Chapter 5
A Geological History for the Alboran Sea Region

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5.1 Introduction

The paleogeography of a marine basin is strongly controlled by its geodynamic evolution, particularly in the case of active basins such as the Alboran Sea. Tectonics are responsible for the widening or narrowing of the marine basin, the modification of bottom surface reliefs through the creation of subsidence zones that can evolve to sedimentary sub-basins, or the formation of positive reliefs—by tectonic deformation, volcanic activity, and diapirism, and even the generation and destruction of oceanic gateways and/or straits to adjacent seas. The modification and formation of such physiographic reliefs can give rise to critical situations for the habitat. Closure of a strait can mean the isolation of biological populations between adjacent seas and could favor the development of endemic species or of new relationships among species. In turn, the opening of gateways can generate episodes of colonization or invasion by non-native species that may leave a print on the biological history of the basin. Therefore, sound knowledge of the geological and geodynamic evolution of a marine basin is needed to face problems involving its current biogeography.

The Alboran Basin is a Neogene-Quaternary basin formed since the late Oligocene through extensional processes, located within the Betic-Rif alpine cordillera and forming part of this orogenic system. The Betic-Rif belt is characterized onland by ranges that form a tight orocline at the junction between the Atlantic Ocean and the Mediterranean Sea. At present, the Alboran Basin has relatively small dimensions, both in length and width (Chap. 6). The basin has been progressively restricted in the last 10 Ma (Tortonian) due to uplift of the adjacent mountainous reliefs, producing the isolation of intramontane and foreland basins (onland) in addition to the current Alboran Sea. Other changes in the basin's dimensions can be traced to shifts in the sea level, most notably desiccation during the Messinian salinity crisis, and eustatism in Pliocene-Quaternary times.

Nowadays, the Alboran Basin is situated along the boundary between the current plates of Africa and Eurasia (Fig. 5.1), and it occupies a strategic position from an ecological/oceanographic standpoint: this basin harbored the connection of water masses and biological populations from the Atlantic Ocean and the Mediterranean Sea.

Thus, ongoing tectonics and the geodynamic evolution tied to differential kinematics of the African and Eurasian plates, plus the Alboran Domain's westward drift since the late Oligocene, are key factors in the configuration of the current basin and continental margins. They likewise determine Atlantic-Mediterranean oceanic

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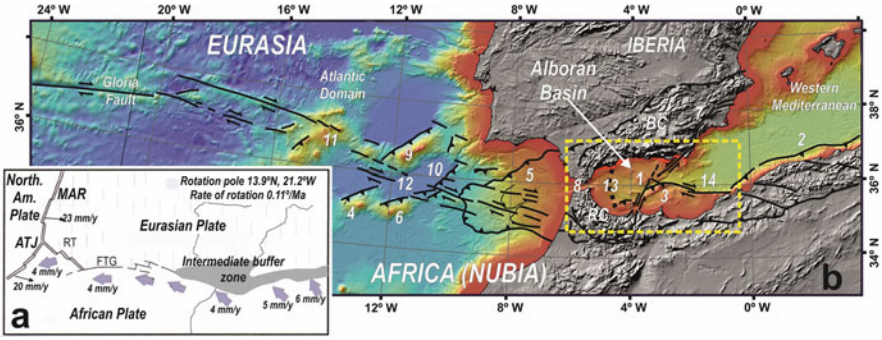

Fig. 5.1 (**a**) Present plate tectonic setting of Africa-Eurasia interaction (taken from Vegas et al. 2008). (**b**) Geodynamic framework between the Western Mediterranean Basin and Gloria Fault in the Atlantic Domain. FTG: Gloria Transfer Fault; MAR: Mid-Atlantic Ridge; RT: Terceira Ridge-Transform plate boundary. BC: Betic Cordillera; RC: Rif Cordillera; 1: Al-Idrissi Seismic Zone; 2: Algerian Margin Subduction Zone; 3: Alboran Ridge Indenter; 4: Ampere Seamount; 5: Allochthonous Unit of the Gulf of Cadiz; 6: Coral Patch Seamount; 7: Eastern Betic Seismic Zone; 8: Gibraltar Arc; 9: Gorringe Bank; 10: Horseshoe Abyssal Plain; 11: Josephine Bank; 12: South West Iberian Margin Fault Zone; 13: Western Alboran Intermediate Seismic Zone; 14: Yusuf-Habibas Fault Zone; Yellow box: Alboran Basin location

dynamics and the sedimentary infilling of basins. The geological and oceanographic framework of this region therefore reflects the dynamics of the Africa-Eurasia plate boundary, the role of the Iberian microplate, and the development of diverse kinematic styles throughout its evolution (Srivastava et al. 1990a; Vergés and Fernàndez 2012).

5.2 Plate Tectonic Settings: Evolution of the African-Eurasian Plate Boundary

To gain a full view of the geodynamic evolution of this region it is necessary to envisage the stages prior to the formation of the basin and understand the influence of African-Eurasian plate interaction. The evolution of these plates during the Mesozoic and Cenozoic—their kinematics and boundary relations—conditioned the formation and evolution of the ocean basins, first the Atlantic and Tethys oceanic basins, and later the Western Mediterranean basin.

The regional background to establish the post-Triassic evolution of the African and Eurasian plates was derived from kinematic reconstructions, adjusting oceanic fracture zones, bathymetric data, and conjugate magnetic alignments (Le Pichon et al. 1977; Srivastava et al. 1990a, b; Roest and Srivastava 1991; Olivet 1996; Schettino and Turco 2009, 2011; Macchiavelli et al. 2017). The geodynamic evolution of the African and Eurasian plates began with the rupture of the supercontinent Pangea at the Permian-Triassic boundary and through middle Jurassic times. Plate reconstructions set Iberia facing Newfoundland and North Africa, approximately between Algeria and Tunisia. During this stage, the Paleotethys oceanic lithosphere was located to the east of the African and Eurasian plates (Stampfli and Borel 2002). Later, relative movements between Africa and Eurasia could be synthesized into four main evolutive phases) (Fig. 5.2) (Dewey et al. 1989; Srivastava et al. 1990a; Schettino and Turco 2011) in which the Iberian plate worked as a single plate or else in association with the African and Eurasian plates.

During the Jurassic (154 Ma) and part of the Lower Cretaceous (120 Ma), the opening of the Central Atlantic and Western Tethys basins took place (Ligurian and Alpine Tethys) (Fig. 5.2a). The motion of Africa with respect to Eurasia (including Iberia) was eastward, implying senextral movements along a plate boundary located south of Iberia. In this paleogeographic scenario, the Central Atlantic opening was prolonged southeast of Iberia into the Ligurian Tethys basin, thus providing for an oceanographic connection between the Atlantic and Tethys oceans, most likely through the Subbetic continental margin (Vegas et al. 2016; Michard et al. 2018; Gómez et al. 2019).

From the Lower Cretaceous (120 Ma) to Late Cretaceous (83.5 Ma), the North Atlantic basin opened, including the Bay of Biscay, whereas the Tethys opening ceased. Hence, processes of divergence and convergence between the two main plates coexisted, maintaining Iberia as an intermediate plate (Fig. 5.2b, c). The opening velocity of the North Atlantic basin was greater than that of the Central Atlantic, causing a relative motion of Africa toward the west (with respect to Iberia and Eurasia), producing dextral movements along the southern plate boundary of Iberia and sinistral movements in the northern boundary (Fig. 5.2b) (Srivastava et al. 1990b). The opening of the Bay of Biscay basin produced an anti-clockwise rotation of Iberia, implying some convergence toward the southeast with Africa, and oblique divergence with Eurasia northward (Sibuet and Collette 1991; Olivet 1996; Vegas et al. 1996; Sibuet et al. 2004; Osete et al. 2011; Vissers and Meijer 2012). The oceanographic connection between the Atlantic and the Tethys basins was maintained in this period, probably through oceanic portals both south and north of Iberia, though in a more restricted gateway to the north (Martín-Chivelet et al. 2019).

Since the Late Cretaceous (83.5 Ma) the opening of the Central and Northern Atlantic basins has continued (Fig. 5.2c). The region was deformed during late Cretaceous-Paleogene times by relative convergence toward the north of Africa with respect to Eurasia (Dewey et al. 1973; Biju-Duval et al. 1978) as a result of South Atlantic and Indian oceanic spreading (Olivet 1996). An orogenic Alpine phase was produced through the subduction of the Ligurian Tethys oceanic lithosphere and the collision of Africa and Eurasia, along with several intermediate continental lithospheric blocks involved in this plate boundary, including minor plates such as Iberia and Apulia (Fig. 5.2c) (Schettino and Turco 2011; Verges and Fernàndez 2012; Vissers et al. 2016). The Pyrenean connection between the Atlantic and the Tethys basins was closed by the convergence between Eurasia and Iberia, and only the south Iberian connection remains.

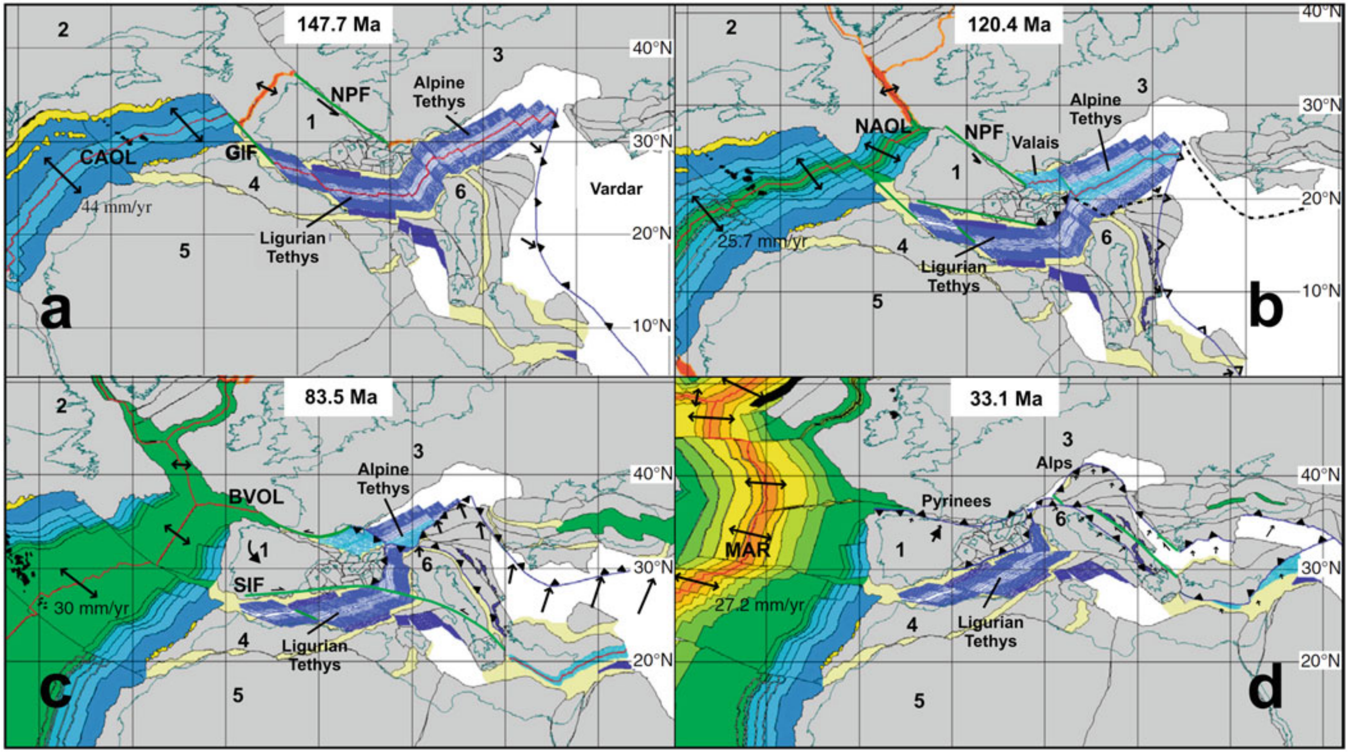

Fig. 5.2 Plate tectonic maps showing the evolution of Africa, Iberia, and Eurasia from the late Jurassic (**a**) to lower Cretaceous (**b**), late Cretaceous (**c**) and middle Oligocene (**d**) times (modified from Schettino and Turco 2011). 1: Iberia; 2: North America; 3: Eurasia; 4: Morocco; 5: NW Africa; 6: Adria; BVOL: Bay of Viscay Oceanic Lithosphere; CAOL: Central Atlantic Oceanic Lithosphere; GiF: Gibraltar Fault; MAR: Mid-Atlantic Ridge; NAOL: North Atlantic Oceanic Lithosphere; NPF: North Pyrenean Fault; SIF: South Iberian Fault

Late Paleocene (33.1 Ma) to Present. The Western Mediterranean oceanic basin began to open inside the western part of the Eocene Alpine Orogen (Fig. 5.2d) in the early Oligocene. The spreading of this basin caused the splitting of several continental fragments (the Alboran domain among them) around it. They collided with the continental margins of southern Iberia, northern Africa, Sicily, and western Apulia to form the Western Mediterranean Belt since the Miocene to the Present Alpine orogenic phase. At this time, the plate boundary between Africa and Eurasia is situated to the south of Iberia and Iberia is part of the Eurasian plate since 10 Ma (Roest and Srivastava 1991).

