferous tectono-metamorphic event, while the Steige and Villé Units have a previous deformation episode (Skrzypek et al. 2014). All these sequences were intruded by diorites and granites in the Carboniferous.

7.6 The Finisterra-Léon-MGCR Terrane; a Proposal

This proposal is based on the stratigraphic, metamorphic, magmatic and structural comparison between the Finisterra, Léon and MGCR blocks which share remarkable affinities. They are resumed below:

(i) An Eo-variscan plutonic event (ca. 420–360 Ma), represented by Devonian granites, is described in the three domains (Cabanis et al. 1979; Chaminé et al. 1998; Dombrowski et al. 1995; Marcoux et al. 2009). In the MGCR and Finisterra blocks this magmatism is partially coeval with HT amphibolite-granulite metamorphism (ca. 390–360 Ma; Zeh and Will 2010; Bento dos Santos et al. this volume). Late Silurian-Devonian felsic magmatism and metamorphism are rare in European Variscides, a period generally associated with eclogite and granulite facies conditions (Moita et al. 2005; Gómez Barreiro et al. 2007; Ballèvre et al. 2009; Schulz 2013; Mateus et al. 2016; Puelles et al. 2017).

(ii) These early magmatic and HT metamorphic processes were contemporaneous of a complex structural deformation. The older deformation is characterized by N-facing folds and thrusts in the MGCR and Léon domains (Faure et al. 2010; Zeh and Will 2010) and W-facing in the Finisterra Block (Fig. 7.11), a

