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Rifting of the Southwest and West Iberia Continental Margins

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Abstract

The West and SouthWest Margins of Iberia started their formation as intra-continental rifts during initial break up of Pangea in Triassic times. The tectono-stratigraphic record of the Algarve, Alentejo and Lusitanian basins and their offshore prolongation documents the syn-rift, post-rift and passive margin phases of the rifting process as well as three magmatic cycles of tholeiite to alkaline affinities. Although the Ocean-Continent Transition has been investigated using deep ocean drilling, seismics, gravimetry and magnetics its nature and location are still matters of debate. The tectonic inheritance of the Paleozoic Orogeny had great influence in the geometry of the rift basins and development of the Neo-Tethys and Atlantic Oceans intersection. Salt tectonics strongly controlled both syn-rift and post-rift basin tectonics and led to the formation of an allochthonous salt nappe in the Algarve Basin.

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6.1 Introduction

At the end of the Paleozoic, Iberia was part of the Pangea super-continent that amalgamated most of the Earth's continental masses. The relative location of Iberia at the meeting point of the Variscan suture zone between Laurasia, Gondwana, and the western end of the Tethys Ocean, created the necessary conditions for crustal stretching around the Iberia microplate. This crustal stretching event occurred at the beginning of the Meso-Cenozoic Wilson Cycle.

The sinistral oblique movement of Africa with respect to Eurasia, spanning the Early Jurassic to the end of Early Cretaceous, resulted in the propagation of the Tethys Ocean towards the North Atlantic rifts. This allowed for the formation of a sinistral transtensional drift between South Iberia and North Africa and the development of rift basins along Southwest Iberia. The best preserved example of these rift basins is the Algarve Basin (AB) in South Portugal. The lateral prolongation of the AB is named Southwest Iberia Margin (SWIM), and includes its western and southern Ocean-Continent Transition Zone (OCTZ).

The West Iberia Margin (WIM) is the continental façade of the Iberia microplate that experienced Mesozoic extensional tectonics. The Lusitanian Basin (LB), located in West Portugal, is one of the best examples of an exposed rift basins in Europe. The Alentejo Basin, of which two small remnants are documented onshore, is the southern rift basin of the WIM. The location of the OCT along the WIM and the age of continental break-up are still matters of debate. However, it is consensual that oceanic spreading was diachronous, propagated from south to north, and occurred approximately from Barremian to Aptian times in SW Iberia ($\sim$ 128 to 110 Ma) and relatively later (Late Cretaceous) in the Northwest part of the WIM (e.g. Pinheiro et al. 1996; Bronner et al. 2011).

6.2 The Paleozoic Tectonic Inheritance

Terrinha P

The pre-Mesozoic basement of Iberia depicts an arcuate structure, the Ibero-Armorican arc, which can be followed through France and the British Isles into Central Europe (Fig. 6.1). The Ibero-Armorican arc was formed in the Late Paleozoic during the Variscan Orogeny. The best exposed Mesozoic rift basins, the Algarve and the Lusitanian Basins, lie on top of NW-trending orogenic belts, the South Portuguese Zone (SPZ) and the Central Iberian Zone, respectively.

The Algarve Basin lies on top of the low-grade metamorphic marine turbidites of the Baixo Alentejo Flysch group (Carboniferous), in the SPZ. The SPZ is a SW-vergent thin-skinned fold and thrust belt detaching on top of Lower Paleozoic basement. The Lusitanian Basin lies on top of the Central Iberian Zone (CIZ), which is an internal zone of the Variscan Orogen of Iberia chiefly made of low- to high-grade metamorphic rocks and syn- to post-orogenic granites.

A Late Variscan faulting event occurred in Permian times after cratonization of the Variscan Orogen within an irregular megashear zone that extended from the Appalachians through Central Europe towards the Urals (Arthaud and Matte 1975). This event led to the formation of a dense network of left-lateral NNE-SSW strike-slip faults in Northwest Iberia, changing to ENE-WSW to the Southwest (Ribeiro 2002; Dias et al. 2017) (Fig. 6.2). The conjugate set (NW-SE to the Northwest and N-S to the Southwest) is not as pervasive as the main one. Some of these NNE- to ENE-trending faults extend for more than 250 km in length, controlled the intrusion of some late orogenic granites (Sant'Ovaia et al. 2000), and occasionally contain basic and hydrothermal quartz dykes in the CIZ. In Southern Portugal, the Messejana-Plasencia dike (Triassic) intruded along a NE-SW late Variscan fault zone that is >500 km long (Fig. 6.2). The NNE-SSW and ENE-WSW striking faults were reactivated as main extensional faults in the WIM and SWIM during Jurassic and Cretaceous rifting.

6.3 Plate Kinematics and the Ocean-Continent Transition

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The tightest configuration of Pangea as a supercontinent was achieved in the Late Permian at the end of the Variscan Orogeny, after Laurussia became attached to Gondwana through the closing of the Rheic Ocean (e.g. Domeier and Torsvik 2014; Nance et al. 2012; Torsvik et al. 2012) (Fig. 6.1). Iberia, one of the Variscan terranes accreted to Laurussia during the Carboniferous, occupied a key position within the Laurussia-Gondwana boundary, near the transition from continental masses to the Tethys Ocean (Pangea-A configuration; Domeier et al. 2012). However, Iberia's precise pre-rift position with respect to Africa and North America is still under debate (e.g. Olivet 1996; Silva et al. 2000; Sibuet et al. 2012), a significant caveat in Mesozoic paleogeographic reconstructions involving the break-up of Pangea and formation of the North Atlantic and western Neo-Tethys.

The kinematics of Iberia for most of the Mesozoic is not consensual (e.g. Olivet 1996; Srivastava et al. 2000; Sibuet et al. 2004). This is mainly due to lack of data and

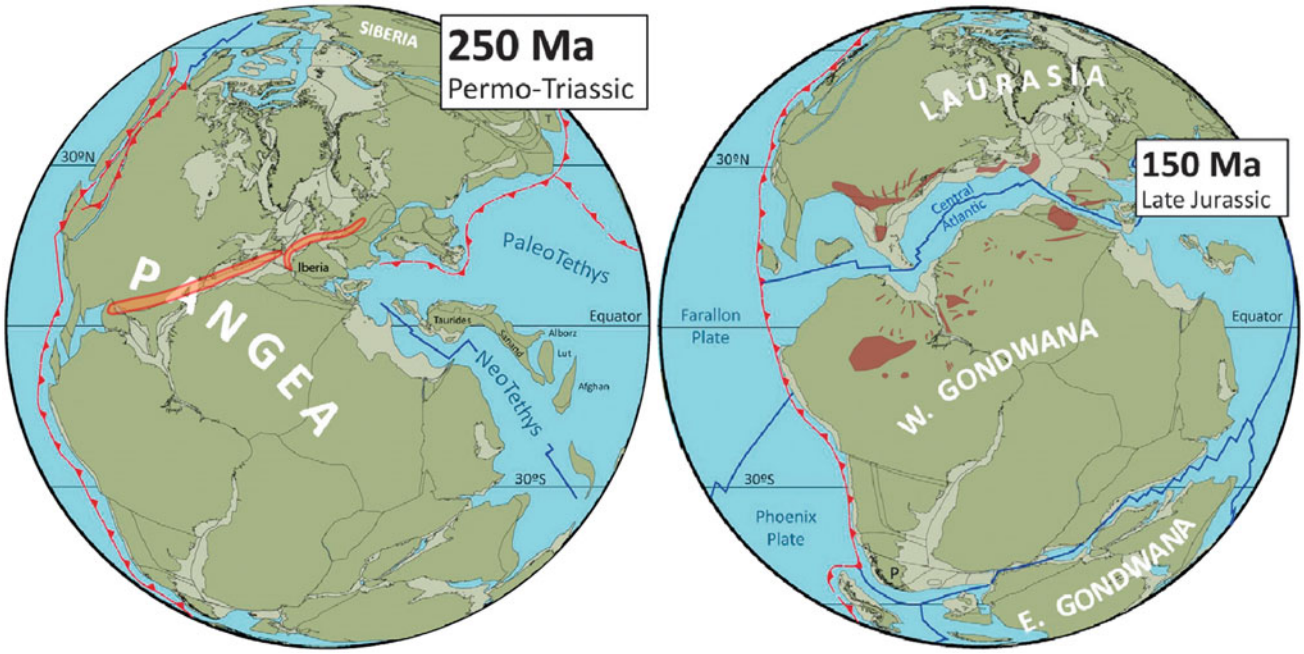

Fig. 6.1 Plate reconstructions for Late Permian (Pangea tightest assemblage) and for Late Jurassic (modified from Torsvik et al. 2012). Red shade in the Permo-Triassic reconstruction locates the Variscan suture of the Rheic Ocean closure (after Nance et al. 2012) and the Ibero-Armorican Arc. Note that the Variscan suture in Iberia coincides with location of the Lusitanian, Alentejo and Algarve rift basins, i.e. Jurassic-Cretaceous propagation of oceanic ridges. The brown features in Late Jurassic reconstruction represent outcrops of dikes and lava flows of the Central Atlantic Magmatic Province (CAMP, see text)

inconsistencies between interpretations, namely when comparing paleomagnetic data and seafloor magnetic anomalies on the WIM (and Bay of Biscay) with geological observations in the Pyrenees and the nature of the lithospheric mantle (Srivastava et al. 2000; Sibuet et al. 2004, 2007; Gong et al. 2008; Jammes et al. 2009; Neres 2013; Neres et al. 2012, 2013; Vissers et al. 2016; Barnett-Moore et al. 2016, 2017; van Hinsberger et al. 2017; Nirrengarten et al. 2016).

The nature and age of marine magnetic anomalies offshore West Iberia (Fig. 6.2) have been central not only for the reconstruction of Iberia kinematics but also to assess the lithospheric structure, evolution of the margin and location of the OCTZ. Uncertainties associated with the nature and position of the oceanic magnetic anomalies are important as they allow for significant differences between proposed locations for the OCT ($\sim$100 km) and of the ages of seafloor spreading (from 145 to 112 Ma). This is shown schematically in Fig. 6.2. Such a caveat has also been tackled by stratigraphic information from onshore and offshore wells, outcrop and seismic data, with the ultimate aim of understanding the absolute ages for key tectono-stratigraphic events, and their geodynamic significance in the whole of Iberia (Alves et al. 2009; Soares et al. 2012 a, b; Alves and Cunha 2018). Of particular interest has been the recognition of major regressive events during tectonic events that are associated with the onset of rifting on the proximal margin (Alves et al. 2009) and subsequent shift of the main rifting axes towards more distal positions on the WIM during continental break-up (Alves and Cunha 2018).

The oldest indisputable magnetic isochron off West Iberia is C33r (79-83 Ma; Gee and Kent 2007) (Fig. 6.2). Hence, 83 Ma (early Campanian) is the oldest reconstruction that can be achieved from the fitting of magnetic anomalies across the North Atlantic. During the Cretaceous Normal Superchron (C34, 83-120.6 Ma; Gee and Kent 2007), until the M0 chron (120.6 Ma; Gee and Kent 2007), ocean-type magnetic anomalies were not generated due to lack of geomagnetic reversals. Whereas in the Central Atlantic, M0 and older M-series chrons are consensually identified offshore Northwest Africa and North America (Klitgord and Schouten 1986; Labails et al. 2010), the identification of the M0 and older anomalies off West Iberia has been a matter of debate. Some authors argue for the existence of M0 adjacently to the J anomaly on the Madeira-Tore Rise (MTR), although with different locations that imply different reconstructions (Olivet 1996; Srivastava et al. 2000). Magnetic anomalies of the M-series were tentatively identified between the MTR and West Iberia and resulted in different interpretations, namely those supporting the presence of a slow-spreading oceanic seafloor (Srivastava et al. 2000) and those in favor of serpentinized exhumed mantle in the same region(s) (Sibuet et al. 2007).

Recent studies have challenged previous interpretations regarding the origin of magnetic anomalies offshore Iberia.

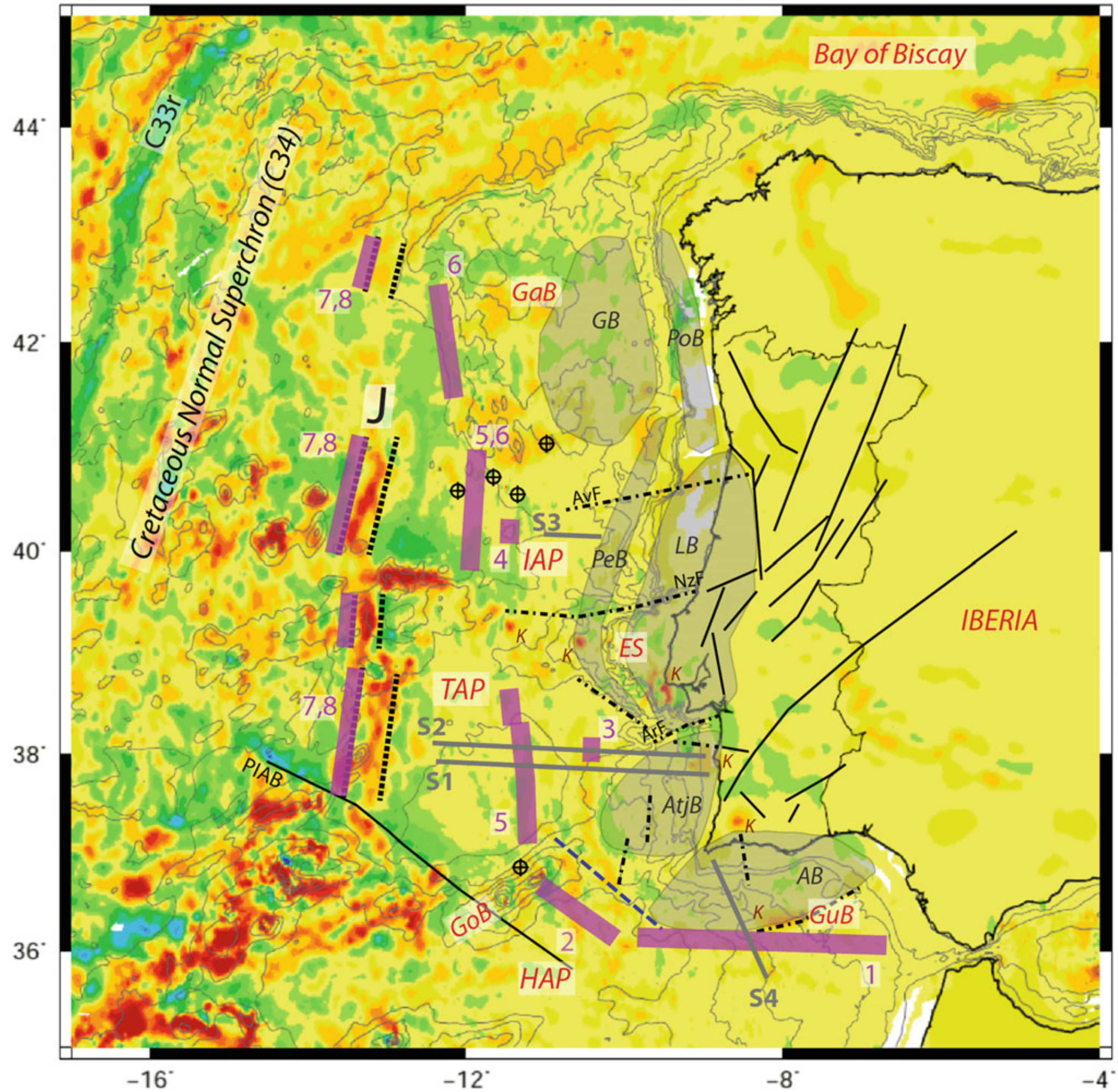

Fig. 6.2 Magnetic anomalies off West Iberia. IAP, Iberia Abyssal Plain; TAP, Tagus Abyssal Plain; CNS, Cretaceous Normal Superchron; 1 to 5, OCT, Ocean-Continent Transition according to 1—Sallarès et al. (2011); 2—Rovere et al. (2004); 3—Afilhado et al. (2008); 4—Dean et al. (2000); 5—Pinheiro et al. (1992); 6—Russel and Whitmarsh (2003); 7—Bronner et al. (2011); 8—Sibuet et al. (2007). K: Upper Cretaceous alkaline magmatism, Sintra, Sines and Monchique complexes from north to south. Late Variscan faults inland (solid black lines) and offshore West Iberia (point-dashed lines); ArF- Arrábida Fault; NzF- Nazaré Faul; AvF- Aveiro Fault; PIAB- Paleo Iberia-Africa Boundary in the Jurassic-Cretaceous transition, according to Rovere et al. (2004). Solid black lines indicate inland Late Variscan faults and point-dashed lines Late Variscan faults offshore. Note that these faults subdivide the WIM into rifting sectors. Blue dashed-point line: Gorringe Bank - Gulf of Cádiz transfer fault. Crossed circles: ODP and DSDP sites. Gray lines S1-S4 locate deep seismic lines in Figs. 6.3 and 6.4. AB- Algarve Basin; AtjB- Alentejo Basin; GB- Galicia Basin; LB- Lusitanian Basin; PeB- Peniche Basin; Pob- Porto Basin

Bronner et al. (2011) suggested that the J anomaly does not result from pure oceanic accretion but from magmatism and underplating, which triggered continental break-up at ~112Ma (early Albian) off central West Iberia. Nirrengarten et al. (2016) agree in that the J anomaly cannot be considered an isochron, suggesting instead that it results from syn- and post-break-up polygenic magmatic events. These authors discard its use in kinematic reconstructions.

The WIM has been intensively studied using geophysical methods such as multi-channel seismic reflection, wide-angle reflection and refraction, magnetic and gravity data. The SWIM and the WIM were first drilled in the scope of the Deep-Sea Drilling Project (DSDP). Later, the Ocean Drilling Project (ODP Legs 103, 149, 173) investigated the nature of the basement and rifting processes at magma-poor margins of slow spreading oceans.