5.3 Current Africa-Eurasia Tectonics

The relative movements of the current African and Eurasian plates could be explained as an anti-clockwise rotation of Africa with respect to Eurasia around a rotation pole that would be situated within the African plate, in the vicinity of the Canary Islands (Buforn et al. 1988) or the Cape Verde Islands (Argus et al. 1989). This tectonic frontier is characterized by progressively changing kinematics (DeMets et al. 2015), implying a gradual transition from convergent areas eastward to wrench zones and divergent areas westward. The plate boundary is framed to the east by the Dead Sea Transform Fault (Weber et al. 2009), which bounds the African and Arabian plates and connects northward with the Alpine-Himalayan orogenic belt. To the west, this plate boundary ends in the Mid-Atlantic Ridge at the Azores triple junction, which separates the plates of Africa and Eurasia from the North American plate (Fig. 5.1). These tectonic characteristics allow the plate boundary to be divided into three sectors as can be summarized from the literature (Serpelloni et al. 2007; DeMets et al. 2010; Argus et al. 2011).

1. The Eastern Mediterranean Basin between the Dead Sea Transform Fault and the Calabrian Arc. The relative movements between Africa and Eurasia are convergent, directed toward the north (DeMets et al. 1990, 2015), where a relict Tethyan oceanic lithosphere is subducting under Eurasia (Hellenic and Cypriot arcs) and Apulia (Calabrian Arc) (Makris et al. 1986; Jolivet and Faccenna 2000; Speranza et al. 2012; Faccenna et al. 2004, 2014).
2. The Western Mediterranean/Gulf of Cadiz region extends between the Arc of Calabria and the Horseshoe Abyssal Plain (Fig. 5.1). It includes the Western Mediterranean basin, the Gibraltar Arc, the Gulf of Cadiz, and the submarine highs around this abyssal plain (Vázquez and Vegas 2000). The convergence direction between Africa and Eurasia progressively rotates from an NW trend in the central Mediterranean to a WNW trend in the Gibraltar Arc and Gulf of Cadiz areas (Dewey et al. 1989; Reilly et al. 1992).

   The plate boundary in this area could be traced by southward subduction along the Algerian continental margin (Aïdi et al. 2018; Hamai et al. 2018), continuing to the west along the Yusuf-Habibas right-lateral transfer fault, the northern

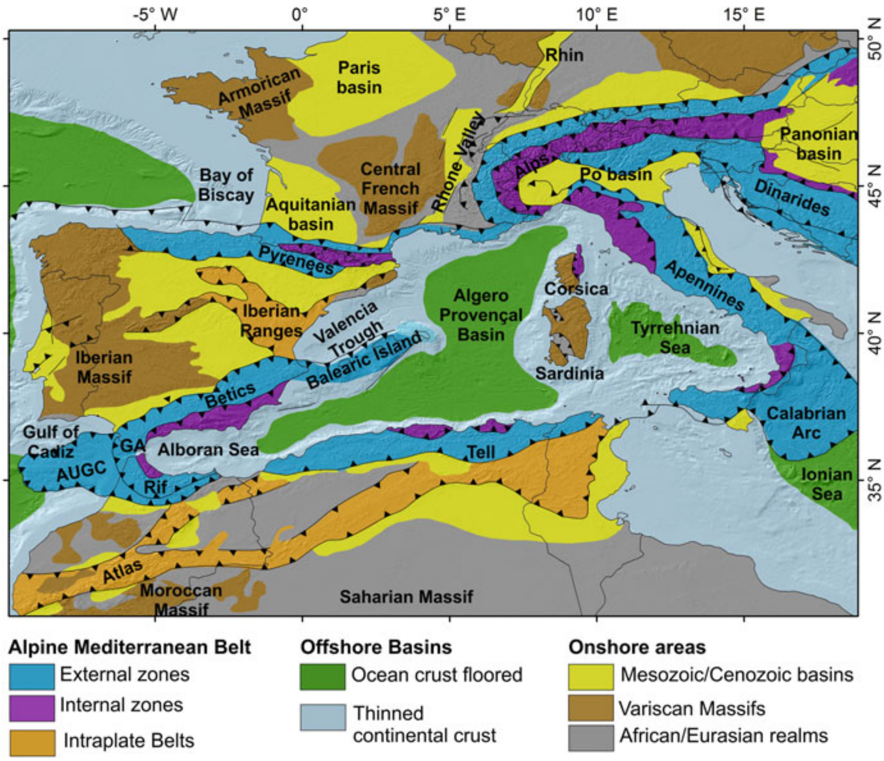

Fig. 5.3 The Western Mediterranean Alpine belt (modified from Sanz de Galdeano and Vera 1992; Faccenna et al. 2014; Leprêtre et al. 2018). GA: Gibraltar Arc; AUGC: Allochthonous Unit of the Gulf of Cadiz

reverse flank of the Alboran Ridge, the Al-Idrissi left-lateral transfer fault (Estrada et al. 2018), and compressive structures at the southern limit of the Rifian belt (Jimenez-Munt et al. 2003; Chalouan et al. 2006, 2014; Pedrera et al. 2011) (Fig. 5.1). In the Gulf of Cadiz sector, the contemporary WNW-ESE convergence produces the SWIM right-lateral transfer zone as well as compressive structures of NE trend that affect the Miocene Allochthonous Unit of the Gulf of Cadiz and several oceanic features (Medialdea et al. 2004; Zitellini et al. 2009; Terrinha et al. 2009; Rosas et al. 2009, 2012; Duarte et al. 2011; Cunha et al. 2012; Neres et al. 2016; Hensen et al. 2019) (Figs. 5.1 and 5.3).

Some debate is related to the relevance of current subduction processes in this sector. Proposals include slow subduction of the Atlantic oceanic lithosphere under the Gibraltar Arc in relation to the Betic-Rif wedge and seismic tomography characteristics of the mantle under the Alboran Sea (Gutscher et al. 2002; Spakman and Wortel 2004; Gutschter et al. 2012, Spakman et al. 2018) or incipient subduction processes associated either with Gorringe Bank or with the transition between the continental and oceanic lithosphere (Maldonado et al. 1999; Duarte et al. 2013).

3. The Atlantic Domain located approximately to the west of Josephine Bank is characterized by the Gloria transform fault (Argus et al. 1989), as demonstrated by the earthquake focal mechanisms analyzed in the region (Buforn et al. 1988; Custódio et al. 2016). This fault has a slightly curved geometry as corresponds to the Africa-Eurasia sliding vector in this sector (Neres et al. 2016; Hensen et al. 2019). The Gloria fault connects to the west with the Azores triple divergent junction between North America, Eurasia, and Africa. The Africa-Eurasia boundary (Terceira Ridge) is directed ESE and interpreted as a transfer-divergent system (Fig. 5.1a) characterized by a transtensive (oblique divergence) regime (Madeira and Ribeiro 1990; Madeira et al. 2015).

5.3.1 Western Mediterranean

The Western Mediterranean region features elongated Alpine orogenic belts, to both the north and south, surrounding marine basins floored by thin continental or Cenozoic oceanic lithospheres (Fig. 5.3). This region has undergone widespread plate tectonic movements since the early Oligocene, comprising divergence and convergence. This area's complex geodynamic evolution entails plate subduction, slab fragmentation, slab rollback and delamination (Lonergan and White 1997; Faccenna et al. 2004; Jolivet et al. 2009; Carminati et al. 2012), escape (Chalouan et al. 2006) and indentation tectonics (Estrada et al. 2018). These large-scale lithospheric movements occurred within a context of continuous Africa-Eurasia convergence (Dewey et al. 1989; Vissers and Meijer 2012) and caused intense crustal deformation in the surrounding orogens (Jolivet et al. 2009). The Western Mediterranean Alpine belt is constituted by several arched orogenic belts. In the westernmost Mediterranean region, the Rif and the Betic Cordillera are connected through the Gibraltar Arc and they extend to the Gulf of Cadiz, forming a tectonic/gravitational wedge known as the Allochthonous Unit of the Gulf of Cadiz (Medialdea et al. 2004). Northward, the Iberian Ranges and the Pyrenees constitute the main alpine chains, which finally connect with the Alps. The Apennines constitute the eastern boundary and delineate a close orocline in the Calabrian Arc, finally extending through northern Africa toward the Tell, in continuity with the Rif.

Development of the Western Mediterranean Alpine belt was mainly governed to progressive southward slab retreat detachment associated with northwestward subduction of the former oceanic lithosphere attached to the African plate, and the development of back-arc basins in an initial Eocene alpine orogenic phase (Cohen 1980; Rehault et al. 1985; Carminati et al. 1998a, 2012; Gueguen et al. 1998; Rosenbaum et al. 2002a; Faccenna et al. 2004). After 35 Ma the northwestward subduction and detachment process of the former oceanic Ligurian Tethys (Carminati et al. 2012) caused extensional processes on the upper plate and formed the Western Mediterranean basin since the early Oligocene. These processes separated and rotated (anti-clockwise) Corsica and Sardinia lithospheric blocks, thereby opening the Algero-Provencal basin as the main oceanic back-arc basin along the Miocene (20–10 Ma) between the Balearic Islands, Tell, Corsica, and Sardinia.

The Alboran Sea, most of the Tyrrhenian Sea (between the Apennines and Corsica-Sardinia block), and the Valencia Trough, developed by the clockwise rotation of the Balearic Promontory with respect to Iberia, constitute the main sedimentary basins formed on strongly stretched thinned continental crust. Moreover, small areas of the Tyrrhenian Sea are floored by Pliocene-Quaternary oceanic crust. The absence of clear seafloor linear magnetic anomalies (Galdeano and Rossignol 1977) impedes a detailed study of the oceanic spreading.

Deformation continues up to the consumption of the Ligurian Tethys oceanic lithosphere (Fig. 5.2) and collision of the accretionary prisms with the forelands that represent the main alpine belts surrounding the Western Mediterranean (Jolivet and Faccenna 2000; Handy et al. 2010). In this context, the Alboran Sea is generated in the interior of the Betic-Rif orocline by extensional forces linked to the contraction resulting from African-Eurasian convergence, which continues up to Present, with the coetaneous retreating of an eastward-dipping slab below the Gibraltar Arc (Pedrera et al. 2011; Faccenna et al. 2014; Molina-Aguilera et al. 2019).

5.3.2 Evolutive Models

In the literature, no consensus exists regarding the tectonic evolution, however, the scientific community agrees that a major change occurred in the subduction regime during the Oligocene (Faccenna et al. 1997; Jolivet and Faccenna 2000). In the Western Mediterranean basin, several hypotheses approach the lithosphere dynamics. They could be synthesized in (1) double westward (Gibraltar Arc) and eastward (Calabrian Arc) slab retreat, (2) slab fragmentation, and (3) continental delamination. These models have been proposed to explain the formation of forearc and back-arc basins, involving the extension, exhumation of metamorphic core units in internal zones and collapse of orogens (Royden 1993; Lonergan and White 1997; Jolivet and Faccenna 2000; Wortel and Spakman 2000; Faccenna et al. 2004; Spakman and Wortel 2004; Handy et al. 2010). In response to slab retreats, several back-arc basins have opened in the Western Mediterranean region: the Liguro-Provençal basin, Algerian basin, Alboran Sea, and the Tyrrhenian Sea, among others (Faccenna et al. 2001).