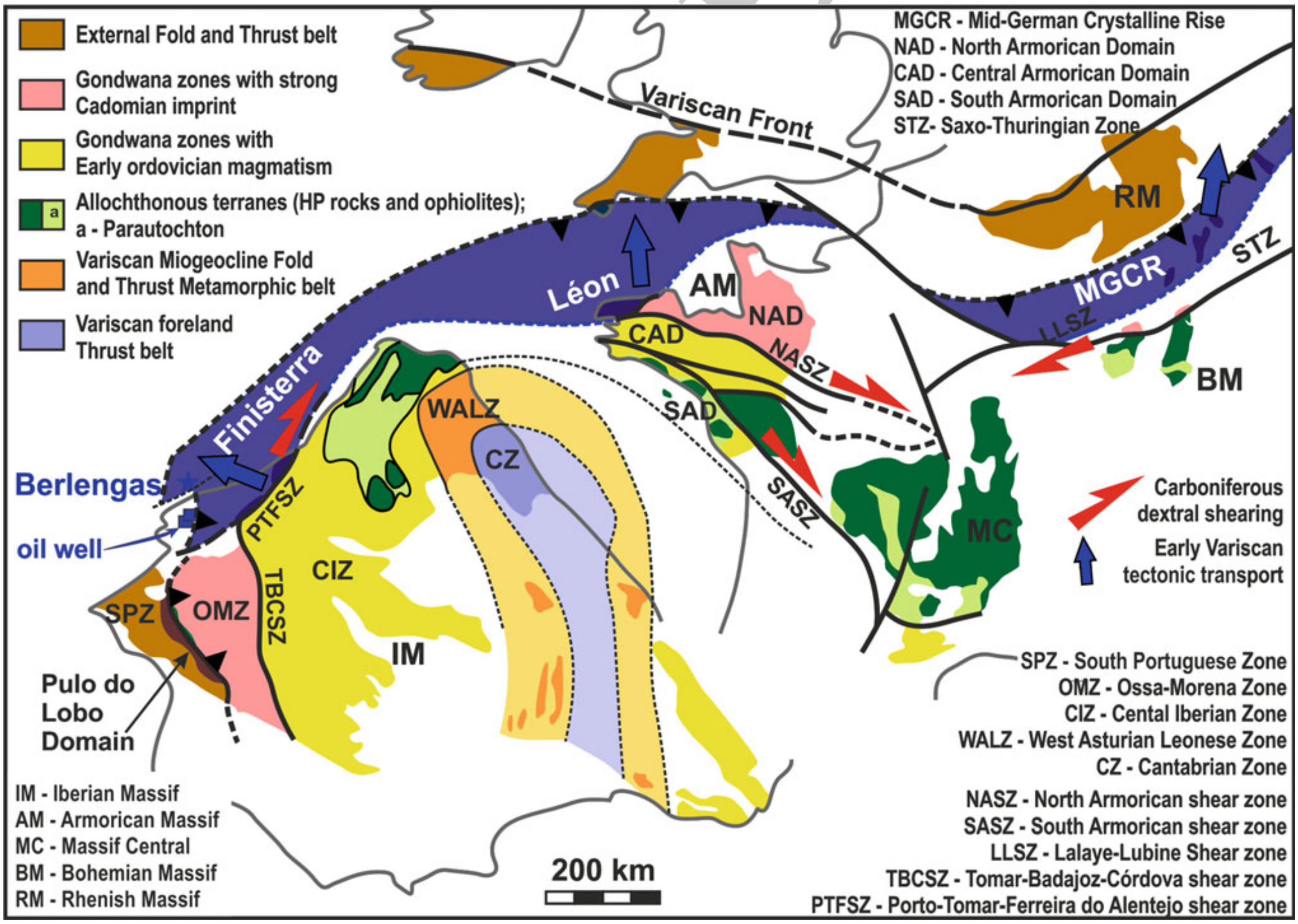

Fig. 7.11 The Finisterra-Leon-MGCR Terrane in the context of the European Variscides (adapted from Ribeiro et al. 2007; Dias et al. 2016; Franke and Dulce 2017)

kinematics compatible with the arcuate structure of Ibero-Armorican Arc (Fig. 7.11; Dias et al. 2016).

(iii) A Silurian-Devonian HP metamorphism with eclogites was also described in the Léon and MGCR domains. These eclogites, which were retrograded during the Carboniferous HT events, are older in the Léon Domain (Silurian; Paquette et al. 1987) than in the MGCR (Upper Devonian; Scherer et al. 2002). Although the age of the MGCR HP rocks are debatable, this suggests a diachronic Variscan subduction during Upper Silurian-Devonian, which may have controlled the early tectono-metamorphic stages of the Finisterra-Léon-MGCR Terrane (Rheic or Rheno-Hercynian Ocean subduction?). Nevertheless, the presence of distinct subductions of two different oceans (e.g. Franke and Dulce 2017) could not be excluded. Eclogites have not been described in the Finisterra Block, probably due to the scarcity of detailed metamorphic studies and/or to the Meso-Cenozoic sedimentary cover of the Lusitanian Basin, which hide a great part of the Finisterra Block (Fig. 7.1b).

(iv) Mafic and ultramafic magmatism, contained in the HT metamorphic units, occurs in all domains, although without well age constrain. The within-plate to MORB geochemistry signature of this magmatism may be the expression of extensional processes during Cambrian-Ordovician or even Silurian related with Variscan Ocean(s) opening;

(v) A similar diversity of lithotypes and the ages of the magmatic and metamorphic events can be found in the Continental Allochthonous Terrane of NW Iberia (Fig. 7.1b; Gómez Barreiro et al. 2007; Mateus et al. 2016). This suggests that this terrane could have been rooted in the Finisterra-Léon-MGCR Terrane and not in Armorica as usually considered (e.g. Ballèvre et al. 2009). This possibility is compatible with the spatial position of Finisterra-Léon-MGCR Terrane in the Ibero-Armorican Arc (Fig. 7.11) and with the SSE nappe transport of the Continental Allochthonous Terrane (Ribeiro et al. 2007).