The deep-water WIM can be divided into three different domains: (i) the Deep Galicia Margin to the North, to the west of the Galicia Bank; (ii) the Iberia Abyssal Plain (IAP) south of the Galicia Bank, and; (iii) the Tagus Abyssal Plain (TAP) to the south. The IAP and TAP are separated by the Estremadura Spur, a tectonic and volcanic relief that resulted from a Late Cretaceous magmatic event and Paleogene and Neogene compression (Ramos et al. 2017a; see Alpine orogeny in the WIM, this volume).

In the past, two main hypotheses have been put forward to explain the nature of the WIM's deep domain between thinned continental crust to the East and thickened crust (coinciding with the J anomaly) to the West. This domain is generally designated as the ocean-continent transitional domain, or zones (OCTZ) as no clear ocean-continent boundary can be defined.

Some authors favoured the hypothesis that the OCTZ consists of oceanic crust formed by slow or ultra-slow spreading (Sawyer 1994; Whitmarsh and Sawyer 1996). This hypothesis is consistent with the model proposed by Cannat (1993), with a crust intruded by serpentinized upper mantle with pockets of gabbro and basalts corroborating the drilling of serpentinized peridotite rocks south of the Galicia Bank and in the Iberia Abyssal Plain by DSDP and ODP (Whitmarsh et al. 1998).

It has also been suggested that the OCTZ in West Iberia consists of a narrow zone of thin faulted continental crust underlain and intruded during rifting by serpentinized peridotite (Pinheiro et al. 1992; 1996; Whitmarsh et al. 1990, Whitmarsh and Miles 1995, Whitmarsh and Sawyer 1996). This hypothesis was based on the integrated interpretation of seismic refraction and reflection data, as well as on gravity and magnetic modelling. Based on this interpretation (Pinheiro et al. 1992, 1996) two ODP Legs (149 and 173) were proposed for the OCTZ in the Iberia Abyssal Plain. These drill holes confirmed the existence of an area of serpentinized peridotite, which is wider than previously thought, marking the OCTZ. Some authors have nevertheless raised the possibility that the ODP sites could coincide with a fracture zone or transfer zone and may not represent the OCTZ (e.g. Boillot and Winterer 1988).

Results from wide-angle reflection and refraction modelling on the IAP and TAP (Pinheiro et al. 1992, 1996; Dean et al. 2000; Afilhado et al. 2008) show that the velocity model at the OCTZ is not compatible with upper or lower continental crust. As such, the hypothesis that is more consistent with all geological and geophysical data points out the OCTZ as an area of serpentinized peridotite separating highly thinned continental crust from oceanic crust. The serpentinized upper mantle could have been exposed or underlays a thin oceanic crust that was likely intruded. The serpentinized peridotites probably correspond to exhumed lithospheric mantle and the variations in the velocity structure derived by seismic methods can thus be explained by a variable degree of serpentinization (Pinheiro et al. 1992, 1996; Beslier et al. 1993; Pickup et al. 1996; Discovery 215 Working Group 1998; Dean et al. 2000; Péron-Pinvidic and Manatschal 2009; Afilhado et al. 2008).

Despite the numerous geophysical and geological surveys and deep sea drilling conducted on the WIM, the origin and nature of the crust at the OCTZ and the locations of the oldest unequivocal oceanic crust remain a matter of debate. Although great progress has been achieved towards understanding the deep structure of the OCTZ the seismic resolution still allows for different structural interpretations (Fig. 6.3). What appears to be clear is that both the SWIM and the WIM were formed by a slow rifting process and slow divergent plate movement. Continental break-up of the WIM occurred at the end of the Early Cretaceous in association with the generation of a North Atlantic Mid-Ocean Ridge. The initial oceanic lithosphere in the SWIM formed due to the westwards propagation of the Neo-Tethys as early as in Mid Jurassic times. The SWIM oceanic domain has never been very wide due to the sinistral transtensional movement of Africa with respect to Iberia (see Fig. 6.1; e.g. Torsvik et al. 2012).

The SWIM comprises two different OCTZs, one to the south of the Guadalquivir and Portimão Banks that bound the Algarve Rift Basin and another one to the west of the continent that is well exposed along the Gorringe Bank. The OCTZ to the south of the Algarve Basin comprises a necking domain that varies in width from 25 to 50 km and where thinning is accommodated through ductile deformation of the lower crust, overlain by synthetic southwards tilted blocks of upper crust. A distal domain is made of extremely thinned tilted crustal blocks underlain by serpentinised mantle that establishes the transition to the oceanic crust of the Neo-Tethys Ligurian Ocean (Sallarès et al. 2011, 2013; Martínez-Loriente et al. 2014; Ramos et al. 2017b, Fig. 6.4). The Gorringe Bank has been drilled at ODP Site 120 and investigated through refraction and reflection seismic. It consists of partially serpentinized continental mantle rocks overlain by sediments of Albian through Present age that were thrusted over the Tagus Abyssal Plain. To the East, the Gorringe Bank is overlain by the rifted SWIM continental margin. The Eastern part of the Gorringe Bank is likely separated from the SWIM by a Mesozoic (syn-rift) transfer fault (Pereira and Alves 2013;

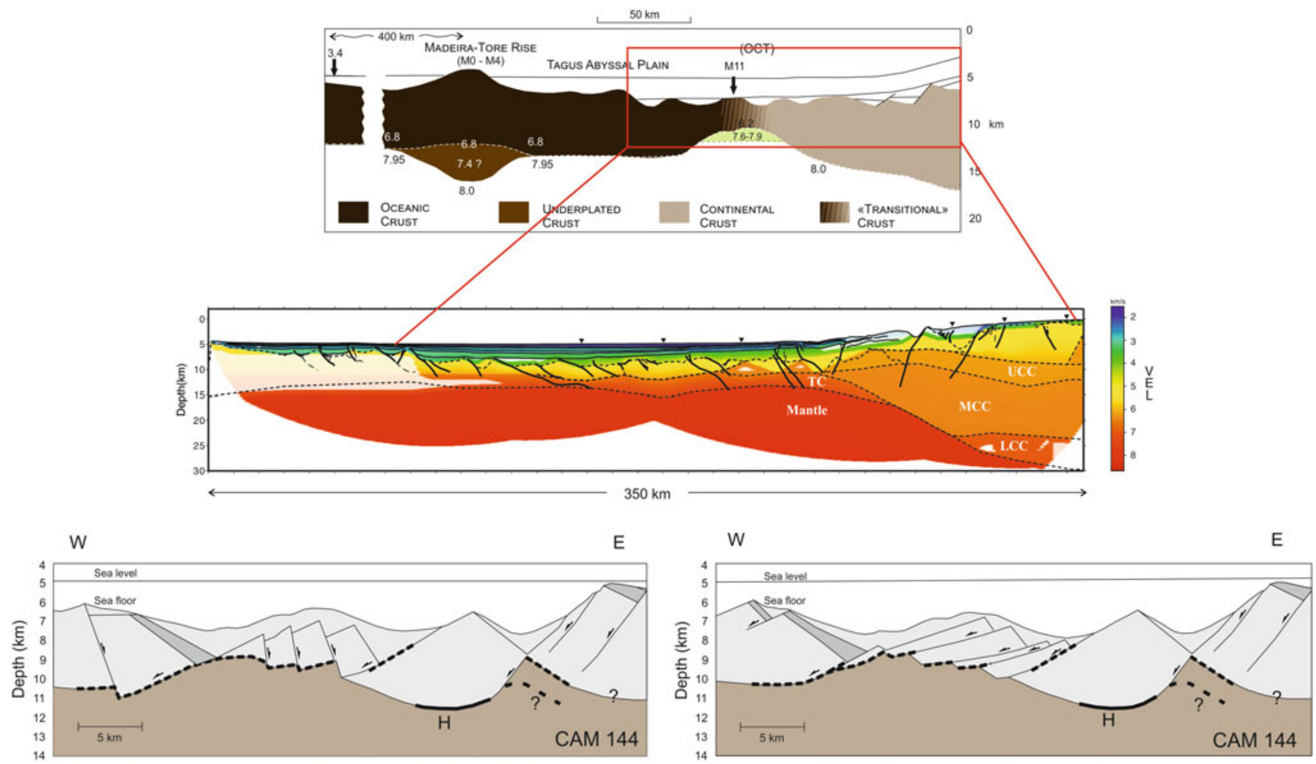

Fig. 6.3 The deep structure of the WIM. Top: structure of the Tagus Abyssal Plain (adapted from Pinheiro et al. 1992); Centre: detailed structure of the continental crust and OCT in the Tagus Abyssal Plain; corresponding to rectangle in top (Afilhado et al. 2008); Bottom: schematic alternative solutions for the crustal structure of the Iberia Abyssal Plain (Dean et al. 2008)

Ramos et al. 2017b, Fig. 6.2). The western part of the Gorringe Bank passes laterally to oceanic crust across the Paleo Iberia-Africa Plate Boundary (PIAB), according to Rovere et al. (2004) (see Fig. 6.2). The PIAB separates an Africa domain of linear magnetic anomalies from an Iberia domain where magnetic anomalies are not evident. The basement extensional fabrics also suggest different extensional fabrics and basement composition on both sides of the PIAB (Rovere et al. 2004; Silva et al. 2017).

At present, the detailed nature of the OCTZ and structure of the crust are still under debate due to the lack of high-resolution seismic refraction and wide-angle profiles (Fig. 6.3). Nevertheless, the rifted margins of the WIM, on the Tagus and Iberia Abyssal Plains, and on the SWIM, display different structures (Pereira et al. 2017). These differences are presumably associated with the different tectonic settings and heritage, rift mechanisms and possibly different mantle compositions.

6.4 Magmatism

Mata J

The lithostratigraphic successions of the WIM and SWIM show evidence for two magmatic cycles preceded by important rifting events. These two magmatic cycles occurred in the Early Jurassic and during the Jurassic-Cretaceous transition, separated by a time gap of ca. 50 My.

6.4.1 Lower Jurassic Magmatism

The first of these cycles is well preserved in the Algarve Basin as part of a volcano-sedimentary complex formed after the deposition of siliciclastic red beds of Triassic age and a Hettangian evaporite-pelite complex. These volcanic rocks, outcropping almost continuously along the W-trending (ca. 150 km) Algarve Basin, were described in detail by Youbi et al. (2003), Martins et al. (2008) and Callegaro et al. (2014). This volcanic complex is part of the Central Atlantic Magmatic Province (see Fig. 6.1 for location; Marzoli et al. 1999).

Volcanic rocks are inter-layered on a marly-carbonate complex and were dated by Verati et al. (2007) using the $^{40}Ar/^{39}Ar$ method on plagioclase. Their results were recalculated by Callegaro et al. (2014), who, by using the decay constant determined by Renne et al. (2009), obtained an age of 199.7 ± 1.4 Ma as the best estimate for Lower Jurassic volcanic rocks. The thickest volcanic sequences occur in the Central Algarve where up to 16 effusive and explosive events can be found intercalated with sediment intervals. These sediments were covered by lava flows when they were still unconsolidated and saturated in water, as shown by the occurrence of peperite, ball and pillow structures (Martins et al. 2008).

Lower Jurassic magmatic rocks are low-Ti tholeiites ($TiO_2 < 1.5\%$) with basic to intermediate compositions. Accordingly, they are characterized by normative quartz and/or hyperstene, a relative rarity of olivine, the presence of pigeonite and interstitial granophyric intergrowths, as well as textural evidence for plagioclase crystallization preceding clinopyroxene (Martins et al. 2008). Lavas do not have primary or primitive compositions (Mg# < 71; Ni < 110 ppm) and are characterized by a small compositional range (59 < Mg# < 71). Clinopyroxene-melt barometry indicates that the crystallization of phenocrysts started at pressures of 0.69 ± 0.11 GPa, corresponding to an average depth of 26 ± 4 km (Martins et al. 2008), which is slightly shallower than the estimated pre-rift Moho depth (Stapel et al. 1996). This indicates that the main fractionation processes occurred at deep crustal levels. However, upper crustal reservoirs must have existed at pressures of 0.35 to 0.05 GPa (Callegaro et al. 2014), suggesting a polybaric character for their crystal fractionation.

As typical of continental tholeiites, Lower Jurassic magmatic rocks in Algarve present: (1) faintly fractionated REE patterns (Ch-normalized La/Yb up to ≈3.5), and (2) multi-elemental patterns characterized by high LILE/HFSE ratios complemented by Nb and P negative anomalies. However, they also show pronounced positive anomalies in Pb.

Most rock samples from Algarve are characterized by a somewhat radiogenic initial $^{87}Sr/^{86}Sr$ (up to 0.70657), which broadly correlates negatively with Nd isotope ratios ($0.1 > \varepsilon Nd > -2.5$). Lead (Pb) isotope ratios plot above the NHRL (Northern Hemisphere Reference Line, Callegaro et al. 2014). The Sr-radiogenic isotope ratios and the high $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ values for given $^{206}Pb/^{204}Pb$ ratios, as well as the observed anomalies in Nb, P and Pb, suggest that contamination by continental crust played a significant role in the geochemical composition of these rocks. However, such an interpretation is still at odds with the low initial $^{187}Os/^{186}Os$ values (0.1298 ± 0.0056) recorded in these same rocks - a character interpreted as reflecting magma genesis by melting of a lithospheric domain affected by previous (Variscan?) subduction-related enrichment event(s). Yet, small amounts of crustal silicate contamination cannot be discarded. Some of the sampled magmatic rocks are characterized by high CaO contents (up to 20.4 wt%), initial $^{87}Sr/^{86}Sr$ ratios (up to 0.70746) and $\delta^{18}O$ (up to 8.7‰), interpreted as the result of limestone assimilation in the magma (Martins et al. 2008).

Volcanic rocks occupying the same stratigraphic position also occur in the Alentejo Basin (Santiago do Cacém region), but outcropping in a significantly smaller area. Their age (see also Verati et al. 2007), petrographic (Martins 1991) and geochemical characters (Martins 1991; Callegaro et al. 2014) indicate that they were also part of the same magmatic province as the Algarve. To this same province also belong, due to its location, age and geochemistry, the doleritic rocks of the 530 km-long, 5 to 200 m-thick Messejana–Plasencia dyke, trending NE–SW across the Iberian Peninsula from Central Spain to southwest Portugal (Alibert 1984; Sebai et al. 1991; Martins 1991; Cebriá et al. 2003; Callegaro et al. 2014).

6.4.2 Tithonian-Berriasian Magmatism

In contrast to the CAMP magmatism, which lasted for ≈2 My, the magmatic pulse spanning the Jurassic – Cretaceous transition lasted for about 8 My. Outcrop evidence for such a pulse is only recorded in the Lusitanian basin within a somewhat restricted area close to the Nazaré Fault Zone, between the latitudes of Rio Maior and Soure. For these rocks, an age range between 141 ± 0.95 Ma and 146.5 ± 1.5 Ma was obtained using U-Pb and $^{40}Ar/^{39}Ar$ methods (Grange et al. 2008 and Mata et al. 2015, respectively).

Tithonian-Berriasian magmatism is documented in the form of intrusive rocks of doleritic (*s.l.*) composition occurring mostly as sills and dikes. These intrusive rocks were chiefly emplaced along sub-meridian alignments that corresponded to rift faults resulting from the reactivation of Permian strike-slip faults, which also controlled the location of spatially related salt walls. The least evolved rocks (Mg# < 67.5) are mildly alkaline, but fractionation at pressures above 8 kbar led the magma to straddle the compositional divider between alkaline and sub-alkaline rocks towards SiO_2-saturated (and oversaturated) compositions. It should be noticed that rocks occurring north of the Nazaré Fault are clearly more evolved than those cropping out to the south. These northern intrusives also seem to derive from magmas generated by lower degrees of partial melting than in the south (Mata et al. 2015). Accordingly, magmas were generated in the presence of residual garnet and sampled a fairly homogenous source characterized by important Sr and Nd isotopic compositions (εNd_i from +1.6 to +4.2), more enriched than the typical N-MORB source. The Tithonian-Berriasian rocks are also characterized by somewhat high $^{207}Pb/^{204}Pb$ for a given $^{206}Pb/^{204}Pb$ (it is very high for one of the samples) as compared with Atlantic MORB (Grange et al. 2008). These isotopic data, and the fact that magmas were generated at the top of garnet zone (≈80 km depth)

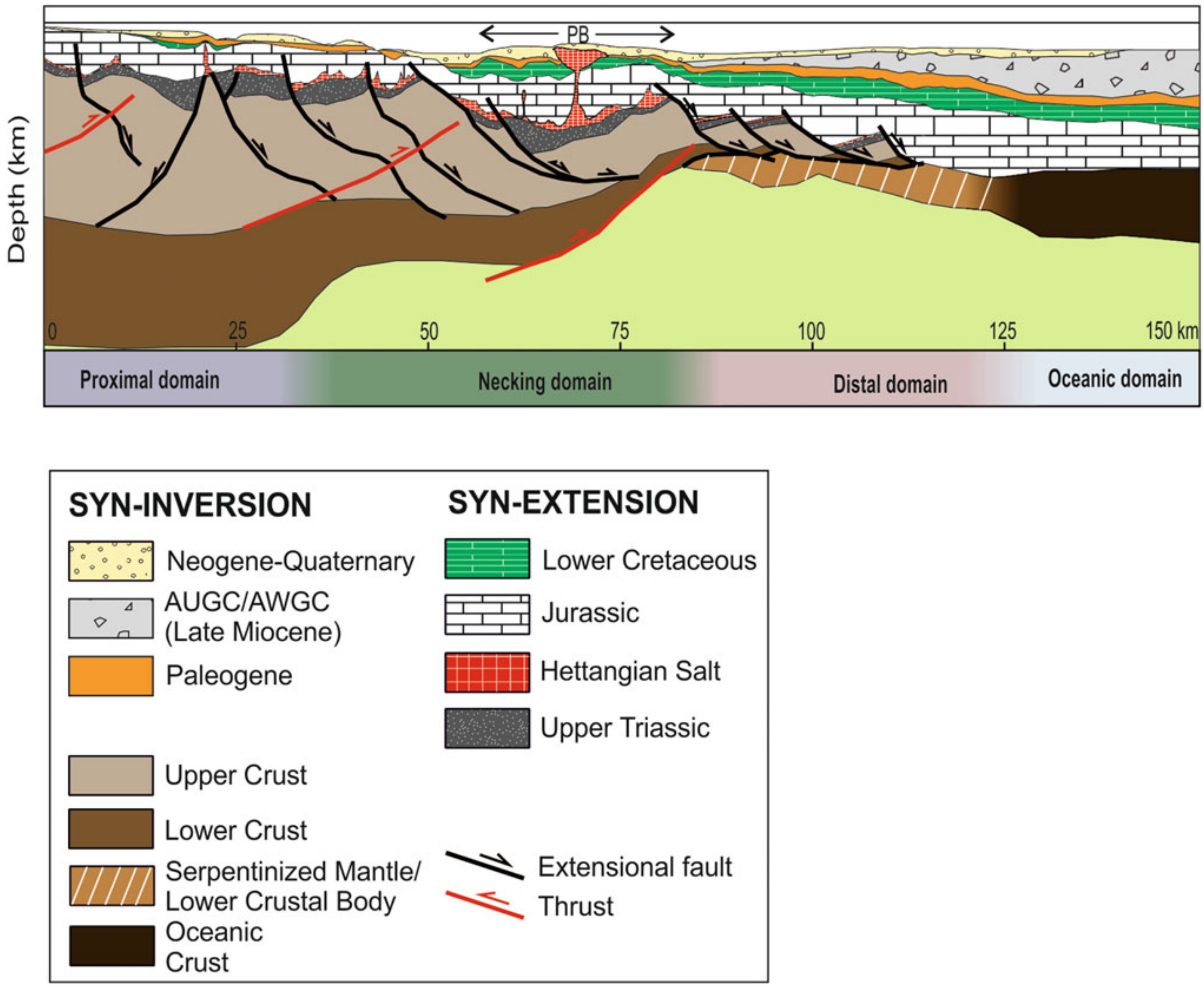

Fig. 6.4 Model for crustal structure and uppermost lithospheric mantle across the SWIM from gravity modeling (adapted from Ramos et al. 2017b) based on deep seismic reflection, gravimetric and magnetic modeling. PB- Portimão Bank

when the lithosphere thickness in the region was ~120 km (e.g. Tesauro et al. 2009), led Grange et al. (2008) and Mata et al. (2015) to propose a sub-continental lithosphere mantle source for their origin.