The Alboran Sea remains one of the most controversial issues in Western Mediterranean geodynamics. The present-day complex geometry of seismic tomography anomalies in the mantle show clear evidence of subducted remnants resulting from progressive slab tearing and detachment (Carminati et al. 1998a, b; Wortel and Spakman 2000; Bezada et al. 2013; Spakman et al. 2018; Molina-Aguilera et al. 2019) associated with a complex mantle convection pattern (Faccenna et al. 2004; Spakman and Wortel 2004; Jolivet et al. 2009; Faccenna and Becker 2010; Sternai et al. 2014). Nevertheless, the exact nature of subducted material remains unclear, varying from oceanic lithosphere to hyper-extended margin (Vergés and Fernàndez 2012; van Hinsbergen et al. 2014).

Numerous hypotheses have been proposed to explain the formation of the Alboran Basin in the context of the surrounding Betic-Rif orogen: (1) extensional collapse of a thickened crust due to convective removal of the continental lithospheric mantle (Platt and Vissers 1989; Platt et al. 2003, 2013); (2) continental lithosphere delamination processes (Seber et al. 1996; Calvert et al. 2000); (3) W-SW retreat of the subduction zone (Royden 1993; Lonergan and White 1997; Gutscher et al. 2002); (4) W-NW retreat of the subduction zone (Vergés and Fernández 2012); or (5) complex models which integrate subduction, delamination event and rollback westward as Bezada et al. (2013).

The geodynamic evolution of the Alboran Domain—comprising the basement of the Alboran Sea as well as the internal domains of the Betic and Rif cordilleras—is mostly controlled by the extension and exhumation of metamorphic core units, driven by the westward retreat of the subduction zone that interplays with African-Eurasian convergence. The metamorphic basement of the Betic Cordillera is made up of three stacked metamorphic complexes, from bottom to top: the Nevado-Filabride, Alpujarride, and Malaguide (Fig. 5.4). In the Rif Cordillera, where no equivalent of the Nevado-Filabride complex exists, the Alpujarride and Malaguide complexes are, respectively, known as the Sebtide and Ghomaride complexes. Today, these metamorphic complexes are separated by crustal-scale extensional detachment shear zones (Galindo-Zaldívar et al. 1989; García-Dueñas et al. 1992; Lonergan and Platt 1995; González-Lodeiro et al. 1996; Augier et al. 2005a; Platt et al. 2005). Exhumation of the Alpujarride/Sebtide complex occurred during the lower Miocene (22–18 Ma) in an N–S to NNE–SSW extensional setting (Monié et al. 1994; Crespo-Blanc et al. 1994; Crespo-Blanc 1995; Kelley and Platt 1999; Platt et al. 2005) while exhumation of the Nevado-Filabride complex occurred from the Lower to Upper Miocene (20–9 Ma—de Jong 1991; Monié and Chopin 1991; Johnson et al. 1997; Augier et al. 2005b; Platt et al. 2005; Vázquez et al. 2011) through ~E-W regional-scale extension (Galindo-Zaldívar et al. 1989; Jabaloy et al. 1992). Two nearly perpendicular superimposed directions of stretching in the basement nappes complexes (Alpujarrides and Nevado-Filabrides) show a progressive yet drastic change in the direction of extensional crustal-scale shear from N-S to E-W, which occurred between 20 and 14 Ma (Jolivet et al. 2008). The exhumed low-angle normal faults are inactive at present; yet the present-day tectonic motion determined by GPS measurements would indicate that E-W extension continues in the Betic Internal Zones (Mancilla et al. 2013; Galindo-Zaldívar et al. 2015a, b; Gil et al. 2017), where it interplays with NW-SE compression that affects the entire Alboran Domain under continual Africa-Iberia convergence.

After the late Oligocene, several sedimentary basins formed above the basement in response to extensional deformation (Galindo-Zaldívar et al. 2019). In the Betics onshore, two main generations of sedimentary basins were formed: the first subsidence pulse took place early in the Aquitanian-Burdigalian up to the Langhian, and the second during the Serravallian-Tortonian (Sanz de Galdeano and Vera 1992; Vissers et al. 1995; Vera 2000). The first generation of sedimentary basins lies

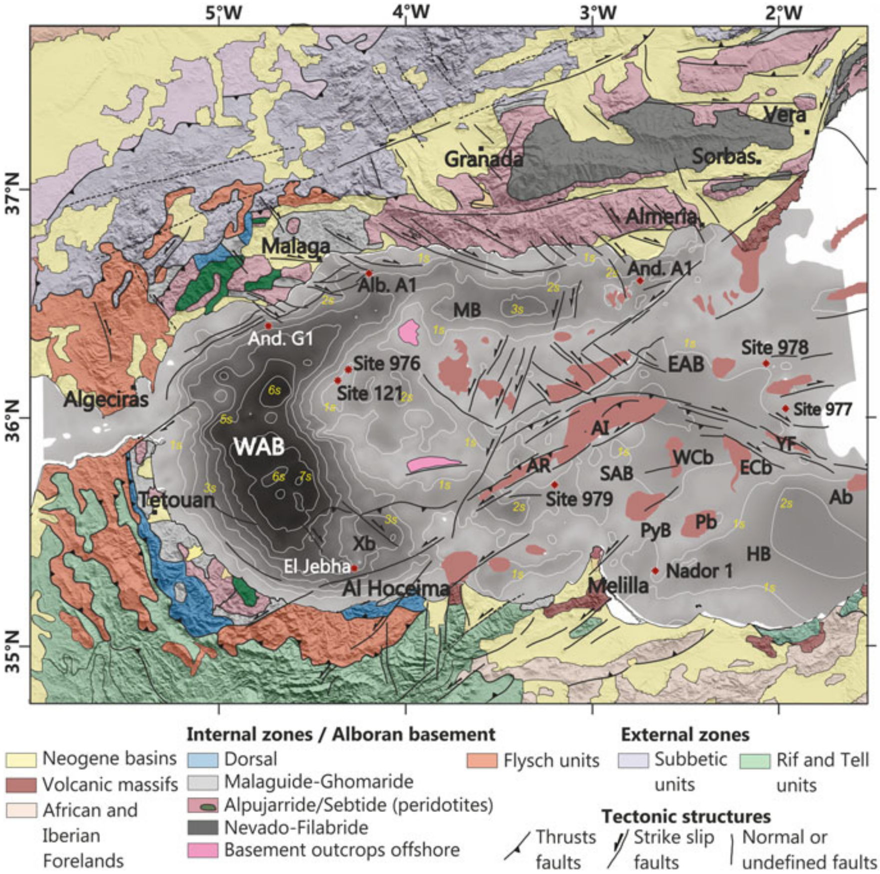

Fig. 5.4 Alboran Sea Region geological map showing major tectonic features of the Betic-Rif belt. Gray shaded background of the Alboran Sea marks the thickness of the sedimentary pile (given in seconds of two-way travel time as interpreted from the seismic records, from <1s, the lightest, to >7s the darkest). ODP sites 976–979, DSDP site 121, and industrial wells are located offshore (And G1: Andalucía-G1; Alb A1: Alborán-A1; And A1: Andalucía-A1; El Jebha and Nador 1). Ab: Alidada Bank; AR: Alboran Ridge; AI: Alboran Island; EAB: Eastern Alboran Basin; ECb: East Cabliers Bank; HB: Habibas Basin; MB: Motril Basin; PyB: Pytheas Basin; SAB: South Alboran Basin; WAB: Western Alboran Basin; WCb: West Cabliers Bank; Xb: Xauen Bank; YF: Yusuf Fault. Modified from Do Couto et al. (2016)

unconformably over the Malaguide-Alpujarride basement and is associated with E-W trending, generally top-to-the-E and top-to-the-N extensional structures (Crespo-Blanc 1995; González-Lodeiro et al. 1996; Suades and Crespo-Blanc 2013). The oldest sedimentary series rework pebbles, blocks, and olistoliths of basement rocks, thus evidencing denudation and intense erosion accompanying the tectonic activity (Serrano et al. 2007; Suades and Crespo-Blanc 2013). The asymmetric geometry of the sedimentary basins, the transgressive character of the older sedimentary series, and the expansion and subsidence of the depocentres demonstrate the intense tectonic activity of this period (Serrano et al. 2007).

The second generation of sedimentary basins formed under a top-to-the-WSW extensional regime (Sanz de Galdeano and Vera 1992; Vera 2000; Rodríguez-Fernández et al. 2011). These basins, defined as intramontane basins (e.g., Sorbas Basin) are mostly located in the southeastern Betics and have similar stratigraphic successions (Rodríguez-Fernández et al. 2011). Extensional structures observed at the border of the basins or within the earlier deposits lead some authors to link their initiation with the latest exhumation stages of the Nevado-Filábride complex (Crespo-Blanc 1995; Meijninger and Vissers 2006; Rodríguez-Fernández and Sanz de Galdeano 2006; Augier et al. 2013; Do Couto et al. 2014). In detail, the asymmetrical distribution of the sedimentary series (e.g., Sorbas, Alpujarras corridor) combined with paleostress analysis show that numerous normal faults accommodated the subsidence of the basin.

Ongoing debates surround the amount of westward movement of the Alboran Domain and its palaeogeographic origin (Jabaloy-Sánchez et al. 2019), the occurrence of STEP faults (Subduction-Transform-Edge-Propagator) (Mancilla et al. 2012, 2013, 2015; d'Acremont et al. 2020), the vergence of the subduction slab and the direction of spreading in the Algerian oceanic basin (Fig. 5.5). In the Alboran Region, recent geophysical measurements show that a remnant of relatively cold lithosphere lies beneath the Gibraltar arc, under the subduction zone, displaying an overall curved shape anomaly that reaches a depth of 600 km (Bezada et al. 2013). The curvature of this anomaly roughly mimics the orogenic arc, hence its length suggests the extent of the Tethys lithosphere slab retreat (Bezada et al. 2013).

5.3.3 Magmatism

The main positive reliefs on the Alboran seafloor are of igneous rocks. The late Eocene—Pleistocene magmatism in the Alboran Domain is related to the collisional orogenic processes resulting from Africa-Eurasia convergence (Duggen et al. 2004, 2008). A roughly 200 by 500 km NE–SW trending belt of middle Miocene to Pleistocene volcanic rocks crops out in this region, which extends from southeastern Spain through the central-eastern Alboran Sea into northeastern Morocco (Fig. 5.6). The igneous activity took place in four main stages (Duggen et al. 2004, Gill et al. 2004; El Azzouzi et al. 2014) that are explained mostly in a context of an eastward subduction and westward roll back of predominantly oceanic lithosphere model (Lonergan and White 1997; Duggen et al. 2003). Nevertheless, there is still discussion about the mechanisms, convective removal of thickened lithosphere and delamination of the continental lithospheric mantle have also been proposed to explain in part the magmatism (Platt et al. 1998; Duggen et al. 2005).

Stage 1 It corresponds to the intrusion of tholeiitic to calc-alkaline dyke swarms at the Malaga area (Fig. 5.6) during the late Eocene to lower Miocene (Turner et al.

