E–W opening of the Algerian Basin (Western Mediterranean)

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ABSTRACT

We present geophysical and geological evidence of fast (2 cm yr\(^{-1}\) in the east and 5 cm yr\(^{-1}\) in the west) E–W spreading between 16 and 8 Ma in the region behind the Gibraltar Arc in the Algerian Basin.

Introduction

There is general consensus about the formation of the Western Mediterranean Basin behind a Miocene volcanic arc. Clockwise rotation of the Corsica–Sardinia Block (Vigliotti and Kent, 1990) and the subsequent opening of the Provencal Basin is corroborated by geophysical and geological evidence (Rehault et al., 1984). The Algerian Basin (Fig. 1) is often presented as an extension of the Provencal Basin that has undergone a similar history but with drift of the internal domain of the Maghrebian system (Kabylies) towards the south-east. However, convergence between Africa and Europe is slow but has induced lateral arcs (Gibraltar and Calabria–Apennine arcs) to migrate quickly towards the west and the east, respectively. The rapid lateral tectonic escape of the arcs and coeval rollback of the subduction zones must have had considerable consequence for the structure of the back-arc basins. Although the Alboran Basin and the adjacent Betic-Rif domain have been extensively studied, the Algerian Basin is as yet poorly known, although several commercial and academic seismic surveys have been performed in this area. We present a new picture (Fig. 2) of the Algerian Basin based on a synthesis of these data.

Tectonic setting

The Eastern Alboran Basin is located between the Eastern Alboran Basin to the west, the Sardinia and Tunisia margins to the east, the Balearic margin to the north and the Algerian margin to the south (Fig. 1). The boundary between the East Alboran Basin and the Algerian Basin is marked by the western limit of diapirism and a N–S magnetic anomaly (Galdeano and Rossignol, 1977), whereas a prominent magnetic anomaly (Galdeano and Rossignol, 1977), here termed Hamilcar, separates the Provencal and Algerian Basins. The eastern boundary of the Algerian Basin is the North Tunisian Fracture Zone (Auènde et al., 1974). The Spanish and the Balearic margins are delimited, respectively, by the very steep Mazarron and Emile Baudot escarpments. This last escarpment, now inactive, shows clear evidence of right-lateral strike-slip motion in the past (Mauffret et al., 1992; Acosta et al., 2001). The steep slope of the Algerian Margin is formed by a complex set of strike-slip and reverse faults (Mauffret, 2003) that are related to the present right-lateral transpression along the Algerian margin.

Basement of the Algerian Basin

In the Eastern Alboran Basin Miocene salt deposits are thin or absent and the Moho very shallow (7 s TWTT). The crust thins from 6 km in the Eastern Alboran Basin to 4 km in the Algerian Basin beneath the salt domes (Comas et al., 1997). The thickness of the crust in the East Alboran Basin is confirmed by high heat flow values (124 mW m\(^{-2}\); Polyak et al., 1996). Therefore, the Eastern Alboran Basin represents the transition zone between the continental crust of the Alboran Basin and the oceanic crust of the Algerian Basin. On the eastern side of the Algerian Basin, south of Sardinia, a CROP (CROsta Profonda Project) profile (Catalano et al., 2000) reveals a 5.8-km-thick crust that thickens toward the east. ESCI (Estudios Sismicos de la Corteza Iberica) seismic profiles that crosses the Valencia Trough, the Balearic Promontory and the northern Algeria Basin (Vidal et al., 1998) and a refraction profile (Hinz, 1972) show a thin crust (4.5 km, Fig. 1A) that is similar in thickness (5.5 km) to the oceanic crust of the Provencal Basin (Pascal et al., 1993). However, the sedimentary cover here is very different. The Quaternary–Pliocene sedimentary pile is 2 km thick in the Provencal Basin but only 0.8 km thick in the Algerian Basin. Moreover, the same contrast is observed for the pre-Messinian Miocene layer, which is up to 4 km thick in the Provencal Basin but only 1.8 km thick in the Algerian Basin. The oldest sedimentary layer dated on the Algerian margin (Alger 1; Burrollet et al., 1978) is late Langhian (14.7 Ma). The Neogene deposits following the main compressional tectonics are of late Burdigalian age (15.8 Ma) in the Grande Kabylie (Magne and Raymond, 1974), whereas a coeval basaltic magmatism (16–15 Ma) occurred (Maury et al., 2000) in the Dells region (Fig. 3B). A heat flow value of 124 mW m\(^{-2}\) (Polyak et al., 1996) may correspond to a 16-Ma crust. We therefore propose that the oldest oceanic crust of the Algerian Basin is 16 Ma, whereas the youngest
Fig. 1 Tectonic sketch map of the Algerian Basin. Extension of the Messinian diapirism is shown. Note the absence of salt diapirism in the East Alboran Basin and Sardinia margin where a thin layer of continental crust has been identified. The Messinian salt is also thin on the Hannibal Ridge between Menorca and the Algerian margin where we suggest that a spreading centre was active until the late Tortonian (8 Ma) and perhaps later (6 Ma). The N–S trend of several magnetic anomalies (Galdeano and Rossignol, 1977) support this hypothesis. Inset A: comparison between the Provencal and the Algerian basins. Note the difference in depth of the Moho (15 and 10 km) related to the difference in thickness of the Pliocene–Quaternary and Miocene sedimentary cover. We propose that the Algerian Basin is younger (16–8 Ma) than the Provencal Basin (21–16 Ma). The Sardinia and the Drepano Thrust Front are drawn from Catalano et al. (2000).