The composition of these onshore rocks is somewhat distinct (more silica-undersaturated and Nd-isotope enriched signatures, εNdi down to +1.6) from the *quasi* contemporaneous magmas emplaced offshore (εNdi down to +2.2). This was interpreted as a result of the less important onshore lithospheric stretching, leading to lower degrees of partial melting, which favoured a higher contribution of lower solidus metasomatised mantle domains to the onshore magmas (Mata et al. 2015).

As shown by Mata et al. (2015), some rocks preserve evidence for exogenous processes that involved Jurassic evaporites, leading to an increase in the Na_2O content (up to 5.56 wt %), normative nepheline and $^{87}Sr/^{86}Sr$ ratios (up to 0.707469). Documented differences, on each side of the Nazaré Fault, concerning magmatism, sedimentation rates (Kullberg et al. 2013), and lithospheric structure (Tesauro et al. 2009), are proof of the major role that this structure had during the Mesozoic, marking the transition between two domains of distinct crustal architectures; the NW Iberia lower-plate and the SW Iberia upper-plate (Pereira et al. 2017).

6.4.3 The Influence of Rifting on Magmatism

Lower Jurassic magmatic rocks occurring in the Algarve and Alentejo basins, as well as forming the Messejana–Plasencia dyke have been included in the Central Atlantic Magmatic Province (CAMP-Marzoli et al. 1999) a Large Igneous

Province that, at present, crops out in SE Europe, NE Africa, South America and North America, and that initially covered an area of about 11 × 10^6 km^2 (e.g. McHone 2000).

Despite the large area covered by volcanism, CAMP magmatism is thought to be emplaced in a very short period of time (≈2Mys; e.g. Verati et al. 2007), which could suggest the operation of a super-plume. However, plume-related magmatism is usually associated with previous domal uplift with offlapping and shoaling of regional sedimentation (e.g. Ernst and Buchan 2003), which clearly is not the case for the CAMP. As emphasized for the Algarve basin (Martins et al. 2008; Callegaro et al. 2014; see also Youbi et al. 2003), rift related magmatism preceded volcanism by some 20 to 30 Ma, strongly suggesting magma generation by adiabatic decompression as a response to a significant stretching event. This same kind of mechanism was proposed by Mata et al. (2015) for the Tithonian-Berriasian magmatism given it immediately follows an important rifting event in the Lusitanian basin.

The magma sources for these two magmatic cycles were, nevertheless, distinct. As shown by Mata et al. (2015) rift-related Mesozoic magmatism in Iberia evolved from tholeiitic (Lower Jurassic) to mildly alkaline (Tithonian-Berriasian). With time, magmas became enriched in incompatible elements, while their sources turned more time-integrated depleted, as can be inferred from the Sr and Nd isotope ratios of unaltered and uncontaminated rocks. The change in magmatic affinities from tholeiitic to mildly alkaline of magmas generated from sub-continental lithosphere was explained by Mata et al. (2015) by important extension rates during the Lower Mesozoic (e.g. Cunha 2008), which induced higher partial melting degrees and shallower depth of magma segregation, both favouring more SiO_2-enriched magmas.

6.5 Rifting Events in the SWIM: The Algarve Basin

Terrinha P, Ribeiro C, Ramos A, Muñoz JA, Fernández O

6.5.1 Rifting Events

The Algarve Basin (AB) is exposed onshore along a stretch of ca. 150 km, parallel to the West-East coastline, and shows a maximum width of 25 km (Fig. 6.5). The onshore Mesozoic AB rift basin exposes a stratigraphic record that spans, at least, the Late Triassic to Cenomanian (lowermost Late Cretaceous), revealing multiple hiatuses and unconformities (Fig. 6.5). Crustal extension in the AB was accommodated by SW to W-striking extensional faults dipping oceanwards; the main basin depocentre was developed towards the southeast. Onshore, the AB is segmented by transfer faults striking N-S and NW-SE, across which sediment thickness varies considerably. Towards the offshore depocentre, the role of transfer faults disappears gradually as their displacement was likely distributed into wider transfer fault zones and thick salt layers. Offshore, the AB is bounded by the Guadalquivir Bank basement horst, located approximately 100 km south of the Paleozoic limit of the basin (Figs. 6.1 and 6.5).

The stratigraphic panel in Fig. 6.5, which schematically reveals multiple unconformity bounded megasequences, suggests the existence of five extensional episodes. Extensional faults showing syn-sedimentary geometric criteria are well exposed along the western sector of the basin (Terrinha 1998).

The first extensional event spans the Triassic-Hettangian, during which the red continental siliciclastic sediments were deposited. The sequences grade upwards from coarse conglomerates to finer clastics, clays and evaporites as one moves away from the basin border. Ramos et al. (2017c) show a zonation, in a N to S direction, from clay diapirs, through anhydrite to halite diapirs.

Continental terrigenous formations of Triassic age were described in detail by Palain (1976), who confirmed a consistent record of paleo-currents from NE to SW. The thickest onshore depocentre along the strike of the AB is located on the hanging-wall of the São Marcos-Quarteira Fault, a Variscan thrust reactivated as an extensional fault during rifting. These observations suggest that, during this extensional event, main depocentres were located to the SW of the AB as a result of widespread extensional reactivation of NW-SE trending Variscan thrusts. The Triassic-Hettangian rifting pulse ended with the Central Atlantic Magmatic Province (CAMP) event (Hettangian-Sinemurian), which produced extensive sub-aerial lava flows and pyroclastic events that outcrop continuously along the AB.

Data from U-Pb analyses of detrital zircon from Upper Triassic sandstones point to the following sources: i) Upper Devonian-Lower Carboniferous sediments of the South Portuguese Zone (Phyllite-Quartzite and Tercenas Formations), and the Meguma Terrane (Pereira et al. 2017). In parallel, the nearby Alentejo Basin, located on the WIM, yielded different sources. This suggests that during the first rifting stages, the WIM and SWIM were already separated by a topographic high (rift shoulder) bounding distinct drainage networks (Fig. 6.6).

The second rifting event in the AB started in Sinemurian times when marine sedimentation was first recorded, and lasted until Toarcian times. In the AB, Toarcian deep-marine marls with ammonoids are erosionally truncated and overlain by coral reefs and calciclastic limestones of Mid Jurassic age, probably belonging to the Aalenian.

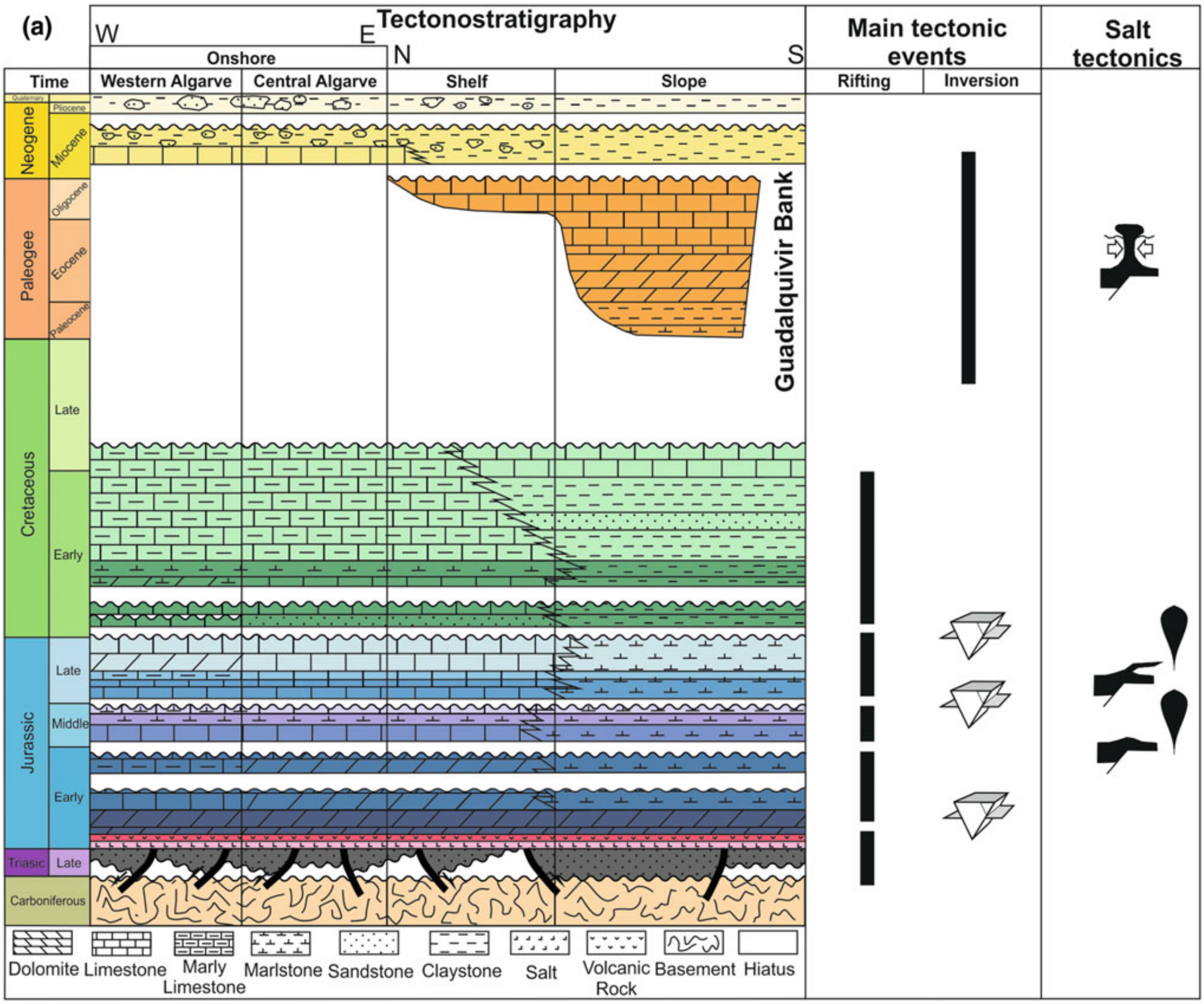

Fig. 6.5 Stratigraphy and extensional tectonics of the Algarve Basin. **a** Schematic stratigraphic table showing main basin tectonic and salt tectonics events (salt pillows, diapirs and allochthonous nappes). Note the existence of various hiatuses/unconformities in the Mesozoic and that some coincide with transient compressive events (adapted from Ramos et al. 2016). **b** Extensional tectonic framework of main basement faults and salt structures. Onshore: schematic location of main W-E to SW-NE faults segmented by N-W to NW-SE faults. The first set is associated to the main Tethyan rifting, the second to the Atlantic rifting. Offshore: main basement faults and salt diapirs, pillows and allochthonous salt layers. Red dashed line: boundary between anhydrite-claystone salt domain and halite salt domain. GoC- Gulf of Cádiz; SMQF- São Marcos-Quarteira Fault (adapted from Ramos et al. 2017c)

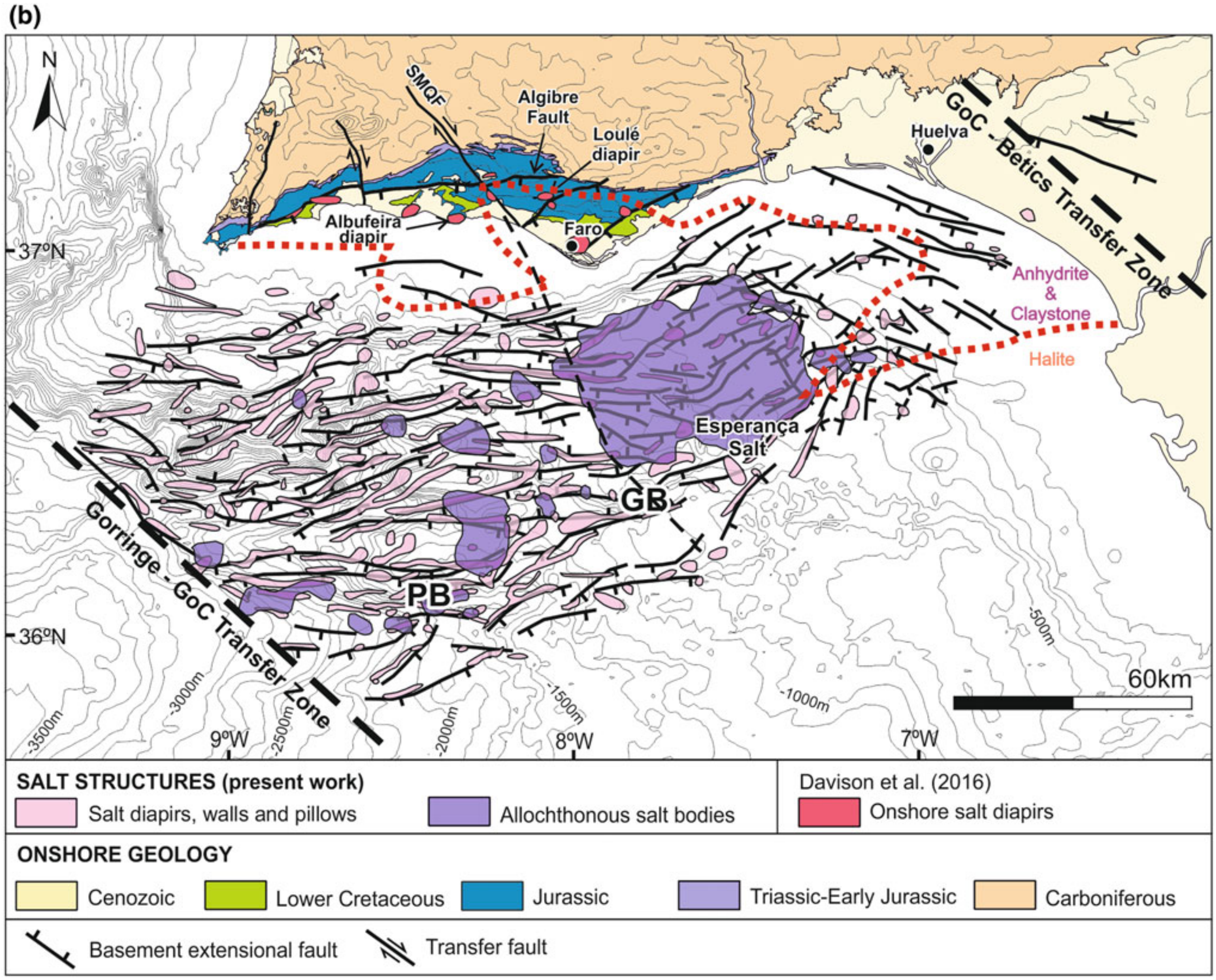

Fig. 6.5 (continued)

This extensional event was not continuous; it was interrupted by a transient compressive and uplift during which hydrothermal SiO_2 (quartz and crypto-cristalline varieties) precipitated in dykes, faults and tension gashes together with a dolomitisation event (Terrinha et al. 2002; Ribeiro and Terrinha 2007). This inversion event lasted less than 2 My affecting external shelf calciclastic sands of early Pliensbachian age that are unconformably overlain by limestones and marls with ammonoids of late Pliensbachian age. The second rifting event in the AB, and transient inversion episodes, occurred during the initial westward propagation of the Neo-Tethys between Africa and Iberia (Sallarès et al. 2013).

A third extensional phase occurred during the Middle Jurassic. The detailed relationships between the Mid Jurassic stages can be observed in the Mareta beach cross-section in Sagres, West Algarve (Terrinha 1998; Terrinha et al. 2002; Fig. 6.7). Coral reefs of probable Aalenian age are karstified and overlain by calciclastic syn-rift deposits of Bajocian age. The Bajocian age extensional faults can be grouped in two sets, NE-SW and N-S striking, i.e. Tethyan and Atlantic rift systems. Bathonian deep-marine marls (deposited during a tectonic quiescent time interval) onlap on top of an erosion surface that truncates the Bajocian syn-sedimentary extensional faults. Extensional tectonics resumed during the Callovian, at the end of which another transient compressive episode occurred – as marked by the occurrence of landslides and slumps at coastal outcrops (Rocha 1976; Terrinha 1998; Terrinha et al. 2002).