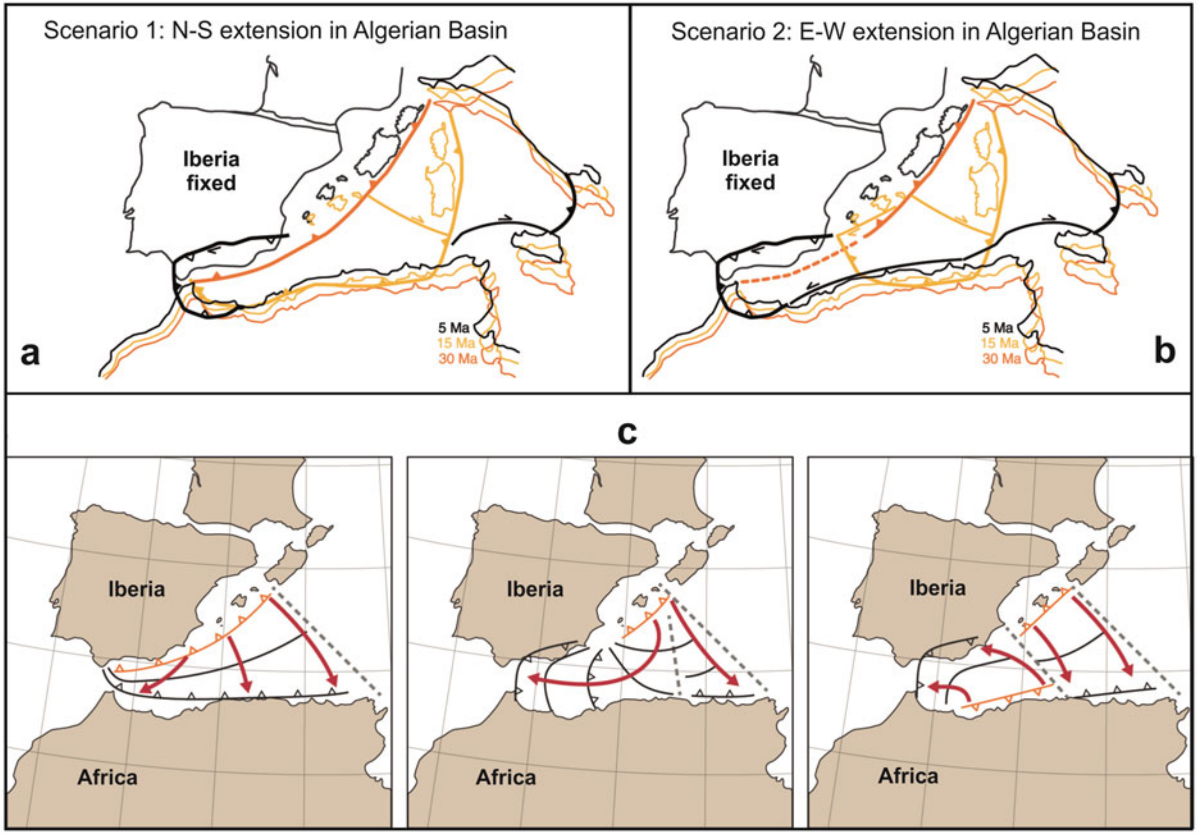

Fig. 5.5 Top: Schematic illustration of the relative position of the subduction zones in the Western Mediterranean since the Oligo-Miocene, assuming (**a**) N-S extension or (**b**) E-W extension in the Algerian basin (from van Hinsbergen et al. 2014). (**c**) Schematic illustration of the three different reconstruction scenarios shown in the paleogeography at ~35 Ma (from Chertova et al. 2014)

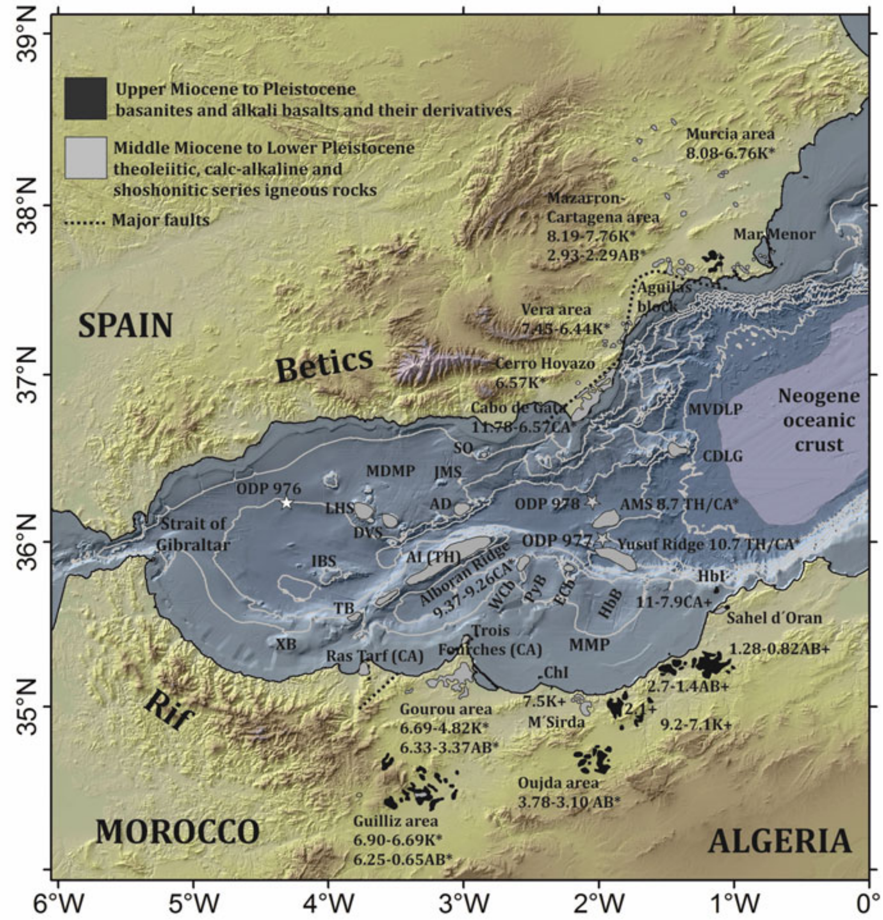

Fig. 5.6 Map of the Alboran magmatic region (westernmost Mediterranean) including southern Spain and northern Morocco. Middle-late Miocene volcanic rocks are distributed in a NE-SW trending belt. Volcanism includes submarine volcanic structures (Al Mansour Seamount—AMS, Yusuf Ridge, Alboran Ridge and ODP Leg 161, sites 977 and 978), Alboran Island (AI), Cabo de Gata and Aguilas area in southern Spain, and Ras Tarf and Trois Fourches in northern Morocco. AR: Adra Ridge; ChI: Chafarines Islands; CDLG: Los Genoveses Ridge; DVS: Djibouti Ville Seamount; ECb: East Cabliers Bank; HB: Habibas Basin; HI: Habibas Islands; JMS: José Medialdea Seamount; LHS: La Herradura Seamount; MDMP: Motril-Djibouti Marginal Plateau; MMP: Moulouya Marginal Plateau; MVDLP: La Polacra high; PyB: Pytheas Basin; SO: El Seco de los Olivos (Chella) Bank; TB: Tofiño Bank; WCb: South western Cabliers Bank. Modified from Duggen et al. (2008)

1999). They were located within the Malaguide Complex and after emplacement (were transported toward the northwest (Torres-Roldan et al. 1986; Martínez-Martínez and Azañón 1997). Geochemistry data and Sr-Nd isotope data favor derivation of the Malaga dykes from a peridotitic (mantle) source through the subduction of oceanic lithosphere (Torres-Roldan et al. 1986; Duggen et al. 2004).

Stage 2 It includes cordierite-bearing volcanic dacite at the Mar Menor, a granitic volcanic clast from ODP Site 977 (Fig. 5.6), and intrusive igneous leucogranites from southern Spain of early Miocene ages (Zeck et al. 1989). These rocks point to a large-scale thermal event occurring in the Alboran region between 23 and 18.5 Ma. There was a partial melting of continental crustal and these rocks formed through crustal anatexis (Munksgaard 1984; Zeck et al. 1989) associated with nappe movement (Platt et al. 2003) during the collision of the Alboran block with the South Iberian and North-African continental margins.

Stage 3 A low-K tholeiitic through high-K calc-alkaline series of volcanic rocks erupted in the Alboran Sea region during the Middle to Late Miocene. Main volcanic complexes onshore include Cabo de Gata, Ras Tarf and Trois Fourches; they correspond to strike-slip fault-bounded blocks on the Alboran Sea margins (Coppier et al. 1989; Aït Brahim and Chotin 1990; Martínez-Díaz and Hernández-Enrile 2004). It also includes most of the volcanic buildings located in the basin: banks of Chella, Djibouti and E and W Cabliers, ridges of Adra, Alborán and Yusuf, and Al Mansour seamount between them (Fig. 5.6). This stage corresponded to the greatest volcanic activity in the basin, especially in the central and eastern sectors, and meant a substantial modification of the seafloor. Age data for Alboran Sea volcanic rocks range from 12.1 ± 0.2 to 6.1 ± 0.3 Ma (Al Mansour Seamount, Alboran, and Yusuf Ridges, ODP 161 Site 977) (Hoernle et al. 1999; Duggen et al. 2008). The volcanism is explained by subduction of oceanic lithosphere in the westward roll back model (Hoernle et al. 1999; Zeck 1996; Duggen et al. 2003; Gill et al. 2004) and the release of hydrous fluids and sediment melts from subducted lithosphere into the mantle wedge beneath the Alboran Basin (Duggen et al. 2004).

Stage 4 The magmatic stage 3 ceased with the occurrence of Late Miocene to Early Pliocene shoshonitic and lamproitic magmatism in Spain and Morocco (Hoernle et al. 1999; Duggen et al. 2003; Gill et al. 2004) that shifted to an intraplate-type, and finally Early Pliocene to Quaternary alkali basalt erupted in southeastern Spain and northern Morocco (Hoernle et al. 1999; Duggen et al. 2003; El Azzouzi et al. 2014). This stage has been explained in the context of a transition from subduction-related to intraplate-type volcanism by lithospheric tearing in the borders of the subducting slab that provides a possible mechanism to explain the shoshonitic volcanism by local melting of the lithospheric mantle (Gill et al. 2004). An associated positive thermal relief is also proposed that overlaps with the Messinian Salinity Crisis (Duggen et al. 2003).

5.4 The Alboran Basin

The Alboran Sea is about 350 km long and 150 km wide, exhibiting a complex seafloor physiography with several ridges and seamounts (Chap. 6). It currently hosts three main sub-basins named after their respective geographic locations: the Western Alboran Basin (hereafter WAB), Eastern Alboran Basin (EAB), and Southern Alboran Basin (SAB). The basins are separated at seafloor by morphostructural features (Fig. 5.4).

5.4.1 Input from Potential Field Data

Gravity and magnetic data constitute what is known as "potential fields" and provide information about the deep structure of a basin. In this case, Bouguer gravity anomalies (Fig. 5.7) were obtained from a global free air dataset (Sandwell et al. 2014) with a resolution of 1 nautical mile. Magnetic anomalies correspond to the second version of the World Digital Magnetic Anomaly Map Project at sea level, which has a resolution of 3 nautical miles (Catalán et al. 2016) (Fig. 5.8). The Analytic Signal (AS) of the magnetic field anomaly is a mathematical expression that highlights the location of magnetic contacts (Roest et al. 1992; Roest and Pilkington 1993; Salem et al. 2002).

The Bouguer gravity anomaly smoothly decreases and narrows from the east (190 mGal) to the west (30 mGal at 4.5° W) in the Alboran Sea (Fig. 5.7). West of 4.5° W it shows low values, yet locally high values are identified at 5° W (M1 and M2, Fig. 5.7) near the African and Iberian coasts respectively.

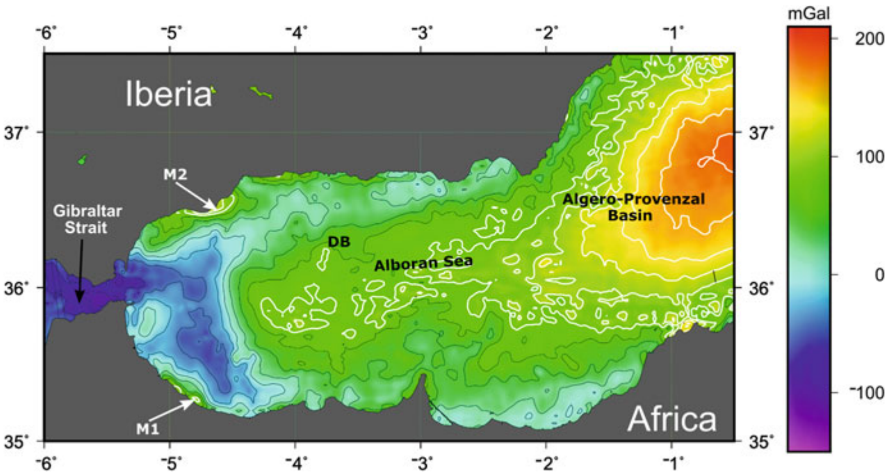

Fig. 5.7 Bouguer gravity anomaly map of the Alboran Sea at 2 km resolution. Contour lines every 20 mGal. M1 and M2: local highs. DB: Djibouti Bank