Putting all things together it seems plausible that the Finisterra, Léon and MGCH blocks were attached together to Gondwana until the Neoproterozoic-Lower Cambrian and were close to Laurussia during the Late Devonian-Lower Carboniferous time. This implies the migration of the Finisterra-Léon-MGCR towards Laurussia as an independent peri-Gondwana Terrane, separated from Gondwana by an ocean realm as indicated by the Silurian-Lower Devonian mafic rocks with MORB signature recognised in the Léon an MGCH Domains. Therefore, the boundaries of these blocks deserve also a close look:

(i) As seen above, the eastern boundary of the Finisterra Block is marked by the PTFSZ (Fig. 7.11; Ribeiro et al. 2007), interpreted as a transform fault with polyphasic deformation at least since the early Variscan Cycle (Ribeiro et al. 2007). Available geophysical data (Silva et al. 2000) suggest that its western boundary is hidden below the Meso-Cenozoic sedimentary cover of the Lusitanian Basin, while its SE continuation is established using the presence of South Portuguese Zone lithotypes found in oil well cores (Benfeito and Monte Gordo; Figs. 7.1b and 7.11; Ribeiro et al. 2013);

(ii) The southern boundary of the Léon Domain is considered the Le Conquet-Penzé Shear Zone whose interpretation is debated, either representing an oceanic suture or the closure of a basin with thinned continental crust (Fig. 7.9b; Faure et al. 2010). Its northern boundary is assumed to represent the Rheic suture zone (Faure et al. 2010).

(iii) The MGCR boundaries are almost totally covered by Permian to Quaternary sediments (Zeh and Will 2010). The contact with the southern Moldanubian Zone corresponds to the Lalaye-Lubine dextral shear zone (LLSZ), superimposed on a previous deformation (Fig. 7.10; Skrzypek et al. 2014). The geodynamical interpretation of this major shear zone is not consensual, seen either as a suture, or as an early Variscan detachment reactivated during Carboniferous (Skrzypek et al. 2014). The northern boundary is not exposed but is indirectly assumed to be placed south of the Northern Phyllite Zone correlated with the Pulo do Lobo Domain of the South Portuguese Zone (Fig. 7.10; Franke and Dulce 2017).

Thus, the northernmost boundary of the Finisterra-León-MGCR Terrane should represent a Variscan Oceanic suture (Fig. 7.11; Rheic and/or Rheno-Hercynian Oceanic Suture?; Franke 2000; Faure et al. 2010; Franke and Dulce 2017). However, its southernmost boundary with Gondwana derived Terranes (Armorica and Iberia) is debatable and two distinct interpretations coexist:

- 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);
- The suture zone of a minor Palaeozoic Ocean (or a stretched continental crust basin) opened during Palaeozoic times, as it was proposed for León Block (Faure et al. 2010).

The first hypothesis could explain the absence of HP rocks in the Finisterra Block and its appearance in León Block and MGCR. In turn, the second one could explain the abundant Ordovician to Silurian mafic and ultramafic rocks with geochemistry similar to MORB to within-plate basalt in all domains (Faure et al. 2010; Zeh and Will 2010; Almeida et al. 2014), as well the Upper Silurian to Devonian HP metamorphic event in León Block and MGCR (Paquette et al. 1987; Scherer et al. 2002).

Since Mississippian, the Finisterra-León-MGCR Terrane and the other peri-Gondwana terranes show similar metamorphic and magmatic events, suggesting a common evolution. This is compatible with the beginning of the collision between Gondwana and Laurentia (Ribeiro et al. 2007; Moreira et al. 2014; Dias et al. 2016). In Mississippian all these terranes where affected by major dextral shear zones (e.g. PTFSZ, NASZ and LLSZ). The pervasive HT metamorphism with melting generation related to the collisional process are superimposed on previous events and almost obliterates 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, also seems to indicate the Northern Gondwana affinities for this composite Terrane. Assuming a possible Cadomian suture in the Espinho Unit, the PTFSZ could be interpreted as a Variscan transform fault reactivating an earlier Cadomian structure, connecting two segments of a Cadomian suture in the TBCSZ and in the northern sector of Finisterra.

Thus, the Finisterra-Léon-MGCR Terrane only has a distinct evolution of Northern peri-Gondwana realm during Early Palaeozoic times (Ordovician to Upper Devonian).

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