Sedimentation and extensional tectonics were resumed in Oxfordian times (internal shelf carbonates in western Algarve), followed by uplift and compression during the Jurassic-Cretaceous transition. Limestones and marls with ammonoids occur in the central Algarve.

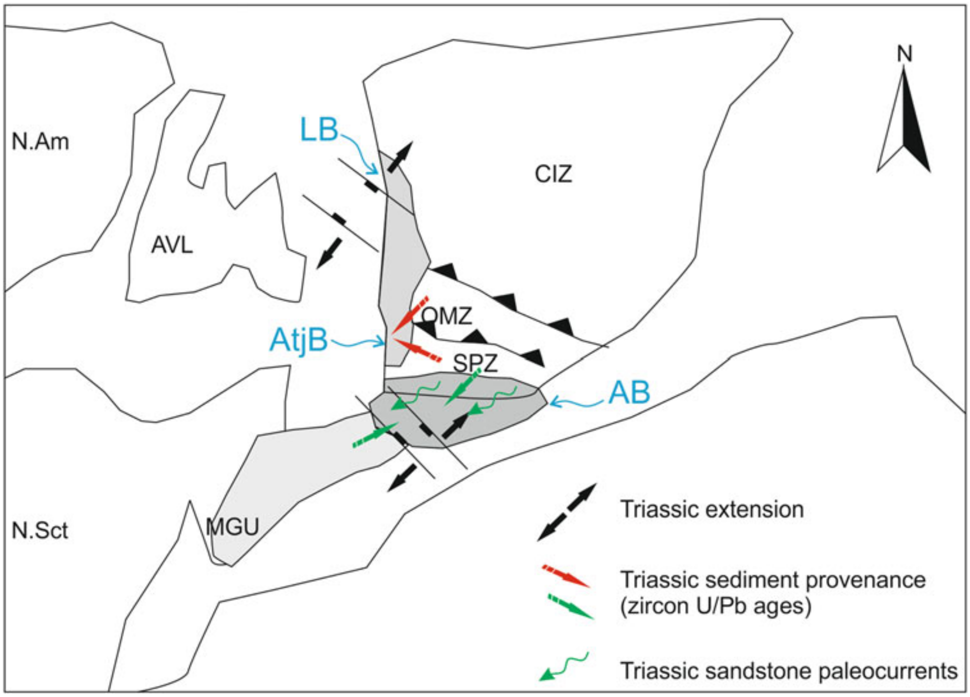

Fig. 6.6 Schematic representation of the Triassic paleogeography within Pangea (cf. with Fig. 6.1). AB, Algarve Basin; AtjB, Alentejo Basin; AVL, Avalonia terrane; CIZ, Central Iberian Zone; LB, Lusitanian Basin; MGU, Meguma terrane; N.Am, North America; N. Sct. Nova Scotia; OMZ, Ossa-Morena Zone; SPZ, South Portuguese Zone. Existence of a depocentre in SW AB results from orientation of drainage pattern deduced from paleocurrent criteria (Palain 1976) and the sources of detrital zircons in AB Triassic sediments (Pereira et al. 2017). NW-SE Triassic-Hettangian salt graben SW of AB, from Ramos et al. (2017c) (cf. with Fig. 6.5b). These data suggest that the LB and AB were separated by a topographic high in Triassic times. The NW-SE striking Monte Real graben in the LB is compatible with a NE-SW oriented tectonic extension in Triassic times, probably extending the NW-SE striking Variscan orogenic structures

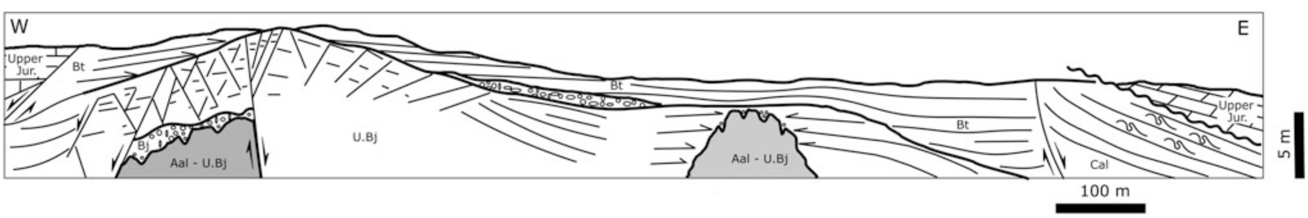

Fig. 6.7 Schematic interpretation of the field section of Praia de Mareta in Sagres, West Algarve. The whole Mid Jurassic section exposes Aalenian (Aal) through Callovian (Cal) strata. Oxfordian sediments are dolomitized across the unconformity in the East and downthrown across normal fault in the west. Note that the Bajocian (Baj) extensional faults in calciclastic limestones with *Zoophycus* are sealed by Bathonian (Bt) marls with ammonoids. The Bathonian does not display syn-sedimentary normal faults; it could be argued that thermal subsidence followed the Bajocian tectonic subsidence. The Aalenian coral reefs were deeply karstified before the deposition of Bajocian strata, indicating a relative sea level drop before Bajocian extension Adapted from Terrinha, 1998

The fifth phase of extension and renewed sedimentation occurred in the Early Cretaceous. Roll-overs and stratigraphic overthickening of hanging-wall sequences can be observed in the Lower Cretaceous of West and Central AB (Terrinha 1998). The Cenomanian occurs in eastern Algarve and is interpreted to be post-rift and contemporaneous with a worldwide sea-level transgression.

Although wedge-shaped sedimentary packages can be described in offshore seismic reflection profiles, allowing the definition of key syn-tectonic formations, the stratigraphic calibration of the Jurassic and its sub-division from periods into ages (e.g. Lower Jurassic into Pliensbachian) is not possible on seismics alone. These syn-rift formations were defined in the field correlating detailed structural observation (geometric and kinematic criteria for syn-tectonic and post-tectonic sedimentation) with ammonoid biozones (e.g. Rocha 1976; Terrinha 1998; Terrinha et al. 2002). The westernmost sector of the Algarve Basin, the Sagres sector, was a structural high during the Jurassic and Cretaceous allowing for the deposition of condensed series that are ideal to observe the correlation of detailed stratigraphy and tectonics.

The Gorringe Bank-São Vicente sector comprises the transition between the SWIM and WIM, the first associated with the westward propagation of the Tethys Ocean, and the second with the northwards propagation of the Atlantic Ocean. The Gorringe Bank exposes a piece of continental mantle exhumed during the Mesozoic rifting that was later overthrust to the NW during the Alpine orogeny.

In more detail, N-S and SW-NE trending extensional rift faults were mapped offshore (see Fig. 6.2). Onshore, in the Cape S. Vicente region, SW-NE to W-E extensional faults of Tethyan affinity are segmented by the Atlantic ~ N-S striking extensional faults. To the north of the AB, in the Carboniferous of the South Portuguese Zone these faults show strike-slip geometries previous to the Triassic extension, showing the Late-Variscan origin of at least a part of the Mesozoic extensional faults.

6.5.2 Salt Tectonics

Salt tectonics is a general term used to describe geological deformation associated with evaporites. However, some similar structures may result if very week rocks, such as overpressured shales are involved during the deformation. Onshore the AB, salt walls can be found associated with the main extensional faults. The Espiche (clay), Albufeira (gypsum, anhydrite) and Loulé (halite) salt walls are the best exposed in the area (Fig. 6.5). The Loulé salt wall has been studied by several authors because salt is exploited and accessible in an underground mine (Terrinha 1998; Terrinha et al. 1990; Machek et al. 2014; Davison et al. 2016). Various types of structures contemporaneous with salt migration can be observed at Loulé, such as recumbent folds contemporaneous with layer parallel salt movement, sub-vertical sheath folds and shear zones (Terrinha 1998; Terrinha et al. 1993; Davison et al. 2016). These salt walls formed due to salt migration along main rift faults and were shortened during Cenozoic tectonic inversion.

Various types of salt structures were identified in seismic reflection profiles, such as salt pillows, diapirs and salt sheets (Terrinha 1998; Lopes C 2002; Lopes F 2002; Lopes et al. 2006; Matias et al. 2011; Ramos et al. 2017c). Some diapirs are being deformed and moving at present as also imaged in swath bathymetry (Terrinha et al. 2009; Ramos et al. 2017c). The existence of an allochthonous salt nappe in the AB was proposed by Terrinha (1998) and demonstrated by Matias et al. (2011). Ramos et al. (2017c) mapped in detail the offshore SWIM salt structures and correlated the interpreted seismic data with industry wells and field observations, reaching important conclusions regarding evaporite facies distribution, paleogeography and tectonics.

The Triassic-Hettangian salt basin displays a lithological zonation; the central part of the Triassic-Hettangian Algarve Basin consists of halite-dominated deposits surrounded by clay-anhydrite deposits to the north, northeast and south. Such a distribution indicates that, in Triassic-Hettangian times, (i) the Guadalquivir Bank (southern boundary of the AB at present and during the Jurassic-Cretaceous rifting) was already a basement high, (ii) the basin depocentre was located in the SW of the AB with maximum salt thickness of 600 m, and (iii) the AB developed between NW-SE trending transfer faults, the Gorringe-GoC and the Betics-GoC transfer faults (Ramos et al. 2017c). One of these transfer-extensional faults, the São Marcos-Quarteira Fault (SMQF, in Terrinha 1998; Terrinha et al. 2002) resulted from the reactivation of a Variscan thrust, which suggests that initial rifting in the SWIM resulted from negative inversion of the Paleozoic thrusts. This interpretation is compatible with the general drainage pattern towards the SW shown by Palain (1976) and could be related with the origin of detrital zircons from the Meguma terrane as proposed by Pereira et al. (2017) (Fig. 6.6).

One of the most striking features of the Algarve basin is the existence of an allochthonous salt sheet that has been named as the Esperança Salt Nappe (Matias et al. 2011; Ramos et al. 2017c; Terrinha 1998). It is a 40 × 60 km^2 salt sheet sourced from salt walls that developed along basement-involved, landward-dipping extensional faults to the north and east of the nappe. Salt evacuation from the linear L-shaped feeder was triggered by sedimentary loading and extension during Mid to Late Jurassic times. The allochthonous salt sheet climbed up into the Jurassic subsalt stratigraphy to the south and west (Fig. 6.8). Welding of the feeder system at Late Jurassic times coincided with the end of the extension. Subsequent salt evacuation of the allochthonous salt sheet by sedimentary loading during Late Jurassic-Early Cretaceous times triggered the development of a roho extensional system and related minibasins linked basinwards with a contractional system at the toe of the Esperança Salt Nappe at the southern and western edges of the nappe (Fig. 6.8). The structure of the Esperança Salt Nappe and suprasalt sediments has been modified by the post-Mesozoic shortening that has resulted into the reactivation of salt structures (Ramos et al. 2017c).

Hence, the initiation of salt diapirism in the AB was probably diachronous depending on basement tectonics and sedimentary loading. The earliest salt pillows and salt rollers are probably of Early Jurassic (post-Sinemurian) age (Ramos et al. 2017c).

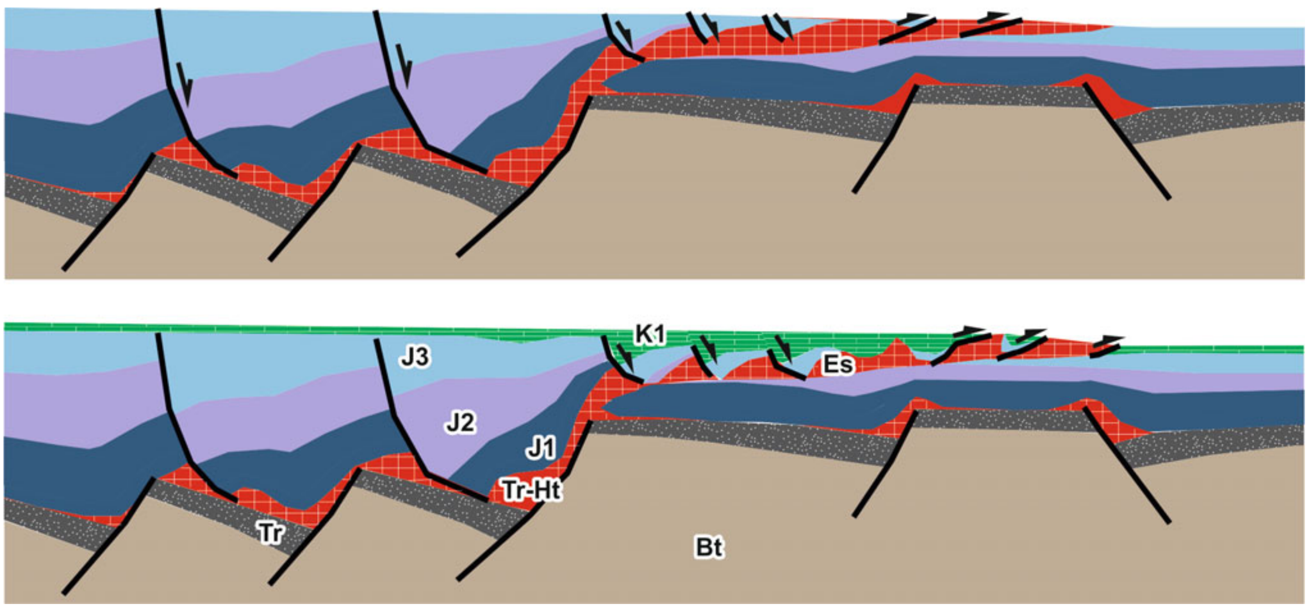

Fig. 6.8 Schematic representation of the mechanism of formation of the Esperança allochthonous salt nappe in Mid Jurassic and Late Jurassic times, with subsequent horizontal propagation and deformation in the Cretaceous Adapted from Ramos et al. (2017c)

6.5.3 Diagenesis, Hydrothermalism and Dolomitisation

Ribeiro C and **Terrinha P**

Among the lithologies of the Mesozoic sedimentary record of the AB, an extensive volume is made up of carbonate rocks, mainly limestones, marly limestones and dolomites, deposited in marine environments of varied depths (Rocha 1976; Terrinha et al. 2013).

In natural systems, the precipitation of dolomite in marine environments can occur at different moments: (i) during the early diagenesis of limestone formations (Perkins et al. 1994; Wu and Wu 1998; Yoo and Lee 1998; Mutti and Simó 1994; Jingquan 1998; Lu and Meyers 1998), (ii) during late diagenetic evolution of limestones (Miller and Folk 1994; Kupecz and Land 1994; Simó et al. 1994; Flood et al. 1996; Moss and Tucker 1996; Reinhold 1998), or (iii) related to hydrothermal events without any causal relation with diagenesis (Coniglio et al. 1994; Tritlla et al. 2001).

The sedimentary record of the AB includes different types of dolomitic formations, most of them of considerable thickness and lateral extent, associated with marine carbonate sedimentary packages: (i) the Upper Triassic-Lower Hettangian “Complexo margo-carbonatado de Silves” (translated as Silves marl-carbonate complex) has a discontinuous dolomitic member near its top, deposited in a margin-coastal lagoon environment (Azerêdo et al. 2003), containing gypsum and anhydrite as evaporite minerals; (ii) the Sinemurian dolomitic limestones and dolomites are a thick package of carbonate rocks without evidence for primary sedimentary structures and a poorly preserved marine fossil content (Rocha 1976), most probably due to the diagenetic character of the dolomitization; (iii) the Lower Jurassic (Pliensbachian) sediments are made up of limestones, dolomitic limestones and dolomites, with clear evidence for a geochemical replacement process during the early stages of diagenesis (Ribeiro and Terrinha 2007); (iv) the transgressive Mid Jurassic formations (Azerêdo et al. 2003) are divided into sediments deposited in a pelagic open marine environment (mainly marls and limestones) and sediments, including dolomitic limestones, deposited in back reef lagoon environment; v) following the regression at the end of the Middle Jurassic pelagic sedimentation in the AB resumed with the deposition of limestones, some of which are dolomitic.

Some of dolomitic lithologies mentioned above show evidence for a substitution process. The lower Pliensbachian limestones and dolomitic limestones exhibit substitution textures, both at the mesoscopic scale—where layers of limestone with fragments of lamellibranches and echinoderms gradually change to a dolomite with remnants of the fossilized exoskeletons—and at the microscopic scale—where textures of calcite dissolution and later precipitation of dolomite are evident and support a secondary origin for the dolomitization (Ribeiro 2006). The dolomitized strata are truncated at their top and unconformably overlain by upper Pliensbachian sediments, showing that dolomitization occurred during early diagenesis.