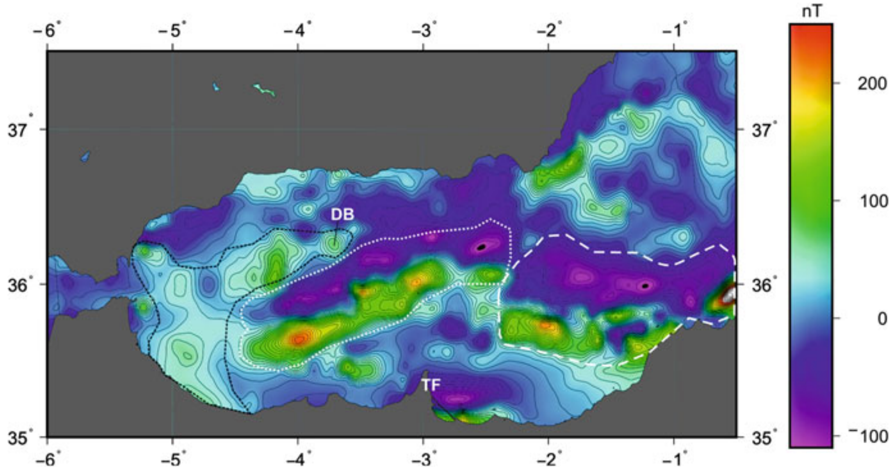

Fig. 5.8 Magnetic anomaly map of the Alboran Sea at 3 nautic miles resolution. Contour lines every 10 nT. Polygons: White dotted lines highlight an ENE-WSW magnetic anomaly; white dashed lines surround a WNW-ESE magnetic anomaly; black dotted line surrounds an N-S magnetic domain turning NE at 4.5° W. DB: Djibouti Banks, TF: Trois Fourches Cape

The Magnetic anomaly map evidences two main ENE-WSW and WNW-ESE linear dipoles (Fig. 5.8) that intersect approximately at 2°30′ W. The ENE-WSW dipole is continuous with an average amplitude of 120 nT, reaching over 200 nT at three locations. The lows show an almost constant value of −80 nT. The NW-SE anomaly shows a more discontinuous picture—its trend changes at 1° W, becoming eastward parallel to the shoreline (Fig. 5.8). East of Trois Fourches Cape, an E-W magnetic anomaly that extends onshore is identified. The westernmost part of the Alboran Basin is dominated by a positive N-S high. North of 36° N this anomaly turns, showing NE trending (Fig. 5.8). This NE elongated anomaly has two peaks geographically correlated with the Djibouti Bank (Fig. 5.8).

Analysis of the ENE-WSW and WNW-SE magnetic anomalies, considering the AS map and the complete Bouguer gravity anomaly map (Fig. 5.9) allows its origin to be determined. The AS signature evidences that the southern boundary of the ENE-WSW magnetic anomalous body lies north of the Alboran Ridge (Fig. 5.9), contrary to what was proposed by Galdeano et al. (1974) but in agreement with Galindo-Zaldívar et al. (1998). This anomaly is also related to a narrow axis of elongated Bouguer anomaly maxima identified at 4.5° W, extending to the east toward the Argelo-Provenzal Basin, whose amplitude is 100 mGal on average (Fig. 5.7), which is interpreted as caused by crustal thinning (Galindo-Zaldívar et al. 1998). The AS signature of the ENE-WSW trending magnetic anomaly fits reasonably well along this axis (seen in thick black contour lines in Fig. 5.9). Its source is located in the lower crust or even beneath the crust.

The AS map suggests an alignment between the WNW-ESE magnetic anomaly and another AS high at the northeast margin that reaches southeast of Cabo de Gata.

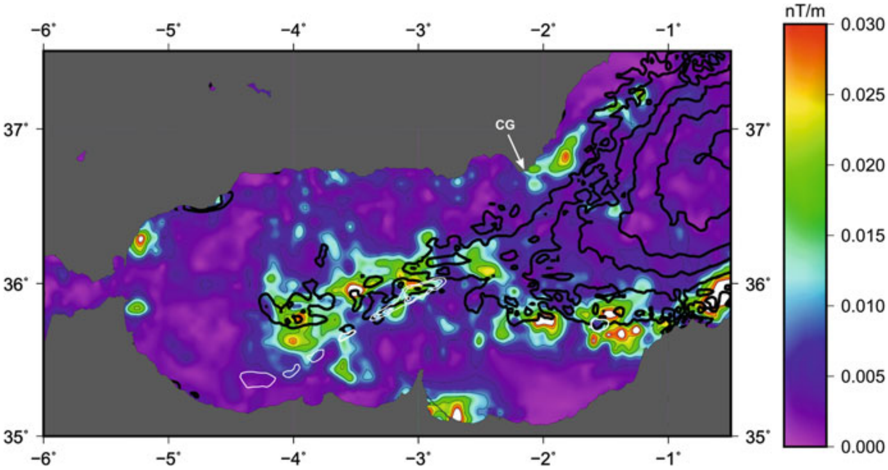

Fig. 5.9 Analytical signal map of the magnetic anomalies of the Alboran Basin. Total Bouguer gravity anomalies are displayed as thick black contour lines. Thick white lines delineate the Alboran Ridge topography. CG: Cabo de Gata

It coincides with the boundary of the complete Bouguer gravity anomaly, which dominates the middle and eastern Alboran Basin. This appoints to the AS maxima lineations are related to a magnetization contrast between the Argelian-Balearic Basin oceanic crust and the northern Iberia and southern African continental crusts (WNW-ESE magnetic anomaly).

In summary, gravity and magnetic anomalies mark three crustal domains: East 1° W denotes amplitude values corresponding to the very thin continental crustal Iberian and African domains, surrounding oceanic areas, or the Algero-Provenzal Basin. From 1° W to 4° W the continental crust has an ENE-WSW magmatic thinning axis. Finally, westward of 4° W, according to the AS map and gravity anomaly values, lies a continental domain. Such a cortical configuration is similar to that proposed by Gomez de la Peña et al. (2018) based on the reflective character of the crust studied by deep seismic reflection profiles; yet these authors discern a continental crust thinned to the north and west of the Alboran Ridge that would correspond to the Alborán Domain, plus a segment of northern Africa crust to the south and east of this morphological elevation.

5.4.2 *Basement Configuration and Major Structure*

The basement paleotopography and main structures are derived from seismic reflection profiles obtained from academia and industry (Bourgois et al. 1992; Comas et al. 1992; Maldonado et al. 1992; Watts et al. 1993; Docherty and Banda 1995; Chalouan et al. 1997; Vázquez 2001; Soto et al. 2010, 2012; Martínez-García

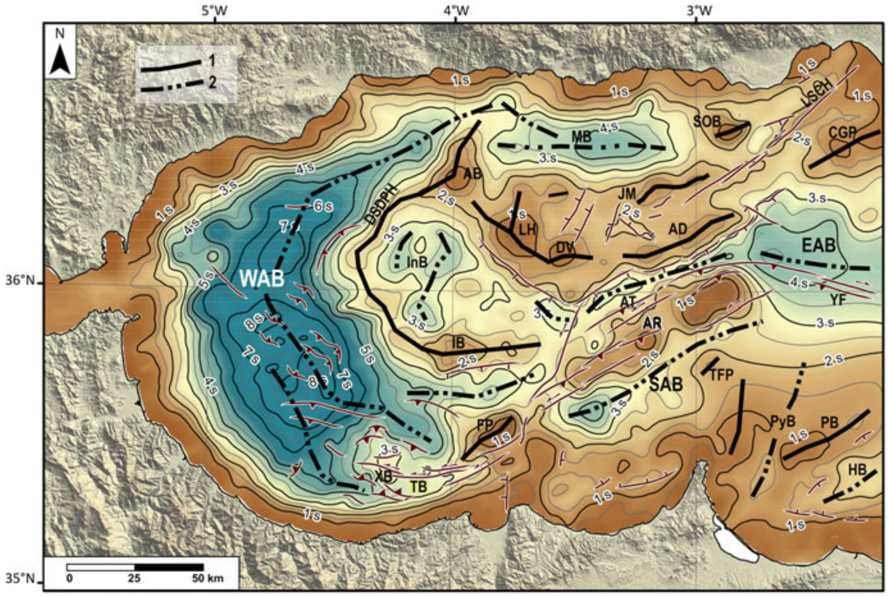

Fig. 5.10 Depth to basement map interpolated from the 2D seismic dataset (in seconds of two-way travel time). This horizon is affected by post-Tortonian deformation (Modified from Do Couto 2014). 1: Structural or volcanic highs axis; 2: Basin axis. AB: Avempace Bank; AD: Adra Ridge; AR: Alboran Ridge; AT: Alboran Central Trough; CGP: Cabo de Gata Promontory; DSDPH: Deep Sea Drilling Project High; DV: Djibouti Ville Bank; EAB: East Alboran Basin; FP: Francesc Pagés Bank; HB: Habibas Basin; IB: Ibn-Batouta Bank; InB: Intermediate Basin; JM: José Medialdea Bank; LH: La Herradura Bank; LSCH: La Serrata-Carboneras High; MB: Motril Basin; PB: Provençaux Bank; PyB: Pytheas Basin; SAB: South Alboran Basin; SOB: El Seco de los Olivos (Chella) Bank; TFP: Trois Fourches Promontory; WAB: Western Alboran Basin; XB: Xauen Bank; YF: Yusuf Fault

et al. 2013, 2017; Do Couto et al. 2016; Gómez de la Peña et al. 2018). Depth to basement map (Fig. 5.10) mainly represents the morphology of the metamorphic and volcanic basement beneath the basin infill.

The basement top surface is characterized by a structure of paleotopographic highs and depressions (Fig. 5.10). The two main sub-basins, WAB and EAB, stand out together with a secondary set of satellite basins found in the central segment (Motril, Southern Alboran, Central Alboran Trough, and Intermediate basins; Vázquez 2001) and the southeastern margin of the Alboran Sea (Habibas and Pytheas basins; Martínez-García et al. 2013) (Figs. 5.6 and 5.10). In addition, several structural highs are seen to have a different expression in the current seafloor morphology. They correspond to the highs of the Motril-Djibouti marginal plateau (Palomino et al. 2011), the Alboran Ridge, the DSDP high (whose superficial expression corresponds to Ibn-Batouta Bank), and the highs related to the Moulouya marginal plateau (Fig. 5.6; Chap. 6). The basement paleotopography would be partially distorted due to renewed uplift of the structural highs produced by the post-Tortonian compressive deformation, especially in the central sector of the Alboran Sea.

The main sedimentary basin formed on the strongly thinned crust of the Alboran Domain corresponds to the WAB, which is also the deepest depression and reaches 8 s twtt at Morocco offshore and 7 s twt at Spain offshore (Figs. 5.4 and 5.10). It is the thickest sedimentary depocenter in the Alboran Basin, estimated to lie at a depth between 10 and 12 km (Soto et al. 1996, 2010; Mauffret et al. 2007; Iribarren et al. 2009; Do Couto et al. 2016). Its geometry in plain view is arched, its axis mimicking the arcuate geometry of the orogenic arc, forming a curvilinear "bean" shape that continues toward the north to a secondary E-W basin. The northern branch has a NE-SW trend that changes to NNW-SSE southward, from latitude 36° N. This basin is largely affected by shale tectonics (Soto et al. 2010, 2012) and associated mud volcanism (Pérez-Belzuz et al. 1997; Sautkin et al. 2003; Talukder et al. 2003; Blinova et al. 2011; Somoza et al. 2012; Gennari et al. 2013).

The Motril Basin has an E-W direction, arranged approximately parallel to the northern coast of the Alboran Sea, with a slightly trapezoidal to elongated geometry, and its maximum thickness is greater than 4 s twtt (Fig. 5.10). The basin lies in the eastern extension of the northern branch of the WAB, just slightly disconnected from it (Vázquez 2001). The Intermediate Basin has no influence on the current physiography of the Alboran Sea. It is situated between the La Herradura-Djibouti Ville Banks and the DSDP highs (Fig. 5.10). Its geometry is oval, with a maximum axis of trend NNW-SSE. The maximum sedimentary thickness (>3.5 s twtt) is found in its northern part. This basin could be considered as an eastern secondary marginal sub-basin related to the main WAB. In turn, the southern sector of the Intermediate Basin would extend to the east along the Alboran Central Trough, of NE-SW orientation, that has a maximum thickness greater than 3 s twtt and a corridor geometry that is highly controlled by post-Tortonian compressive processes.

The continental crust described as belonging to the North Africa domain hosts three sedimentary basins. The longest is the SAB, located between the southern flank of the Alboran Ridge and the Trois Fourches Promontory (Vázquez 2001), having a NE-SW direction and a slightly trapezoidal to elongated geometry. The maximum thicknesses (>4 s twtt—Do Couto et al. 2016; locally ~4 km of sediments—Martínez-García et al. 2017) are located in its southwestern sector (Fig. 5.10). The formation of this basin and its current geometry are controlled by the post-Tortonian compressive deformation. It is separated from the WAB owing to progressive Alboran Ridge uplift since the upper Tortonian (Martínez-García et al. 2017) by folding and thrusting northward (Estrada et al. 2018).

Two sedimentary basins have been described to the east of the SAB: the Pytheas and Habibas basins. The Habibas basin (at least) is floored by continental metamorphic rocks (Medaouri et al. 2012, 2014). Both basins have an elongated geometry trending NE-SW to ENE-WSW and have an expression on the seafloor at their northern segment. The Habibas basin has a maximum thickness close to 4.5 s twtt (Gómez de la Peña et al. 2018) and is bounded to the north by the Yusuf Fault (Figs. 5.4 and 5.10), to the east by the Alidade Bank, and to the west by the E

Cabliers Ridge and the Provençaux Bank (Fig. 5.4), and progressively shoals southward; meanwhile the Pytheas Basin has a maximum thickness of 4 s twtt (Gómez de la Peña et al. 2018), is bounded to the east by the E Cabliers Bank, to the west by the W Cabliers Bank and the Trois Fourches Promontory, and to the south by the Provençaux Bank (Figs. 5.4 and 5.10).

Finally, above the thinned continental and magmatic crust lies the EAB. It has triangular geometry, its northern flank being controlled by the NE-SW La Serrata-Carboneras fault zone and constituted by the Adra Ridge and the Cabo de Gata Promontory; its southern flank corresponds to the NW-SE Yusuf fault zone and progressively opens eastward in transition with the oceanic lithosphere of the Algero-Provenzal Basin, so that its maximum thicknesses (>4.5 s twtt) would be in the western segment (Vázquez and Vegas 1996; Booth-Rea et al. 2007) (Figs. 5.4 and 5.10).

The structural highs—the NE-SW oriented Alboran Ridge, and its prolongation to the southwest in the Francesc Pagés and Tofiño banks—correspond to an antiform and a volcanic alignment (Figs. 5.4, 5.6 and 5.10), although the bulk of uplift was mainly produced by folding and inversion of its northern flank verging northwards, during the post-Tortonian compressive phase (d'Acremont et al. 2020). The banks of La Herradura and Djibouti Ville are probably located in relation to the NW-SE eastern faulted boundary of the Intermediate basin and correspond to two volcanic edifices formed in the Upper Miocene. However, their relief continues to the NE (in a NE-SW trend) in two branches, the northern one is constituted by the José Medialdea and Chella banks, and the southern one by the Adra Ridge, El Sabinar and Pollux banks and the Cabo de Gata Promontory (Figs. 5.6 and 5.10). The DSDP high displays a horseshoe geometry, its NE-SW directed northern branch taking in the Avempace bank and the smooth high that disconnects the WAB and Motril basins, while a central NNW-SSE branch separates the WAB of the Intermediate Basin and a southern E-W branch includes the Ibn-Batouta Bank (Fig. 5.10). Notwithstanding, some inversion and uplift of the northern and southern branches may have occurred during the post-Tortonian compressive phase.

Several structural highs have been defined to the east of the Alboran Ridge on the African margin. Two of them are the Trois Fourches Promontory and the Porvençaux Bank, of NE-SW trend (Fig. 5.10) and marking metamorphic basement highs (Gómez de la Peña et al. 2018). Another two are the NW and E Cabliers ridges (Figs. 5.4 and 5.6), respectively found in the northward prolongation of the aforementioned basement highs and described as volcanic edifices (Duggen et al. 2008). These structures were uplifted in the post-Tortonian compressive phase and produced tilting of pre-Messinian sedimentary units at least (Gómez de la Peña et al. 2018), a deformation apparently greater to the northeast, especially in relation to the intersection of NE Cabliers Bank and the Yusuf Fault (Figs. 5.4, 5.6 and 5.10).

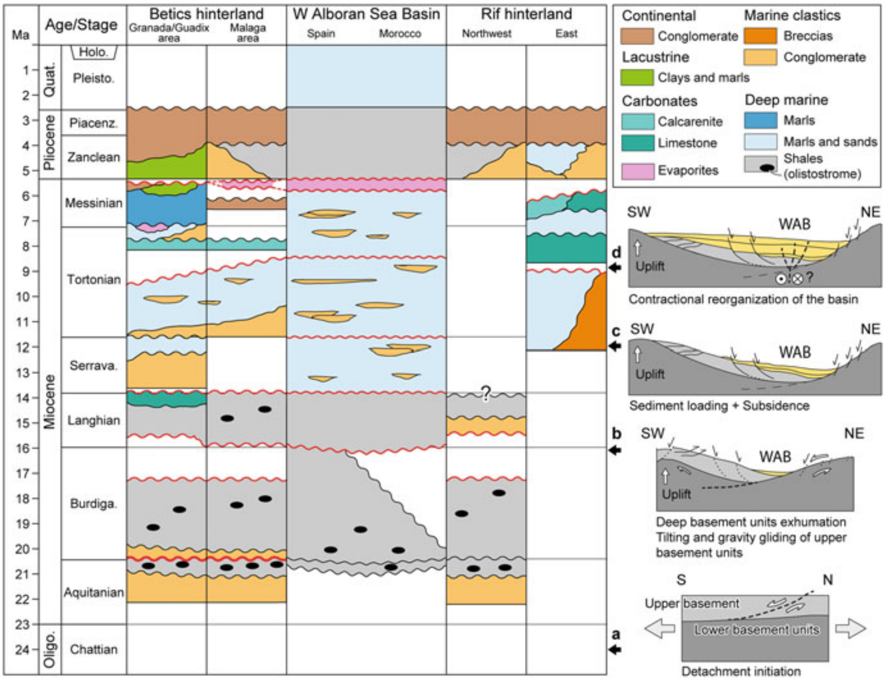

Fig. 5.11 Onshore-offshore chronostratigraphic correlation from Spain to Morocco, showing the main unconformities that affect the whole domain (see Do Couto et al. (2016) for more details), together with synthetic cross-sections reflecting the evolution of the western Alboran basin from late Oligocene (**a**) to late Tortonian, and the shift from extensional to contractional reorganization (**d**)

5.4.3 Tecto-Sedimentary Evolution

The different sub-basins share a common Pliocene-Quaternary history, although each of them formed and evolved differently over time (Fig. 5.11). The WAB contains the most complete stratigraphic record, from the lower Miocene (Aquitanian-Burdigalian) to Quaternary (Comas et al. 1999). Analyzing the geometries of seismic stratigraphic sequences sheds light on the evolution of the basin linked with crustal deformation.

5.5 Miocene

In the WAB, most tectonic and stratigraphic markers indicate that the basin began to develop in early Miocene times, probably above an extensional detachment that controlled subsidence (Do Couto et al. 2016). Then, in the middle Miocene, the

WAB underwent relatively homogeneous yet intense subsidence, interpreted as being controlled by the pull of the dipping subducting lithosphere and relative thermal cooling of the crust probably after the magmatic stages 1 and 2 (Torres-Roldán et al. 1986; Duggen et al. 2004), thereby explaining the considerable thickness (10 km) of the sedimentary infill (Do Couto et al. 2016). It is worth noting that deep structures in the basement imaged by means of deep seismic reflection profiles in the northern half of the Alboran Basin have been interpreted as extensional shear zones, associated with major extensional detachments in the Betic Cordillera (Watts et al. 1993; Vázquez et al. 1995; Vegas et al. 1995; Comas et al. 1997). The Motril Basin (northern E-W elongated part of the WAB) exhibits a rather different pattern, with an asymmetric geometry (Comas et al. 1999) comparable in shape to the asymmetric intramontane basins described inland (e.g., Sorbas Basin, Fig. 5.4), even though the total thickness of sedimentary rocks is far greater (<7 km). While the southern branch of the WAB is strongly affected by late-Miocene-to-present compression that produces its inversion, folding, and uplifting to constitute the Xauén and Francesc Pagés banks (d'Acremont et al. 2020; Lafosse et al. 2020).

The SAB and basins located to southeast differ from the WAB in that it lies above a crustal domain pertaining to Africa (Estrada et al. 2018; Gómez de la Peña et al. 2018). The sedimentary record appears to be younger than that of cores drilled in the WAB, starting in the middle Miocene (Serravallian) (Martínez-García et al. 2017). The EAB is located above thinned continental and magmatic crust (Booth-Rea et al. 2007) and the sedimentary record start also in the middle Miocene in its western part (Alvarez-Marrón 1999) but probably the initial age and the total thickness of the sedimentary column increases towards the east, in the transition with the oceanic crust of the Algero-Balearic basin (Comas et al. 1999; Booth-Rea et al. 2007). According to recent studies, whereas the late-Miocene-to-present compression reorganizes the morphology of the Alboran sub-basins, deeply buried Miocene normal faults demonstrate the widespread character of extensional tectonics around Alboran at that period (Martínez-García et al. 2017).

The *Lower Miocene* (Aquitanian-Burdigalian) is marked by the sedimentation of clays with interbedded sandy intervals above a basal pebbly sandstone in a syn-rift setting. Well data point to undercompacted shales, nowadays presenting overpressure in boreholes. Together, the seismic facies and borehole lithologies indicate that the lower Miocene consists of olistostromes containing polymictic rocks (olistoliths and rock breccia) embedded in an undercompacted shale matrix. The spatial distribution of these units is highly heterogeneous, the thickest depocentres occupying the center of the WAB (Do Couto et al. 2016), or Motril Basin (Comas et al. 1999), and so far nowhere else. The lower Miocene undercompacted shales are involved in the mud volcanic activity of the area (Sautkin et al. 2003; Gennari et al. 2013) and likewise in the shale tectonics that affect the WAB (Chalouan et al. 1997; Pérez-Belzuz et al. 1997; Talukder et al. 2003; Soto et al. 2010). It is difficult to estimate a paleobathymetry with respect to the sedimentary facies.

The undercompaction of shales in the lower Miocene series (Jurado and Comas 1992; Soto et al. 2010), the lack of late Oligocene sediments (Martínez del Olmo and Comas 2008), and the metamorphic paths of basement units are cited by authors

(Platt et al. 2013) who suggest rapid subsidence at the WAB initiation. The lower Miocene deposits are found overlying a basement composed of metamorphic units related to the Alpujarride-Sebtide Complex (Comas et al. 1999), or to the Malaguide-Ghomaride Complex (Do Couto et al. 2016). The origin of subsidence responsible for the deposition and fast burial of the sediments (causing undercompaction) can be found in the extreme thinning of the crust associated with the extensional detachments described onshore in the Beticas-Rif system.

The *Middle Miocene* (Langhian—Serravallian) witnesses the deposition of two sub-units: (1) a lower unit made of clays, interbedded silty to sandy clays, and basal coarse-grained sandstone, the unit displaying typical features of undercompaction with abnormal overpressures; and (2) an upper unit consisting of well-graded sand-silt-clay turbidites and turbiditic muds interbedded with volcanogenic layers. The lower Langhian unit, presenting signs of overpressure, was deposited in syn-rift settings as in the lower Miocene. However, at the end of the Burdigalian, the extensional setting of the Alboran Domain changes from regional N-S extension to widespread E-W extension. Such a transition triggered a moderate subsidence step in the WAB and adjacent basins. During the Serravallian, E-W extension proceeded westward, leading to the exhumation of E-W elongated metamorphic domes inland, thereby triggering the onset of subsidence in the southeastern Betics. This modification of the lithospheric configuration largely affected the WAB, which recorded the onset of a strong subsidence phase, signaled by an up to 3 km-thick Serravallian sequence (Do Couto et al. 2016).

The geometry of the Serravallian sequences, exhibiting overall landward to basinward onlap geometry, indicates that the WAB underwent relatively homogeneous yet intense subsidence ("sag basin") during part of its evolution. This period coincides with high volcanic activity in the Alboran Sea, recorded in the thick volcaniclastic series. Subsidence evaluation indicates two successive Langhian and Serravallian subsidence pulses (e.g., Comas et al. 1999) leading to the deposition of deep sea turbiditic fans. Elsewhere in the Alboran Sea, the Serravallian is also present in the SAB, marking the deepest deposits (Martínez-García et al. 2017). The sag basin geometry is indicative of subsidence along a relative thermal cooling period of the crust (White and McKenzie 1989). Besides, subsidence was accompanied by flexural basement tilting, both to the southeast in the norther margin and to the northwest in the southern one (Vázquez 2001; Soto et al. 2010; Do Couto et al. 2016). These processes favor a slope tectonics of overpressured shales caused a downslope migration, which also resulted in secondary basin subsidence (Soto et al. 2010).