The 87Sr/86Sr isotopic ratios of the Pliensbachian dolomitic rocks are significantly different from the isotopic ratio of sea

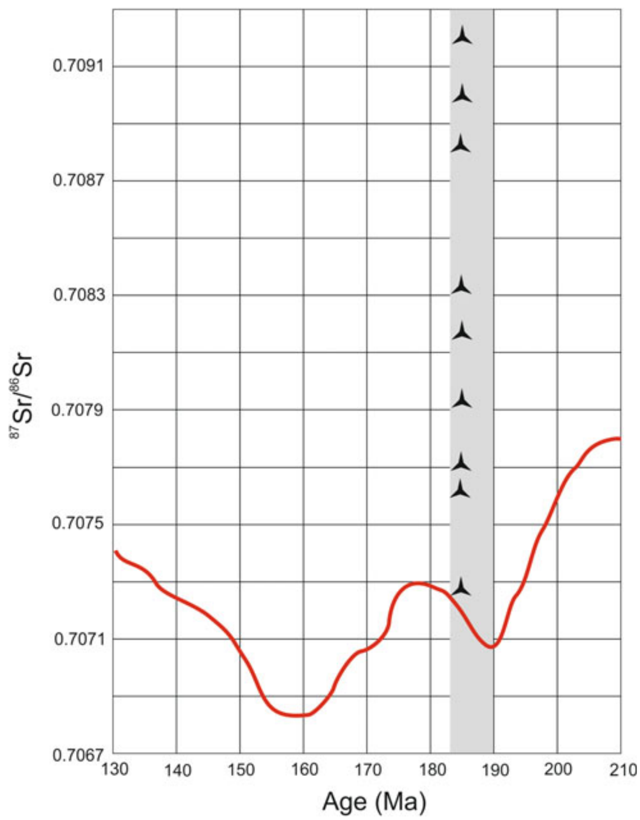

Fig. 6.9 87Sr/86Sr ratios for dolomites and dolomitic limestones (stars) from the lower Pliensbachian of the SW Algarve Basin. The red curve represents the strontium isotope ratio of the sea water between 210 Ma and 130 Ma. The gray bar marks the biostratigraphic limits (based on Ammonoid fauna) of the formation. The enrichment in 87Sr of the dolomitic rocks relatively to sea water is the result of the interaction of the carbonates with exotic water influx during the early diagenesis. Adapted from Ribeiro, 2006

water (Fig. 6.9), supporting a secondary origin of the dolomitization by reaction with exotic fluids likely sourced from emerged areas during a relative sea-level drop (Ribeiro et al. 2010).

Apart from dolomitization, post-depositional events of silicification are observed in the Lower Jurassic succession in the AB. The most pervasive of these events is the silicification of calciclastic sediments that originated cherts (Ribeiro and Terrinha 2007), observable both in the western (Sagres) and eastern (Tavira) parts of the AB. In the western part of the basin quartz dikes were injected along normal faults and quartz veinlets in extensional joints cutting the carbonate lithologies (Fig. 6.10). Once again, the confinement to certain stratigraphic levels supports an early diagenetic origin of the events of chertification and quartz dikes and veinlets formation. However, microthermometric data on fluid inclusions of the quartz dikes are in the range of hydrothermal systems (Ribeiro 2006).

Both events (chert formation and quartz dikes and veinlets formation) imply the percolation of silica-rich fluids in the lower Pliensbachian limestones before the sedimentation of the upper Pliensbachian, where silica phases are absent. Chert and quartz infillings probably represent phases of diffuse flow and focused flow of the hydrothermal system, respectively. The latter are hydrothermal processes associated with the early Pliensbachian extensional tectonic event, while the dolomitization was probably associated with a Late Pliensbachian sea-level drop.

6.6 WIM Rifting Events: The Lusitanian Basin

Kullberg JC and **Alves T**

The Lusitanian Basin (LB) evolved on the West Iberia Margin (WIM) from the Triassic to the uppermost Early Cretaceous. It is located in the central part of the WIM and is divided into an internal, proximal, well exposed onshore basin and a distal deep offshore basin, named Peniche Basin by several authors. These basins are separated by the N-S striking Berlengas Horst.

The onshore part of the LB is exposed along a stretch that is about 200 km parallel to the coast, or a maximum of 100 km of width (Fig. 6.11). The LB exposes a rather complete stratigraphic record from, at least, Upper Triassic

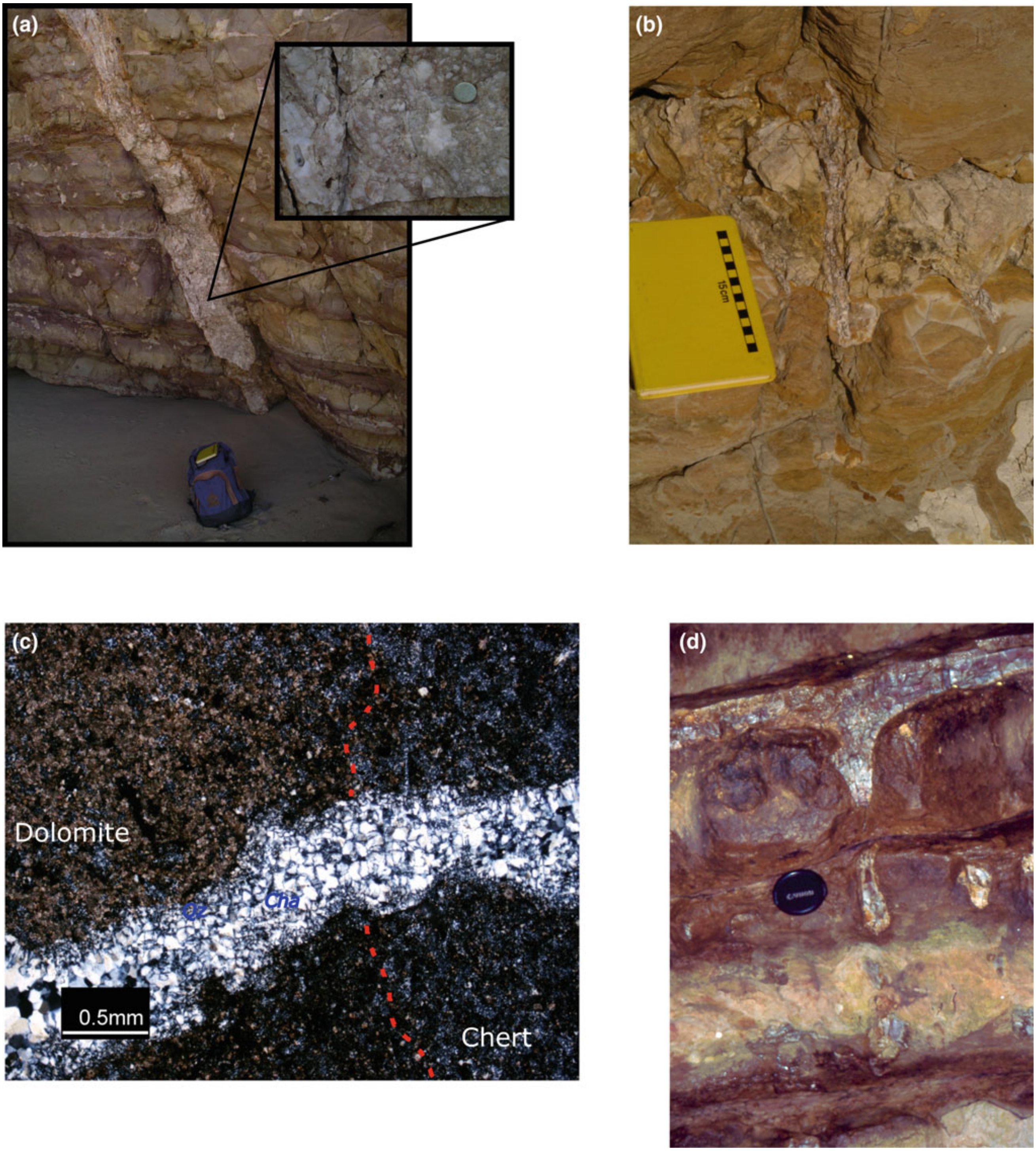

Fig. 6.10 Early diagenesis of the lower Pliensbachian sediments. **a** Quartz dike in normal fault; inset shows the cataclastic texture of the infilling supporting the existence of several episodes of fault movement and flow of SiO_2-rich fluids. **b** Quartz and chalcedony veinlet in an extensional joint cutting dolomitic limestones and chert. **c** Thin section of a contact between a chert nodule and dolomite (dashed red line), cut by a chalcedony + quartz tension gash with evidence of polyphasic opening (large chalcedony crystals – Cha - at the center of the veinlet bordered by small quartz crystals at the contacts of the veinlet with the host sediments). **d** Chert layers and nodules formed by the substitution of calcareous sands during early diagenesis; calcareous sands were injected into tensile joints both downwards and upwards forming chert dikes up to 3 m in length and tens of meters of height (Ribeiro and Terrinha 2007). The observed silicification is associated to the percolation of hydrothermal fluids during the early Pliensbachian extensional tectonic event of the AB, while the dolomitization is related to the interaction of the limestone with meteoric fluids during a sea level drop Adapted from Ribeiro, 2006

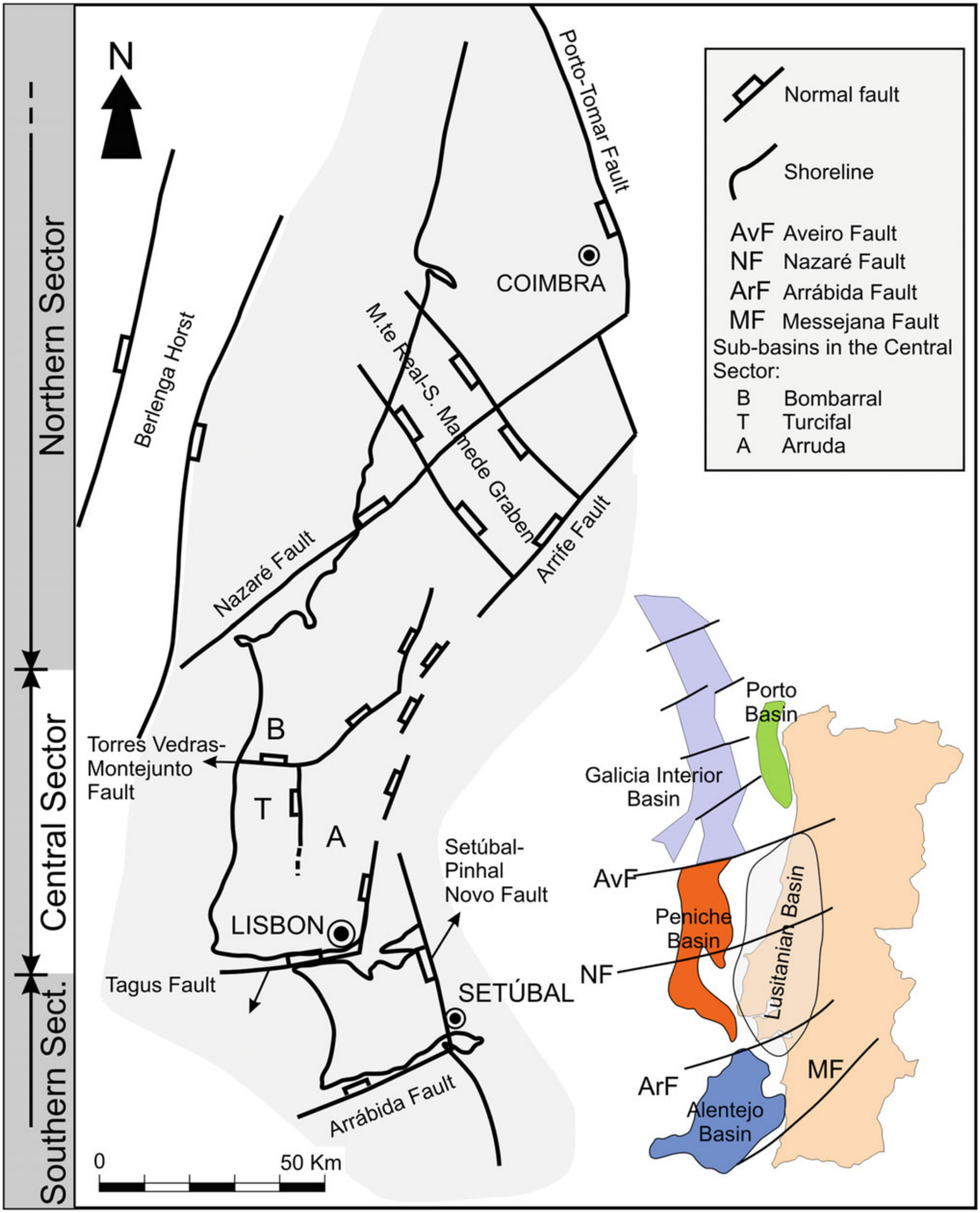

Fig. 6.11 Structural map of the extensional rift faults of the Lusitanian Basin and its sub-division into three sectors. Inset shows the relative position of the Lusitanian Basin with respect to other rift basins of the West Iberia Margin (WIM) (Adapted from Kullberg et al. 2013)

(Soares et al. 2012a) to the uppermost Lower Cretaceous with several unconformities of tectonic origin that mark different rifting episodes on the WIM. Upper Cretaceous post-rift sedimentary units, associated to eustatic processes, unconformably overlay the syn-rift successions. Exposed areas of the onshore LB were exhumed due to tectonic inversion of the WIM in the Cenozoic, mainly during the Neogene, associated with a deep-seated foreland detachment rooted in the collisional front of the Iberia-Africa orogen (Ribeiro et al. 1990; Kullberg et al. 2013).

Extensional tectonics in the LB was relatively constant and followed an approximate E-W direction, accommodated by N-S to NNE-SSW striking extensional faults dipping to opposite sides that controlled an overall graben geometry (Fig. 6.11). The present day contour map of the top basement of the onshore LB shows an approximately symmetrical basin with a N-S striking depocentre roughly located along its central part (Ribeiro et al. 1996; Rasmussen et al. 1998; Kullberg 2000) (Fig. 6.12). In the Peniche basin, rotational extensional faults dipping to the offshore are dominant, giving rise to a general deepening towards the external deep margin (Alves et al. 2006). Immediately to the west of the Berlengas High, however, a simpler graben geometry is observed in continental slope basins, contrasting with the more asymmetric (half-graben) geometry of basins in the most distal parts of the WIM.

These different geometries controlled by the overall geometry of the top of the basement, are responsible for two different tectonic styles. In the LB thick-skinned extensional tectonics prevail, whilst the rotating blocks of basement in continental slope basins to the west of the LB records a mixed thin- and thick-skinned style.

The main confining structures of the onshore LB are: i) the Berlengas Horst to the West, that separates the internal and external domains of the WIM, and ii) the Porto-Tomar Fault Zone in the Eastern border of the LB (Kullberg et al. 2014).

The Berlengas Horst consists of Paleozoic and Proterozoic basement uplifted during the Mesozoic rifting and subsequent continental break-up phases. The sedimentary record shows that this structural high was tectonically active during the 135 My that lasted the evolution of the basin. The Porto-Tomar Fault is an inherited Paleozoic suture of the Variscan orogeny. Apparently, this tectonic boundary of the LB was mainly active from Late Jurassic times.

The LB is segmented by WSW-ENE striking transfer faults that played an important role during both the Mesozoic rifting and Cenozoic Alpine orogeny. During rifting, these transfer faults subdivided the LB into various sectors with important thickness and facies variations. At a larger scale they segmented the WIM into different rift basins, the Porto and Inner Galicia offshore basins in the north, the LB in the centre and the Alentejo Basin in the south. The LB and the Alentejo Basin are separated by the Arrábida fold and thrust belt of Miocene age, formed due to one of these transfer faults that acted as a buttress fault.

Within the onshore LB, the Arrife-Lower Tagus and the Nazaré transfer faults segment the basin into three sub-basins (Fig. 6.11). This evidence led Soares and Rocha (1985) to subdivide the Lusitanian Basin in three sectors: (i) the Northern sector, to the North of the Nazaré Fault; (ii) the Central sector, limited by the Nazaré and the Arrife-Lower Tagus Fault System; and (iii) the Southern sector to the South of this fault system.

Inspection of the simplified stratigraphic table in Fig. 6.13 suggests the existence of four rifting intervals marked by basin wide unconformities. Each one of these pulses began with an increase of the extensional tectonics marked by modifications in the stratigraphic record, by important changes in the overall basin geometry and/or rotation of the stretching direction. Leinfelder (1987), Wilson et al. (1989), Rasmussen et al. (1998) and Leinfelder and Wilson (1998) proposed distinct seismic-stratigraphic frameworks for the interpretation of syn-rift strata in the Lusitanian Basin. Based on these, Alves et al. (2003a) undertook a review of systems tracts in the Central and Northern part of the basin to tackle the differences between the stacking patterns of fault—and diapir-bounded basins (Fig. 6.14). In essence, north to north–northeast—and northwest-trending faults follow the strike of Variscan structures identified in west Iberia, with the Turcifal and Arruda sub-basins being divided by the nearly 20-km-long, north- to north–northeast-striking Runa Fault (Fig. 6.13). The Turcifal and Arruda half-grabens are separated from the Bombarral-Alcobaça sub-basin by a 70-km-long structure, the Torres Vedras–Montejunto lineament (Fig. 6.13). Distinct salt anticlines occur on the western and eastern margins of the Bombarral-Alcobaça sub-basin.

In the context of the Lusitanian Basin, Alves et al. (2003a) demonstrated that propagation of sub-salt faults during the rift climax phase (late Oxfordian) was signed on outcrop and well data by the deposition of coarse-grained turbidites and carbonate olistoliths, depositional facies that are associated with tectonically active periods (Alves 2015). This event marked the through-going phase of fault development (Gupta et al. 1998; Gawthorpe and Leeder 2000) with: (1) gently warped sub-basins in Arruda and Turcifal sectors evolving as separate half-grabens from the late Oxfordian onwards; (2) a carbonate-dominated setting of the rift initiation phase changing abruptly to a siliciclastic-dominated regime in which transverse-derived drainage systems developed (Fig. 6.15). Most of the latter transverse drainage systems consisted of canyon- and gully-related submarine fans gradually filling the accommodation space created during the rift climax (Fig. 6.15).