The *Upper Miocene* (Tortonian-Messinian) is a key period in the WAB. It encompasses the transition from a regional tectonic extensional setting to a compressive regime. The lower Tortonian is marked by the deposition of graded sand-silt-clay turbidites interbedded with a few volcanogenic layers. The widespread spatial distribution and rather isopach character of this stratigraphic sequence indicate the end of the mostly extensional period. Its topmost surface is nowadays folded and partly eroded. Above, the upper Tortonian comprises sandstone intervals, with claystones and silty clay beds, also corresponding to turbidite facies with some channel-like features. This sequence was deposited under a mostly compressional regime, flexural subsidence remains, and generated secondary accommodation.

The Messinian features the superposition of two main sub-units in the Alboran Sea: (1) a lower unit made up of marine sandy turbidites interbedded with carbonates and volcaniclastic layers; and (2) an upper unit made up of gypsum and anhydrite embedded in clays, linked to the Messinian Salinity Crisis deposits. Indeed, the continuous compressional regime closed marine corridors linking the Mediterranean Sea to the Atlantic Ocean during the Messinian and led to the drying up (desiccation) of the Mediterranean Sea. While the Messinian deposits containing remnants of evaporites have never been mapped, borehole records tend to show they are relatively thin in the basin and seem to have eroded before the reflooding of the whole basin close to the Pliocene.

In general terms, post-Tortonian (~8 Ma) tectonics can be understood as a contractive reorganization affecting the Alboran Domain and modifying the architecture of Miocene basins and margins (Comas et al. 1992) as well as the local dimensions of sub-basins, favoring uplift of the Alboran Ridge and structural highs, while deforming by folding and faulting the sedimentary record (e.g., Estrada et al. 1997; Ammar et al. 2007; Vázquez 2001; Marín Lechado et al. 2007; Ballesteros et al. 2008; Martínez-García et al. 2011, 2013, 2017; Gómez de la Peña et al. 2018; d'Acremont et al. 2020).

5.6 The Opening of the Strait of Gibraltar

The Messinian Salinity Crisis (MSC) ended in the latest Miocene, when the isolated (or very restricted) the Mediterranean Sea became reconnected to world ocean currents through the Strait of Gibraltar. Initial models of the MSC establish that the Mediterranean Sea was progressively restricted from the Atlantic Ocean as the Gibraltar Arch shifted to the west and the African plate to the north, resulting in the closure of the Mediterranean-Atlantic gateways. However, a key question has remained unexplained: the thickness of salt accumulated in the deep basins would have required intermittent sea water inputs.

Over the past decades some authors have postulated that during the MSC, the Mediterranean was still connected to the Atlantic Ocean through the Strait of Gibraltar. Thus the deep desiccated basin model (Hsü et al. 1973) was replaced by one entailing stratified dense brines upon a deep non-desiccated basin under a normal sea level (Roveri et al. 2014). No direct proof supports a permanent Gibraltar gateway throughout the MSC, leaving its existence as an open question. Unclear as well is the contribution of the Paratethys Sea to the MSC salt precipitation.

The Alboran Basin proves key area to understand how the opening of the Strait of Gibraltar occurred. Detailed analysis of the Alboran Basin stratigraphy, by means of a dense net of seismic profiles correlated with commercial and scientific wells, indicates that the Strait of Gibraltar formed at the end of the MSC, in the transition to the Zanclean period, as a result of an unprecedented catastrophic event—the

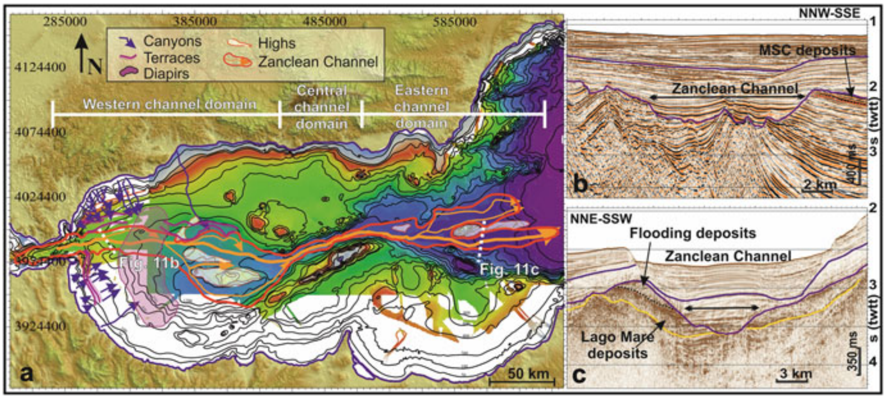

Fig. 5.12 (**a**) Present-day bathymetric map showing the location of the main morphological features related to the opening of the Strait of Gibraltar (modified from Estrada et al. 2011); (**b**) and (**c**) seismic profiles showing the erosive channel related with the Zanclean reflooding of the Mediterranean Sea. Legend: Yellow line corresponds to the time lapse of salt deposition during MSC paroxysm; purple, top Messinian; blue, the base of Quaternary. Horizontal scale in meters and vertical scale in seconds (twtt); see (**a**) for location

Zanclean Megaflood (Fig. 5.12) (Blanc 2002; Garcia-Castellanos et al. 2009, 2020; Estrada et al. 2011). The seismic stratigraphic record of this basin at the transition from Messinian to Pliocene indicates that MSC-related deposits have a characteristic thin, chaotic subunit (Fig. 5.12b) (Martínez del Olmo and Comas 2008) linked to the sea level drop of the MSC (Do Couto et al. 2016). Likewise, remnants of the MSC Lago Mare upper unit (UU) are identified in the easternmost basin (Fig. 5.12c), and patches of primary lower gypsum (PLG) are found in the nearby Malaga and Melilla areas. The lack of salt deposits is remarkable (Mobile Unit, MU). Overlying the MSC erosion and deposits, an open marine Pliocene-Quaternary unit develops, governed by the opening of the Strait of Gibraltar (Fig. 5.12b, c) (Juan et al. 2016).

The Messinian-Pliocene boundary is characterized by a regional polygenic unconformity (Estrada et al. 2011; Do Couto et al. 2016). Unlike other areas of the Mediterranean, where Messinian erosion mainly develops on margins, the Alboran Basin displays its strongest erosion in comparatively deep areas. Several erosional features (channels, terraces, and canyons) characterize this regional surface (Fig. 5.12a). The most prominent feature is an incised E-W channel crossing the entire basin (390 km long and up to 488 m deep) that may signal flooding in the wake of the Atlantic-Mediterranean re-connection (Garcia-Castellanos et al. 2009, 2020; Estrada et al. 2011) (Fig. 5.12). The incision depth of the channel is variable, suggesting local variations in the erosive capacity of Atlantic inflow, conditioned mainly by the regional basin topography and the local presence of topographic highs (Fig. 5.12a). Adjacent to this channel along the Spanish and Moroccan margins, and near the Strait of Gibraltar, several submarine terraces at different depths suggest a pulsed flooding of the Alboran Basin (Estrada et al. 2011).

The age of this striking channel is constrained by the eroded sediments and erosional surfaces. The channel clearly cuts the regional erosional surface as well as the chaotic sediments related to the MSC sea level drop, during coeval salt deposition in other areas of the Mediterranean Sea. In the easternmost Alboran Basin, the channel also incised the lattermost deposits of the MSC (UU), therefore indicating an age between the latest Messinian and Zanclean. As a result of the Zanclean Flooding, and postdating the MSC UU, several chaotic lenticular bodies developed atop the Zanclean channel margins, probably constituting a rare example of megaflood deposits coeval to those described for the Alboran Sea (Periáñez et al. 2019) and the nearby sill of the Strait of Sicily (Micallef et al. 2018).

What caused the opening of the Strait of Gibraltar? This is an unanswered question. Certain hypotheses evoke tectonic control, sea level changes, or retrogressive fluvial erosion, yet the most plausible cause is a combination of tectonics and retrogressive river erosion (Blanc 2002; Loget and Van Den Driessche 2006; Garcia-Castellanos and Villaseñor 2011).

5.7 Pliocene-Quaternary

The Pliocene-Quaternary deformation of the Alboran Sea is evidenced by the change from a mostly compressional regime (Campos et al. 1992; Maldonado et al. 1992; Woodside and Maldonado 1992; Rodriguez-Fernandez and Martin-Penela 1993; Estrada et al. 1997) and by a shift in the orientation of the convergence vector between the Eurasian and African plates, from NW-SE to WNW-ESE (Mazzoli and Helman 1994; Rosenbaum et al. 2002b; Merkouriev and DeMets 2008; Martínez-García et al. 2013). It contributed to the reactivation, propagation, and uplift of the NE-SW oriented structures (Estrada et al. 1997; Comas and Soto 1999; Martínez-García et al. 2013). To accommodate this convergence, indentation tectonics developed in the central Alboran Sea gave rise to two conjugated sets of dextral WNW-ESE and sinistral NE-SW to NNE-SSW faults, while in the same context ENE-WSW-oriented folds and antiforms developed (Estrada et al. 2018).

Three major shortening phases have been defined to characterize the Pliocene-Quaternary tectonic history of the Alboran Sea (Martínez-García et al. 2013): (a) earliest Pliocene (*ca*. 5.33–4.57 Ma), mainly deforming the Alboran Ridge; b) late Pliocene (*ca.* 3.28–2.59 Ma), uplifting the Alboran Ridge until closing the gateway that had connected the SAB and WAB; and c) Pleistocene (*ca.* 1.81–1.19) (e.g., Estrada et al. 1997; Martínez-García et al. 2013), entailing further deformation and uplift of the Alboran Ridge, deformation in the Yusuf Lineament (Martínez-García et al. 2013; Estrada et al. 2018), and northward propagation of the Al-Idrissi Fault zone (d'Acremont et al. 2014; Estrada et al. 2018).

The Alboran Basin narrowed considerably during the Pliocene due to the emergence of onshore sub-basins and the progressive uplifting of the Alboran Ridge as a result of the first two major phases of shortening (Martínez-García et al. 2013). Basement flexure along Pliocene caused a general deepening of sedimentary units (particularly in the WAB) (Docherty and Banda 1992; Rodríguez-Fernández et al. 1999) and tectonics altered the sizes of the easterly sub-basins (Estrada et al. 1997; Comas et al. 1999; Martínez-García et al. 2013). The uplift of the Alboran Ridge propagated from NE to SW, splitting the EAB and WAB (Martínez-García et al. 2013). The Xauen and Tofiño banks area was extensively folded and uplifted (Bourgois et al. 1992; Chalouan et al. 1997; d'Acremont et al. 2020; Lafosse et al. 2020) and, together with the Alboran Ridge, progressively led to the formation of a long morphologic barrier in the late Pliocene that is presently ~130 km long and 1.75 km high (Ammar et al. 2007; Martínez-García et al. 2013; Vázquez et al. 2015). The re-orientation of previous tectonic structures due to the changes in the stress field during the Pliocene and Quaternary, along with the uplifting of diapirs in the WAB (Pérez-Belzuz et al. 1997; Talukder et al. 2003), also conditioned smaller morphotectonic changes behind other evolving tectonic features, such as basement highs, structural scarps, mud volcanoes, and diapiric ridges.

The morphotectonically active setting that characterized the Alboran Sea during the Pliocene and Quaternary involved a changing seafloor landscape and basin configurations. They largely controlled the distribution and lateral continuity of the Pliocene and Quaternary seismic stratigraphic units. The most common stratigraphic boundaries of the overlying Pliocene and Quaternary stratigraphy were described in Ryan et al. (1973) and Campillo et al. (1992), primarily along the Spanish margin. A recent analysis identified the main tectonic/sedimentary reflectors in the SAB and EAB (Martínez-García et al. 2013). All these sources were recently combined and updated by Juan et al. (2016).

The Pliocene-Quaternary sequence of the Alboran Sea has been previously addressed by means of industrial and academic wells (Jurado and Comas 1992; Comas et al. 1999) and is known to consist of pelagic to hemipelagic marls and shales (Fig. 5.11), interlayered with sandy turbidites and/or contourites (Alonso et al. 1999; Juan et al. 2020). Its base is marked by a discontinuity recognized across the entire basin, linked to erosion and the Messinian Salinity Crisis (Estrada et al. 2011). The overall geometric configuration of the Pliocene sequence and its units consists of irregular subtabular sedimentary bodies and roughly wedge-shaped bodies pinching out upslope and, locally, downslope (Juan et al. 2016). The Quaternary sequence and its units primarily display wedge shapes along the Spanish and Moroccan margins (Ercilla et al. 1994; Hernández-Molina et al. 2002), subtabular shapes in the basin domains, and mounded shapes at the bases and walls of the seamounts and escarpments, on the Alboran Ridge, and in the Alboran Trough (Juan et al. 2016).

Throughout the Quaternary, sea level changes related to climate forcing constrained the edification of continental shelf physiography, especially since the middle Pleistocene, as a consequence of the greatest importance of the 100 k year cycles (Hernández-Molina et al. 2002). In low sea level events, the continental shelf is sculpted and grows seaward through the deposit of shelf-margin wedges tied to continental sediment input from river mouths; in turn, high sea level events saw the migration of the coast towards the continent and the deposit of coastal deltas and wedges (Ercilla et al. 1992, 1994; Hernández-Molina et al. 1994, 1996; Chiocci et al. 1997; Chap. 6).

Although early studies mostly carried out along the Spanish margin concluded that confined and unconfined downslope processes and hemipelagic settling have played a dominant role in outbuilding the continental margins and infilling basins, bottom currents recently came under the spotlight (Ercilla et al. 2002, 2016; Palomino et al. 2011; Juan et al. 2016, 2020). The greater depth of Pliocene basins favored the action of a strong, dense countercurrent in the WAB (as modeled by Alhammoud et al. (2010), in a shallow sill situation), which helps explain the thick Pliocene deposits as well as the broad sheeted and plastered drifts. The uplifting of the Alboran Ridge would have weakened this dense countercurrent (Juan et al. 2016); hence, a circulation model conditioned by the complex topography could be established. The Quaternary topography gave rise to a separation of light versus dense water flows into small-scale branches, forming minor contourite features within narrow passages and at the base of the seamounts (Juan et al. 2016). The changes in the contourite systems have allowed to interpretate an increase in the flow energy from the Pliocene to the Quaternary in this area (Juan et al. 2020). Furthermore, the scarcity of turbidite systems in the Morocco margin has been explained by the action of water masses that produces dispersion of the sediments (Ercilla et al. 2019), and the reduction in the number of turbidite systems since the Pliocene and decrease in their sizes could be also related to the enhanced of water mass action on sediment transport (Juan et al. 2020).

5.8 Conclusions: Paleogeographic Implications

The evolution of the area between Africa and Iberia, the region where the Alboran Sea is currently located, has successively been constrained since the Jurassic by the movement of the African and Eurasian plates, by the intermediate role of the Iberian plate, showing a certain independence from the main plates during a period of its development, and by orogenic processes since de Late Cretaceous first in the Ligurian Tethys oceanic basin and later in the Western Mediterranean oceanic basin. Related to the last process, a lithospheric terrane now known as the Alboran Domain collided with the continental margins of South Iberia and North Africa since the Upper Oligocene. The westward drift of this domain during the Miocene generated an extensional basin inside, characterized by a very stretched continental lithosphere and significant magmatic activity in its central and eastern sectors, that evolved to the current basin of the Alboran Sea after compression from the Late Tortonian.

One crucial element of the geological evolution between Africa and Iberia is their geographical location: at the point of connection of two oceanic basins, the geodynamic plates evolution was responsible for the creation and destruction of several ocean basins and their oceanographic gateways. Thus, tectonics controlled

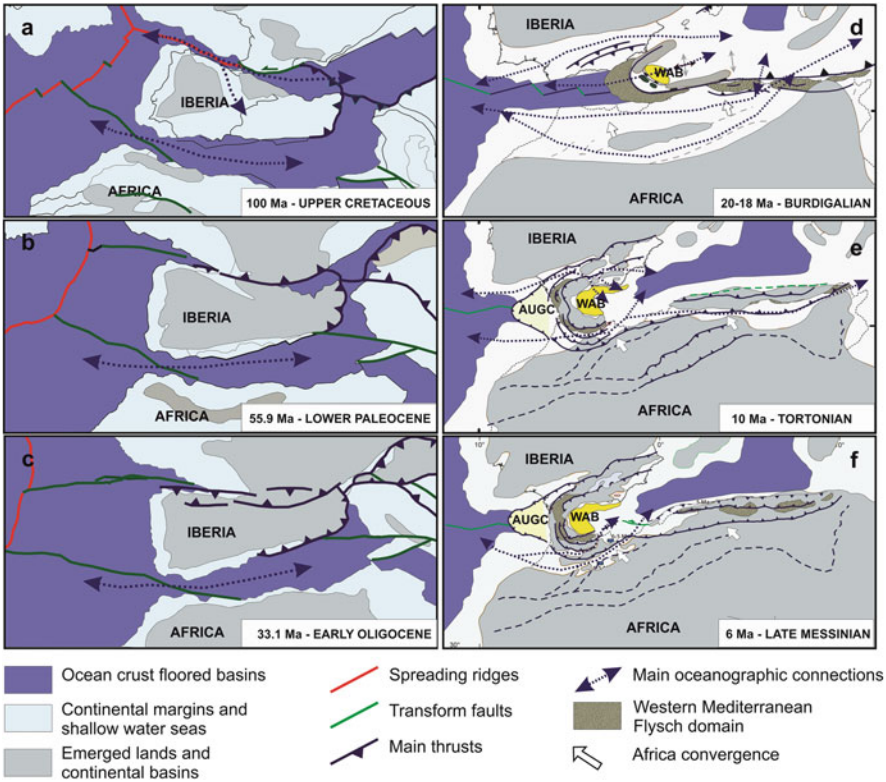

Fig. 5.13 Evolutive scheme of palaeoceanographic relations between the Atlantic Ocean and the different oceans and seas in the Tethys domain, east to the Iberia, (**a**, **b**, **c**) from the Upper Cretaceous to Eocene-Oligocene boundary based on Ziegler (1988), Schettino and Turco (2011), Advokaat et al. (2014) and Martín-Chivelet et al. (2019); (**d**, **e**, **f**) Palaeoceanographic scheme of the Alboran Sea region along the formation and evolution of the Betic-Rif Belt and the Alboran Sea Basin (modified from Do Couto 2014; Do Couto et al. 2016)

the oceanographic connections between the seas situated to the west and the east of Iberia.

Within the specific region currently occupied by the Alboran Basin (yet previous to its formation), there was an oceanographic connection between the Atlantic and Ligurian Tethys oceans in southern Iberia in the Late Jurassic, and a secondary connection occurred in northern Iberia (Ziegler 1988; Schettino and Turco 2011). Since the Late Cretaceous (Fig. 5.13a) the Iberian block has undergone compression due to the convergence between Africa and Eurasia; indeed, it led to the closure of the connection in northern Iberia during the Paleocene (Ziegler 1988; Rosenbaum et al. 2002a) (Fig. 5.13b) and finally to the Pyrenean orogen as part of the Alpine belt in the Eocene phase (Schettino and Turco 2011). Some uncertainty surrounds the characteristics of the southern oceanographic connection. Did it occur only across the continental margins of South Iberia and North Africa? Elsewise, if oceanic seafloor spreading was involved, could it be associated with the stretching of the continental margins located at the western end of the Ligurian Tethys oceanic basin? (Vergés and Fernàndez 2012; Vergés et al. 2019).

The closure of the Ligurian Tethys basin is clearly linked to the subduction of the African oceanic lithosphere northward, under Eurasia, in the Eocene (Fig. 5.13c). It would have favored the progressive disappearance of the Atlantic-Tethys oceanographic connection. Still, the change in subduction described for the Oligocene immediately led to the extensional collapse of the Eocene alpine orogen and opening of the Western Mediterranean oceanic basin. A further consequence was the splitting—to the east, south, and west—of several lithospheric fragments of this orogen (Faccenna et al. 2004; Jolivet et al. 2009; Carminati et al. 2012). This new geodynamic scenario welcomed the maintenance of the oceanographic connection between the Atlantic Ocean and the Western Mediterranean Sea as the Alboran Basin was being formed, and its additional connection to the Ligurian Tethys while this basin was gradually closed due to subduction of its oceanic lithosphere.

Westward drift of the Alboran Domain throughout the Miocene resulted in the formation of the Betic-Rif Belt when it collided with the continental margins of southern Iberia and northern Africa. Inside this orogen, in the back-arc region, a marine basin developed in the wake of extensional stretching of the lithosphere (Comas et al. 1999; Comas and Soto 1999). At this stage, several W-E oceanographic connections resisted, by means of the Northbetic area, the Alboran Sea, and the Rif area (Braga et al. 2001, 2003; Martín et al. 2009; Do Couto et al. 2016; Capella et al. 2017) (Fig. 5.13d). Thus, the Tortonian marine basin would have had a much greater width (even several hundred miles wider) than the current basin (Fig. 5.13e). This is attested to by the distribution of Tortonian coral reefs along the main mountainous reliefs of both the Betic and Rif chains, reliefs that existed as large islands within the Tortonian Sea (Braga et al. 2003; Galindo-Zaldívar et al. 2019).

A shift in the direction of convergence of Africa and Eurasia in the Late Tortonian slowed the Alboran Domain's westward drift. At the same time, tectonic stacking at the Gibraltar Arc front (Balanyá et al. 2007), deformation related to block rotations (Crespo-Blanc et al. 2016), and the progressive uplift of the Betic and Rif ranges onland (Braga et al. 2003), because of increased thickness owing to the emplacement of the Alboran Domain. The Upper Miocene also witnessed the main stage of magmatic activity (Duggen et al. 2008): several volcanos formed in the Alboran Basin, very likely associated with a thermal expansion of the seafloor. Furthermore, it has been proposed that the joint action of mantle resisted slab dragging and slab tearing also influenced the closure of these gateways (Capella et al. 2020). Given this compressional setting, together with the generalized sea level fall in the Messinian (Jolivet et al. 2006; Loget and Van Den Driessche 2006), an imminent consequence was the closure of marine water corridors through which the Mediterranean Sea communicated with the global ocean (e.g., Martin et al. 2001; Betzler et al. 2006; Gibert et al. 2013; Flecker et al. 2015; Achalhi et al. 2016; Capella et al. 2018; Krijgsman et al. 2018) (Fig. 5.13f). The Mediterranean Sea dried out nearly 5.9–5.5 Ma ago. A thick series of evaporites was deposited in the deepest basins, while the margins, including the Alboran Domain, underwent subaerial erosion (Estrada et al. 2011; Gorini et al. 2015). This crisis came to an end after a major breakdown of the Gibraltar dam, which caused quick, intense reflooding into the Mediterranean Sea (Garcia-Castellanos et al. 2009, 2020).

From this point onward, now in the early Pliocene (Zanclean), a new gateway opened between the Atlantic Ocean and the Mediterranean Sea. On the basis of tectonic, gravitational, and erosive effects studies near the Gibraltar Arc, the opening can be situated in close correspondence to the current Strait of Gibraltar (Loget and Van Den Driessche 2006; Estrada et al. 2011). After an initial erosive phase owing to Atlantic water mass flooding, the currently existing oceanographic dynamics progressively prevailed; hence, the region's present state can be traced to the late Pliocene-Quaternary boundary (Juan et al. 2016). Yet coetaneously, throughout this phase, the bulk of uplift affecting main, positive reliefs of the basin took place. In particular, the Alboran Ridge and its continuation toward the SW toward the bank of Xauen were substantially elevated (Martínez-García et al. 2013; Lafosse et al. 2018). This elevation is therefore a consequence of compressive processes that acted during the Pliocene-Quaternary, along with the indentation of a very resistant cortical block associated with the Alboran Ridge, all in conjunction with the compressive ridge uplift and the development of WNW-ESE right-lateral and NNE-SSW left-lateral conjugate fault systems, bounding it respectively to the east and to the west, and propagating deformation toward the northern margin of the basin (Estrada et al. 2018).

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