During the immediate post-rift phase, the progressive infill of the Central Lusitanian Basin was accentuated in

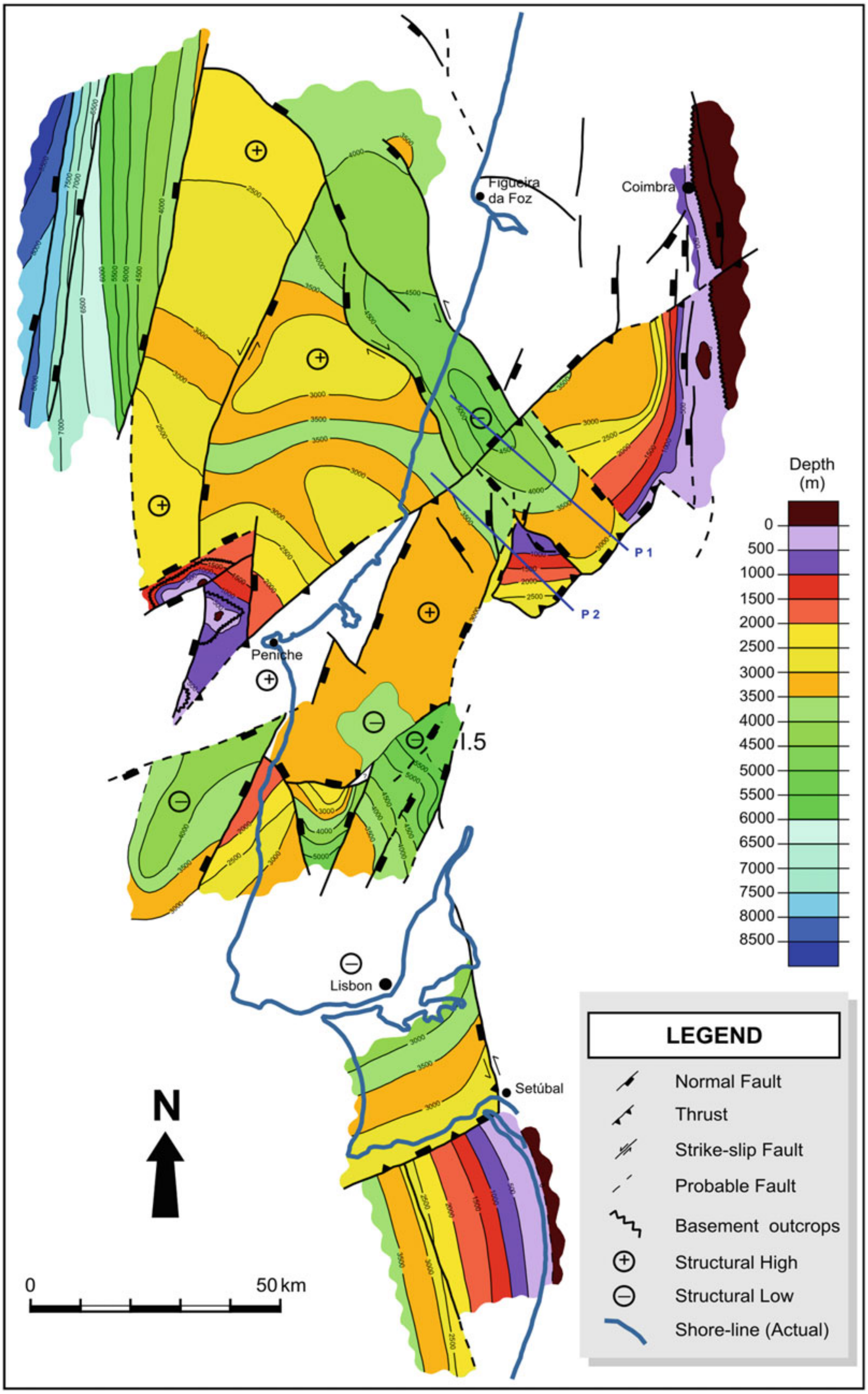

Fig. 6.12 Top of the basement map based on sub-surface information from onshore and offshore industry wells, depth converted maps (adapted from Ribeiro et al. 1996; Rasmussen et al. 1998; Kullberg 2000). The dominant extensional faults strike is NNE-SSW; the Nazaré Fault (cf. with Fig. 6.11) has been reworked at the basement level; ENE-WSW striking basement faults also recognized in this model. Note the NW-SE striking Triassic-Hettangian graben of Monte Real in the northern part of the LB

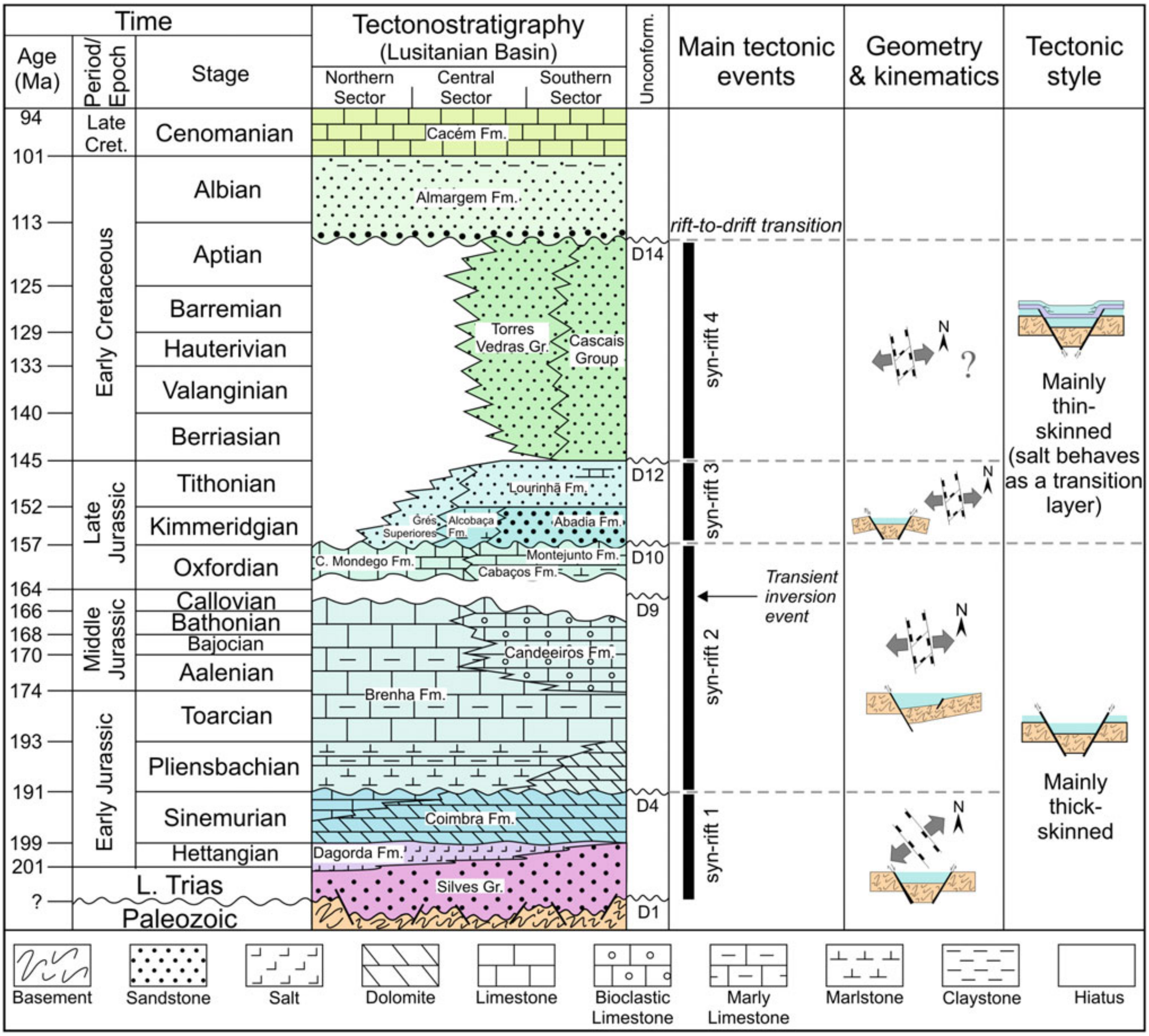

Fig. 6.13 Lithostratigraphic table and synthesis of main tectonic extensional events, geometry and tectonic style (adapted from Kullberg at al. 2013; Kullberg and Rocha 2014)

association with a relative decrease in fault-related subsidence (see also Stapel et al. 1996). This relative decrease in basin subsidence contributed to the progradation of axial and transverse drainage systems in the study area which, entering the basin from north, northeast and northwest, began to feed a southeast prograding continental slope (sequence A3) (Fig. 6.16). Such a rapid infill of the sub-basins by the prograding deposits of Sequence A3 suggests a relatively high sediment yield during the immediate post-rift phase in the Lusitanian Basin, now understood to be related to enhanced subsidence in offshore basins anticipating continental break-up on the WIM (Alves and Cunha 2018). This 'post-rift' sequence in the Lusitanian Basin was almost immediately followed by important shallowing in facies in a passive-margin setting (Alves et al. 2003b) (Fig. 6.16).

Similarly to the Algarve Basin (AB), the first rifting episode spans the Late (?) Triassic through the Sinemurian. This sequence, underlain by a first order unconformity that marks the beginning of a new Wilson Cycle post-dating the Variscan Orogeny starts with deposition of continental coarse-grained siliciclastics in the Triassic with progressive evolution to evaporitic clays and marls, typical of transitional environments during the Hettangian. The sequence ends with marine, confined primary dolomites of Sinemurian age. The continental to transitional units belong to the Silves Group, which is similar to the AB. It is relevant to point out

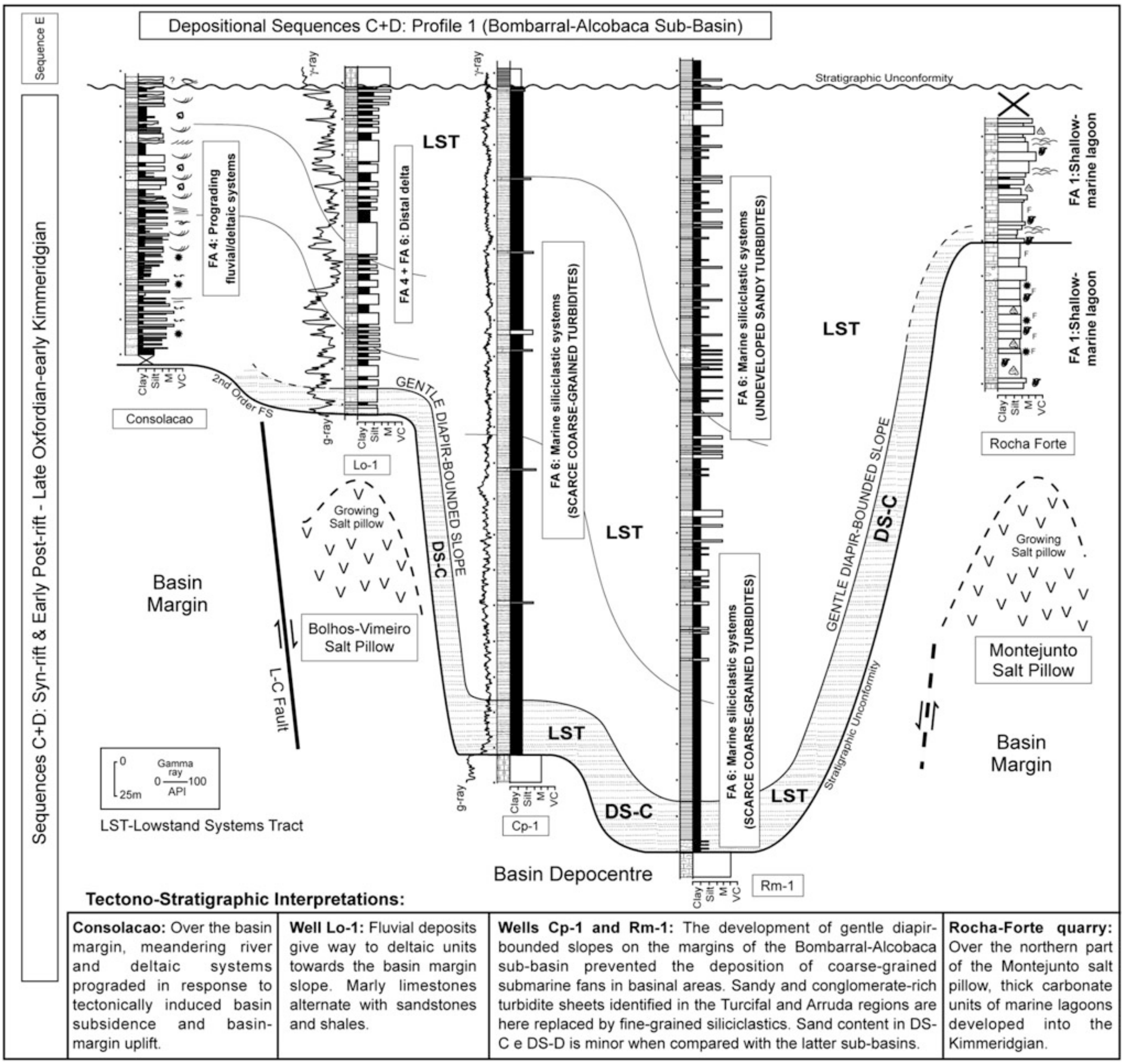

Fig. 6.14 Correlation panel for Depositional Sequences C and D in Bombarral-Alcobaça (see Figs. 6.15 and 6.16 for location of key wells in this correlation panel). Figure modified from Alves et al. (2003a)

that the evaporitic transitional sediments are, in places, more than 1,000 m thick, therefore controlling several later processes in the LB; namely the diapirism and the thin-skinned style of Cenozoic inversion in the context of a passive margin.

It is worthwhile referring that detailed field structural analyses and geological mapping allow us to speculate about the onset of a "pre-Lusitanian Basin". Soares et al. (2012a, b) reported coarse-to fine-grained red continental facies siliciclastic beds arguably of Permian (?)/Lower to Middle (?) Triassic age. Offshore the LB a NW-SE striking, 60 km long to ca. 10 km wide graben, was mapped, which is compatible with the onshore mesoscopic extensional structures (Figs. 6.11 and 6.12). Coincidentally, Palain (1976) reported extensive NE to SW oriented drainage of the Triassic siliciclastics of the AB that could also be associated with NE-SW oriented initial stretching of the WIM and SWIM, perpendicular to the main Variscan orogenic trend (Fig. 6.6).

The second rifting episode in the LB started in the base of the Pliensbachian, and is underlain by a basin wide unconformity (D4 unconformity for Soares et al. 1993;

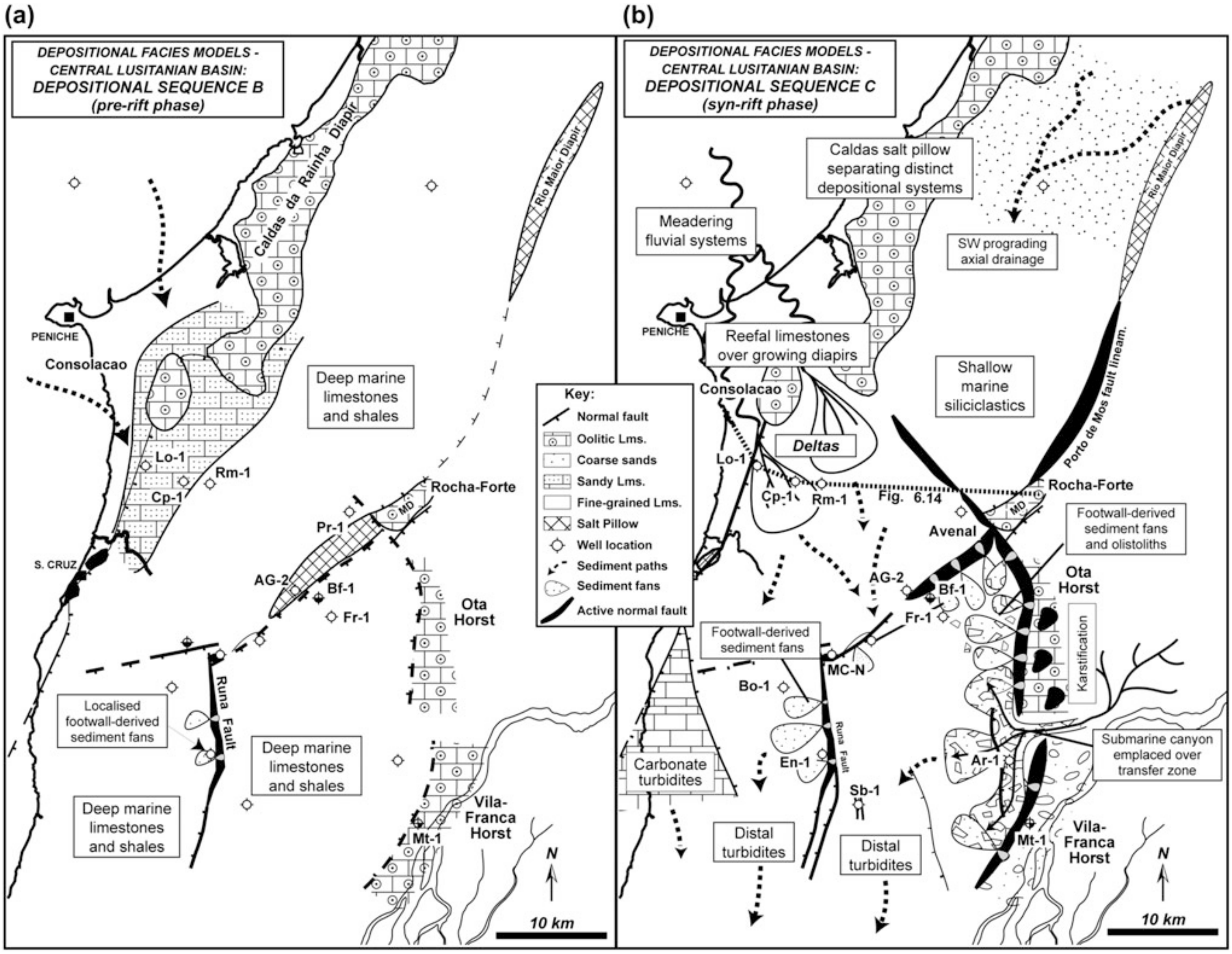

Fig. 6.15 Depositional models for the Central Lusitanian Basin: **a** Middle-Late Oxfordian pre-rift stage (DS-B); **b** Late Oxfordian syn-rift (DS-C) Modified from Alves et al. (2003a)

Rocha et al. 1996; see Fig. 6.13) that marks major modifications in the geometry and consequent facies distribution in the basin. The LB was, at this time, structured as a homoclinal ramp deepening to the West, although the differentiation in sedimentary environments was conditioned by ENE-WSW main transfer faults. The deepening trend in the basin is, at this time, from SE to NW. For some authors (e.g. Wilson et al. 1989; Ravnås et al. 1997) the second rifting episode is bounded at the top by a hiatus in the Mid-Late Jurassic transition (D9), already mentioned for the AB, and by other authors (e.g. Rocha et al. 1996; Terrinha 1998; Azerêdo et al. 2002). This is the most expressive unconformity in the Lusitanian Basin, showing a basinwide expression, and represents a hiatus of about 3 My from the upper Callovian to the Middle Oxfordian. It is expressed as: (i) an angular unconformity, (ii) an erosional surface, or (iii) a paraconformity with a sudden facies change, but no evidence for condensation horizons or clastic influx. The origin of D9 (Fig. 6.13) has been discussed and different possible causes for its occurrence have been proposed, all involving tectonics at a regional, supra-basinal scale, namely: (i) regional uplift in connection with opening of the Central Atlantic (Rasmussen et al. 1998); (ii) tectonic inversion caused either by thermal uplift of a distal stretched portion of the lithosphere or by distal compression induced by thermal subsidence (Terrinha et al. 2002); (iii) regional factors, including climatic, which intensified in Iberia the global sea-level regressive trend recorded at this time (Azerêdo et al. 2002). After this event, the basin changed progressively from lacustrine to a marine deep-external carbonate platform topped by turbidites. At this moment the basin experienced an acceleration of the subsidence rate (up to 200 m/My), anticipating a sudden and widespread modification of the basin morphology.

The third rifting episode is marked by unconformity D10 of earliest Kimmeridgian age (Rasmussen et al. 1998;

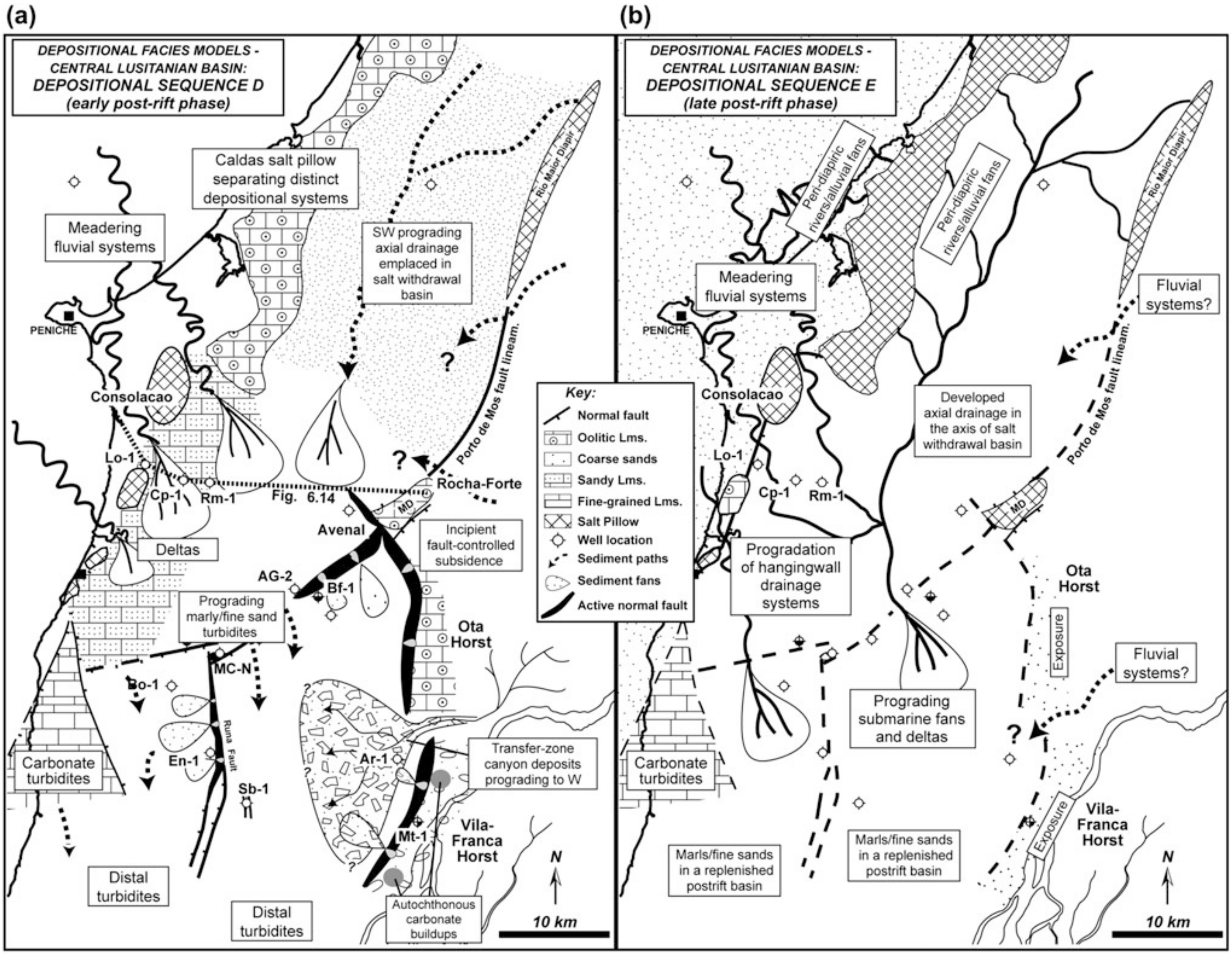

Fig. 6.16 Depositional models for the Central Lusitanian Basin: **a** latest Oxfordian-early Kimmeridgian early post-rift stage (DS-D); **b** upper Kimmeridgian DS-E (late post-rift). (modified from Alves et al. 2003a)

Kullberg 2000; Kullberg et al. 2014) that marks the most important and sudden morphological change of the basin topography and facies modifications. These changes were associated with a strong increase of lithospheric stretching that favoured a generalized influx of siliciclastic sediments fed from the exposed Variscan basement. For the first time in the LB's evolution, there is clear evidence of a proximal rift shoulder in its Eastern margin, in a vast extent materialized by the Porto-Tomar Fault. The proximal marginal sequences (to the East and the West) are marked by arkosic gravels deposited in a submarine fan system that is more than 2,200 m thick (Wilson et al. 1989), while in the distal (deeper) parts of the basin epibathyal clays were deposited (approximately 1,000 m thick). The whole basin shows a trend for progressive infill, firstly with peri-reefal carbonates during the Late Kimmeridgian and secondly by continental fluvial siliciclastics until the end of the Tithonian. Important sub-basin development occurred during the third rifting episode, particularly in the Central sector, where the Bombarral, Turcifal and Arruda sub-basins were separated by deeply rooted normal to vertical faults (Ravnås et al. 1997; Alves et al. 2003a, b).

Although good stratigraphic markers are scarce in the continental siliciclastic deposits of Tithonian age onwards, it is generally accepted that the onset of the fourth rifting episode occurred during the Berriasian and lasted until de Aptian-Albian when the rift-to-drift transition of the Iberian Abyssal Plain probably occurred. The fourth rifting phase is marked by a basal regional unconformity characterized by an angular unconformity over tilted half-grabens overlain by a siliciclastic sequence of 300 to 500 m in thickness. In the offshore area it is not recognized to the North of the Nazaré transfer fault in the Lusitanian Basin, but occurs in exploration wells (and seismic data) from the Porto Basins (Moita et al. 1996; Alves et al. 2006). According to Rasmussen et al. (1998), this regional unconformity is probably due to thermal uplift induced by lithospheric stretching during the final rifting phase that generally precedes crustal separation (Ziegler 1992). In fact, the initiation of the fourth rifting episode can be related to the formation of oceanic crust in the Tagus Abyssal Plain (Pinheiro et al. 1996; Alves et al. 2003a; Soto et al. 2012; Alves and Cunha 2018) and the Tithonian-Berriasian magmatic event in the LB—also, rift-shoulder uplift as recorded on the multichannel seismic reflection data and at DSDP SITE 398 (Alves et al. 2006, 2009).

6.7 WIM Rifting Events: The Alentejo Basin

Kullberg JC and **Alves T**

The Alentejo Basin is located in the Southern part of the West Iberian Margin and just a small part of this basin crops out onshore—the Santiago do Cacém Basin (Fig. 6.11). The offshore Alentejo Basin has been studied more intensively based on the acquisition of high-quality seismic data for the last few decades. Two wells (Golfinho-1 and Pescada-1) were drilled in the 1970's on the continental shelf, constituting the only direct information that exists for the offshore basin. In addition, a large dredge campaign reported by Matos (1979) collected samples of Paleozoic basement, Jurassic-Cretaceous syn-and post-rift and Cenozoic strata (see also Mougenot 1988). This is the only basin of the WIM where the magmatism of the tholeiitic cycle is recognized, both at onshore outcrops (Inverno et al. 1993) and in the core samples from the Golfinho-1 well.

The tectonic limits with the neighbouring basins to the North and the South were pointed by some recent works as being, in the uplifted and exhumed onshore, the Tagus Fault (TF) and the Messejana-Plasencia Fault Zone (MPFZ), respectively (e.g. Pereira and Alves 2013; Pereira et al. 2016). These faults strike NE-SW to NNE-SSW, which is a very oblique direction to the general N-S trend of the WIM basins, including the Alentejo basin itself and the prevailing direction of the normal faults that accommodated the Mesozoic extension. Although the MPFZ seems to be a convincing limit with the Algarve Basin (Pereira and Alves 2013), since it is a major tectonic feature and the Algarve basin has an approximate perpendicular strike (including its internal faults), for the Tagus Fault arguments are missing to support it as the limit with the Lusitanian Basin. In fact, the well exposed lithostratigraphic units, the continuity of the bordering structures, the continuity of depth of the basement at both sides of the Tagus Fault sustain the location of the limit between both basins in the Arrábida Fault (Kullberg 2000; Kullberg et al. 2013).

In a broad approach we can consider that the general geometry of the basement shows some similarities to the central WIM. Interpretation of seismic profiles in the Alentejo basin allows to separate an inner and an outer proximal margin (Pereira and Alves 2013), the first showing a horst-graben geometry, and the second a geometry of rotated blocks, resembling the geometry of the onshore and the offshore Lusitanian basin, respectively.

The few direct observations of the stratigraphic record in the Alentejo Basin allow evidencing some similarities with the neighbouring basins, mainly for the earliest units, although two important differences exist: (1) general thickness is much lower than, for example, in the onshore Lusitanian Basin, even in the Arrábida sector of the LB; (2) a wide important hiatus that affects almost all the Middle Jurassic is other differentiating feature of the Alentejo Basin lithostratigraphy (Fig. 6.17).

A key difference between the (proximal) Alentejo Basin and the LB is that, in the Santiago do Cacém area, syn-rift strata precede a latest Kimmeridgian relative sea level fall (Ramalho 1971; Inverno et al. 1993). Consequently, Tithonian strata are not unequivocally represented at outcrop in southwest Iberia, although dredge data from offshore Alentejo has recovered latest Jurassic-earliest Cretaceous shallow-marine fauna in carbonate samples (Berthou 1977). Tithonian strata comprise shallow marine to transitional carbonate successions at Lagosteiros and well Pescada-1, strata that are lateral equivalent to prograding continental units deposited in the eastern Arrábida region (GPEP 1986) (Fig. 6.18). An equivalent latest Kimmeridgian sea level fall is also recorded in the central part of the Lusitanian Basin but siliciclastic sedimentation predominated in the basin until the earliest Cretaceous (Ravnås et al. 1997).

Offshore, Mesozoic units are relatively thin (less than 1.5 s twtt) in continental shelf basins that are close to the present-day coastline. They are also markedly deformed by pervasive sets of normal faults. High-amplitude folds and pervasive faulting of Mesozoic strata are observed in association with Miocene to Holocene compression. In contrast to the proximal parts of the Alentejo basin, well-developed tilted blocks occur on the outer proximal margin, where they are draped by relatively thick Cretaceous and Cenozoic strata (Fig. 6.19). As with the LB, a major fault system separates tilted blocks on the outer continental margin from the inner proximal margin. This slope-bounding fault system (SFS) delimits the principal locus of Mesozoic subsidence to a 150-km-wide region west of 9° 30 W, oriented along the margin.

The Alentejo basin records three major Mesozoic extensional episodes: Triassic (rift 1), Sinemurian–early Pliensbachian (rift 2) and late Oxfordian-earliest Kimmeridgian? (rift 3) (Leinfelder and Wilson 1998; Stapel et al. 1996; Alves et al. 2002). Late Oxfordian–early Kimmeridgian extension in the Lusitanian basin has been considered the precursor of ocean spreading in the Tagus Abyssal Plain (Wilson et al. 1989) with recent models indicating a Valanginian age for this latter event (Pinheiro et al. 1992).

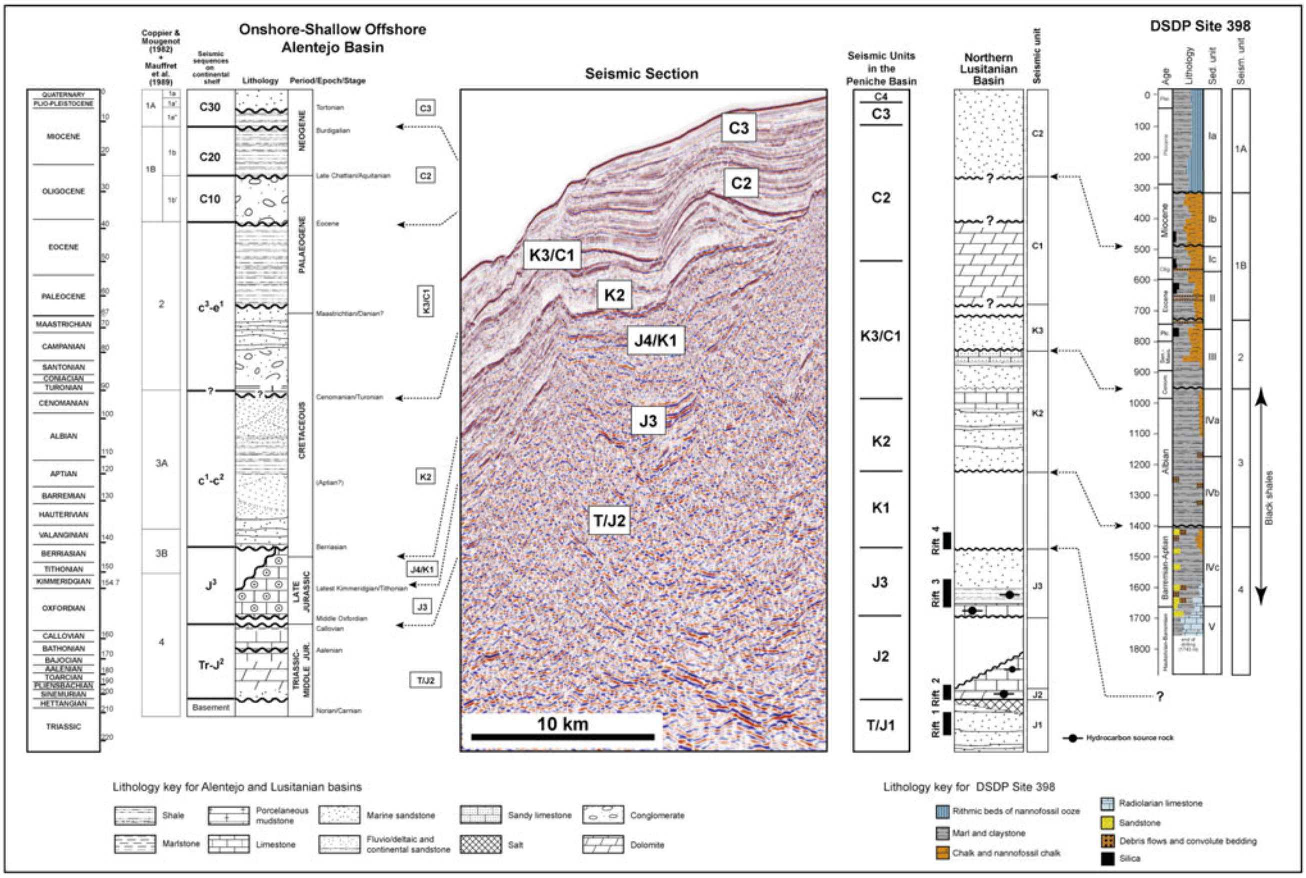

Fig. 6.17 Correlation panel showing interpreted seismic-stratigraphic units from offshore Alentejo and well data from the Northern Lusitanian Basin (Alves et al. 2003) and DSDP Site 398 (Groupe Galice 1979; Réhault and Mauffret 1979). Seismic stratigraphic units on the Alentejo and Lisbon continental shelves are taken from Coppier and Mougenot (1982) and Alves et al. (2003a). Data from the Peniche Basin (Alves et al. 2006) are also correlated with the interpreted units (modified from Alves et al. 2009)

On seismic data, and in the absence of better stratigraphic control, the basal seismic-stratigraphic unit on the WIM comprises Triassic to Mid Jurassic strata, most of which consisting of carbonates and shales underlying Jurassic–earliest Cretaceous synrift strata deposited over tilted blocks (Alves et al. 2009). These strata are overlain by a Late Jurassic syn-rift megasequence (J3) that is preserved in fault-bounded grabens and half grabens on the outer proximal margin. Above this unit is observed a latest Jurassic-earliest Cretaceous sequence that is mainly transparent to low-amplitude and characterized by sub-horizontal reflections. Uppermost Kimmeridgian-Valanginian strata are absent at outcrop (Santiago do Cacém) but comprise shallow marine carbonates in well Pe-1 (Inverno et al. 1993) and dredged units around the Príncipe de Avis Seamounts. This latest Jurassic-earliest Cretaceous unit is overlain by a thick succession of Cretaceous-Cenozoic units that, comparatively, are much more developed than to the North of the Nazaré Fault.

Alves et al. (2009) (Fig. 6.20) have demonstrated the existence of six main sub-basins on the most distal part of the margin—named "outer proximal margin" because the bulk of subsidence in this region is clearly post-rift. The greatest thickness of strata in the Alentejo basin coincides with the location of Late Jurassic depocentres but do not account for the existence of up to 1.7 s twtt of Triassic–Middle Jurassic strata in underlying pre-rift successions. Hence, continental slope basins offshore the Alentejo can comprise more than 9.0 km ($\sim$5.0 s twtt) of strata, which include Triassic–Middle Jurassic successions at the base of rotated tilted blocks. This value is similar to that recorded in the deeper parts of the Peniche Basin (Alves et al. 2006).

Developed salt diapirs and associated minibasins as those observed offshore Peniche have not been recognized in the Alentejo basin. The absence of developed salt structures also contrasts with the deep-offshore area east of the Jeanne D'Arc basin (e.g. Jansa et al. 1980), with the Algarve Basin (Terrinha 1998; Ramos et al. 2017c) and with parts of the

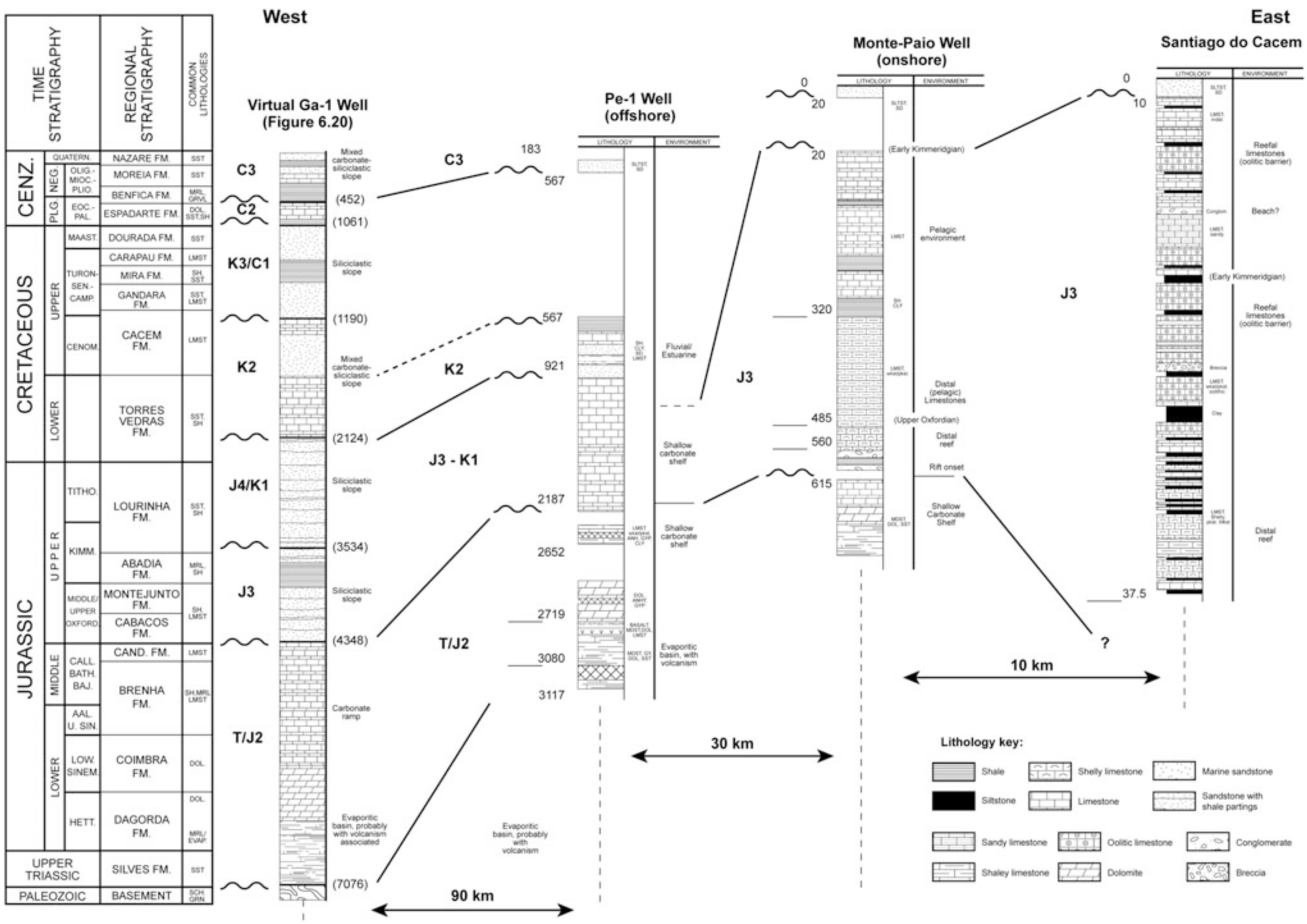

Fig. 6.18 Correlation panel between relevant outcrop and well data from onshore and offshore Alentejo basin. The relative location of virtual well Ga-1 (Gamba-1) is shown in Fig. 6.20 (modified from Alves et al. 2009)

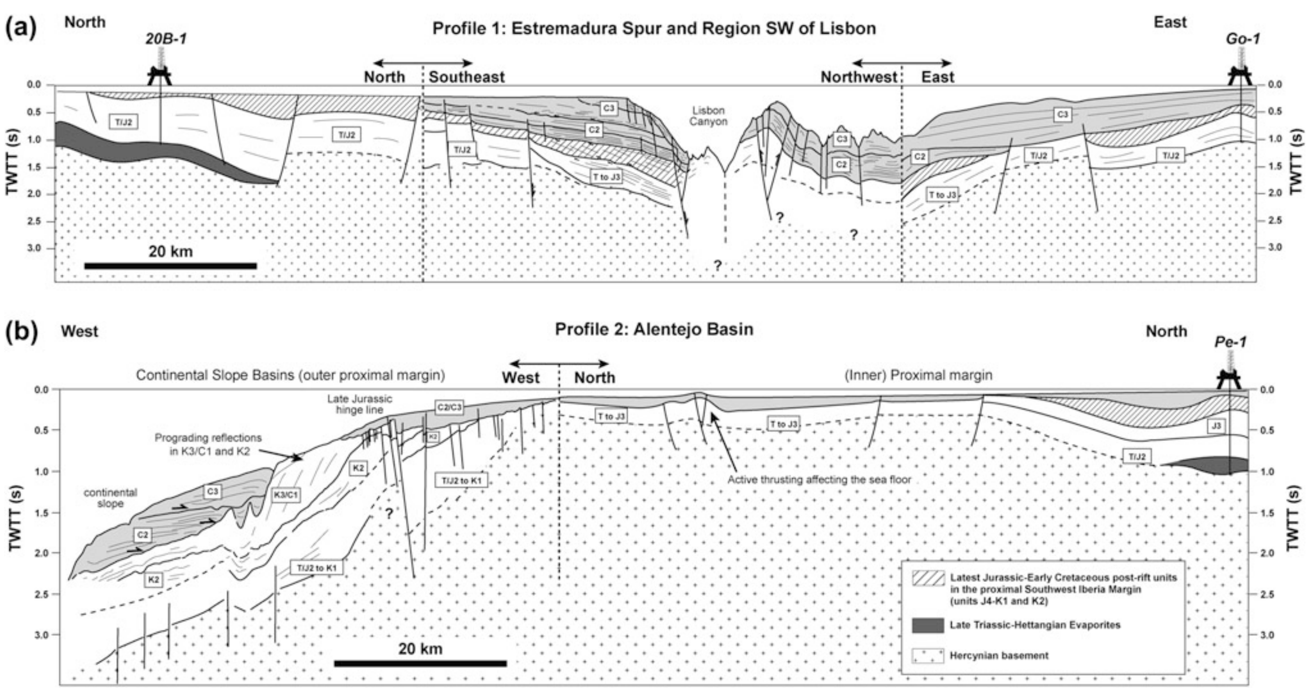

Fig. 6.19 Line drawings showing interpreted seismic profiles across the Southwest Iberian margin (see Fig. 6.20 for location of key well in this figure). The diagrams used 2D seismic data from the proximal margin and were tied to well data whenever possible (modified from Alves et al. 2009)

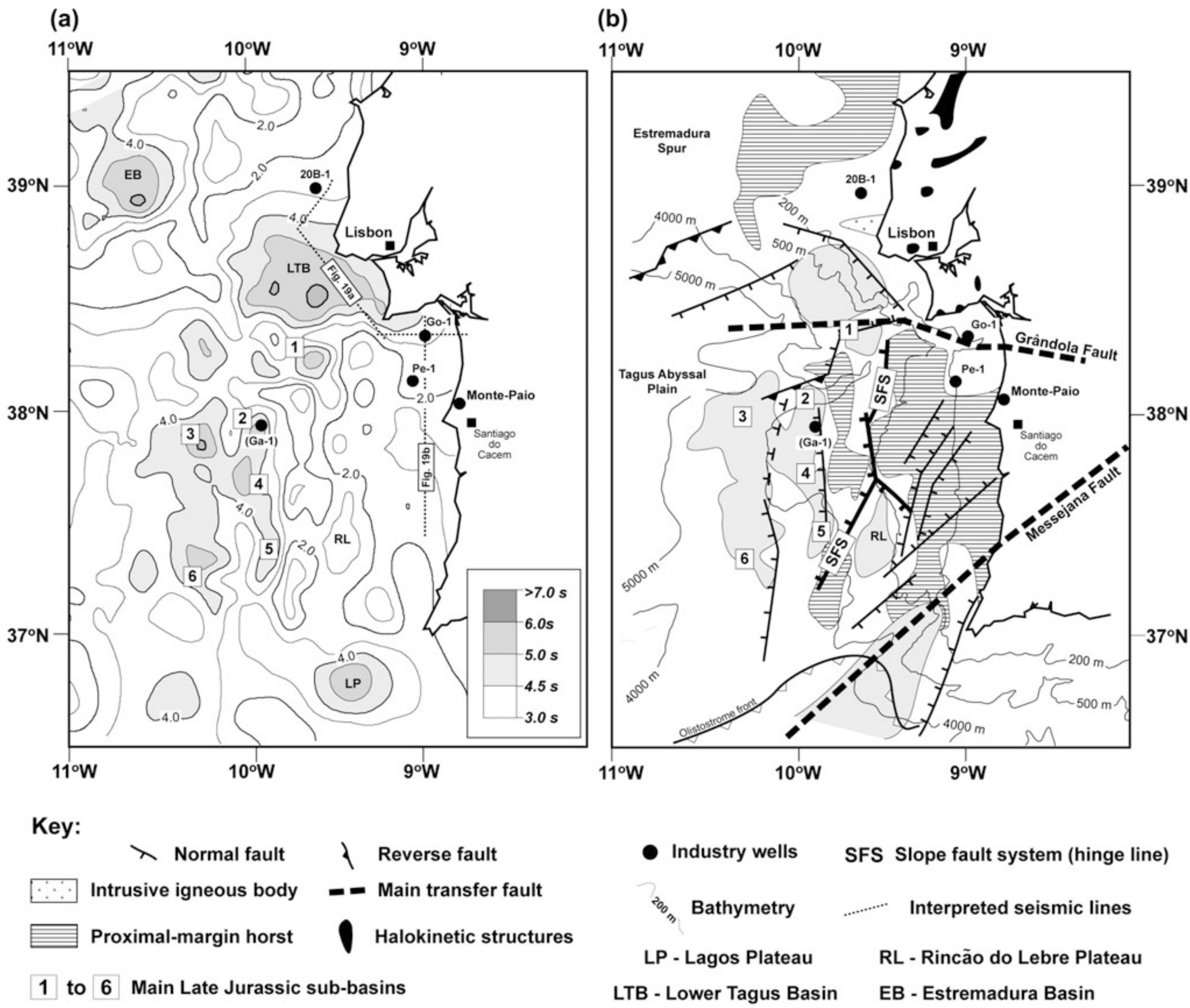

Fig. 6.20 (a) Near base Late Jurassic TWTT isochron map and (b) corresponding structural map of the Alentejo Basin. See text for a detailed description of these maps (modified from Alves et al. 2009)

Lusitanian Basin (Alves et al. 2003b) but correlates positively with outcrop information from the Santiago do Cacém area where Triassic-Hettangian evaporites are absent (Inverno et al. 1993). These are replaced by continental to shallow marine successions at Aldeia dos Chãos, south of Santiago do Cacém. At this location, lowest Jurassic basalt sheets and small-scale intrusions highlight a period of intense magmatism at the start of the Jurassic syn-rift succession. This magmatic pulse is confined to the Alentejo and Algarve basins, i.e. to the region south of Lisbon and is associated with CAMP magmatism (Kullberg 2000; Martins et al. 2008; Kullberg et al. 2013).

In summary, the Alentejo Basin is marked by intense syn-rift movements on a carbonate-dominated setting. Similar seismic-stratigraphic markers to the Lusitanian Basin are recorded here, except for the fact that post-rift (i.e. Berriasian and younger) strata are much thicker in continental slope basins when compared to Peniche or Estremadura. Carbonates dominate the proximal (shallow) parts of the basin and lead to shales and interbedded siliciclastic units in the deeper parts of SW Iberia, a setting that resembles the AB and demonstrates a clear Tethysian influence.

After a widespread period of syn-rift subsidence, the most significant event in the region occurred after Tithonian–Early Cretaceous continental break-up and comprises the development of a prograding slope system from the Late Cretaceous to the Early Cenozoic. Originally interpreted as a thick olistostrome unit of probable late Cretaceous age (Baldy 1977), this prograding slope system was likely related to regional tilting and thermal uplift in the central

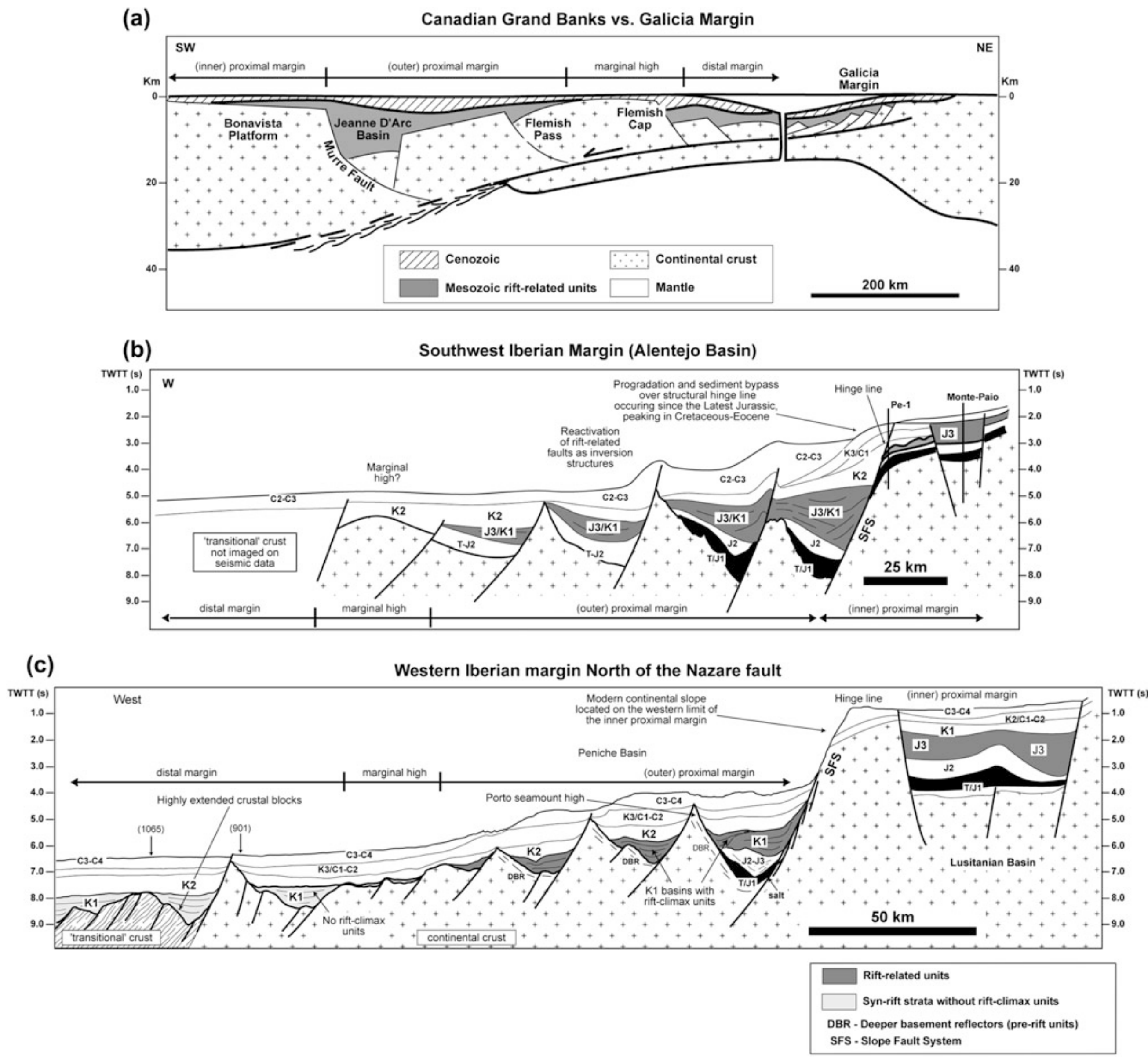

Fig. 6.21 Line drawings depicting (a) the structure of the conjugate Grand Banks-Galicia margin pair (modified from Tankard and Welsink 1987) (b) detailed structure of the Alentejo Basin and adjacent deep-margin; (c) detailed structure of the Peniche Basin and adjacent deep-margin area. The locations of ODP sites are shown in parentheses indicating sites projected onto the line of section. Relevant industry well data is shown in (b)

proximal part of the margin (38°N) during the intrusion of the Sines igneous complex (see Fig. 6.2 for location), an intrusive alkaline complex of gabbro and syenite emplaced near the end of the Cretaceous in the Sines region (Campanian-Maastrichtian; Inverno et al. 1993) (Fig. 6.21).

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