Fig. 2 (A) Depth cross-section of the Hannibal Ridge. Note the thinning and deformation of the Messinian layers. (B) Seismic section crossing DSDP Site 371, which bottomed in Messinian evaporites (Hsu et al., 1978). The salt layer is absent on the high; beneath a thin evaporitic cover, dipping reflectors suggest that the DSDP 371 high to the east, correlated with a large magnetic anomaly, is volcanic. This volcanic activity is pre-Messinian, although several steps in the Messinian sediments and the basement suggest a persistence of activity during the Messinian. (C) Depth section without vertical exaggeration. (D) Enlargement of the seismic line J 103 showing dipping reflectors (white arrows) just beneath the Upper Messinian evaporites and above a horizontal layer. (E) Block diagram of the Hannibal Ridge. (F) Depth to basement map showing the NNE–SSW trend of several highs of the Hannibal Ridge.
crust of the Provencal Basin has the same age (16 Ma) as suggested by the end of rotation of Corsica–Sardinia (Speranza et al., 2002).

The Hannibal Ridge is a N–S-trending basement high (Figs 1 and 2). Several prominent magnetic anomalies are observed in the Algerian Basin (Galdeano and Rossignol, 1977). One of these anomalies is correlated to a high where the DSDP Site 371 has been drilled (Hsü et al., 1978). Just beneath the Messinian some dipping reflectors (Fig. 2, insets B and D) suggest that the volcanic feature was active during the late Tortonian although some Messinian volcanic activity and normal faulting may be responsible for the disturbance of the Messinian evaporitic layer. The depth to basement shows (inset F, Fig. 2) that the highs trend NNE–SSW in an en echelon pattern. A heat flow value of 132 mW m$^{-2}$ (Fig. 4) without sedimentary correction (Erickson, 1978) suggests an apparent age of 13 Ma. On the southern continental extension of Menorca, a value of 81 mW m$^{-2}$ has been recorded (Scientific Party, 1996) at ODP Site 975 (Fig. 4). On a tomographic map at 50 km depth a low-velocity anomaly (Piromallo and Morelli, 2003) correlates with a high temperature regime.
exceeding 1400 °C at 50 km depth (Yegorova et al., 1997). Therefore, we suggest that the Hannibal Ridge is a young (Late Tortonian, 8 Ma) and hot structure that was a high relative to the Messinian salt diapiric province (Fig. 2, inset A). However, some volcanic activity and normal faulting may have persisted during the Messinian. In summary, we suggest that a 560-km-long strip of oceanic crust was generated between 16 Ma and 8 Ma.

In the following sketches of evolution in the Mediterranean region we use largely the reconstructed convergent motions between Africa and Europe presented by Gueguen et al. (1998) and Rosenbaum et al. (2002). The tectonic sketch is superimposed on a tomographic map at 50 km depth (Piromallo and Morelli, 2003). Note the low velocity (in red), probably corresponding to a hot area beneath the Hannibal Ridge. Note the high heat flow in the Eastern Alboran Basin (124 mW m⁻², Polyak et al., 1996) and the Algerian Basin (132 mW m⁻², Erickson, 1978). A distance of 160 km separates the Hannibal Ridge from the Hamilcar magnetic anomaly, which may correspond to the boundary between the Provencal Basin and the Algerian Basin, whereas the oceanic area between the Hannibal Ridge and the Eastern Alboran Basin is 400 km wide. A present-day zone of intermediate seismicity between 50 and 150 km depth crosses the Western Alboran Basin (Bufrom et al., 1997; Lopez Casado et al., 2001), suggesting an almost vertical lithospheric slab. However, an apparent gap of seismicity between 0 and 50 km suggests a slab breakoff. The late Miocene (12–6 Ma) calc-alkaline to transitional volcanic rocks in the Eastern Alboran Basin (Maury et al., 2000; Duggen et al., 2004) may be related to a volcanic arc that trends N–S. In Terra Nova, Vol 16, No. 5, 257–264

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**Fig. 4** Sketch of the Mediterranean Sea during the Tortonian (8 Ma) based on the reconstructions of Gueguen et al. (1998) and Rosenbaum et al. (2002). The tectonic sketch is superimposed on a tomographic map at 50 km depth (Piromallo and Morelli, 2003). Note the low velocity (in red), probably corresponding to a hot area beneath the Hannibal Ridge. Note the high heat flow in the Eastern Alboran Basin (124 mW m⁻², Polyak et al., 1996) and the Algerian Basin (132 mW m⁻², Erickson, 1978). A distance of 160 km separates the Hannibal Ridge from the Hamilcar magnetic anomaly, which may correspond to the boundary between the Provencal Basin and the Algerian Basin, whereas the oceanic area between the Hannibal Ridge and the Eastern Alboran Basin is 400 km wide. A present-day zone of intermediate seismicity between 50 and 150 km depth crosses the Western Alboran Basin (Bufrom et al., 1997; Lopez Casado et al., 2001), suggesting an almost vertical lithospheric slab. However, an apparent gap of seismicity between 0 and 50 km suggests a slab breakoff. The late Miocene (12–6 Ma) calc-alkaline to transitional volcanic rocks in the Eastern Alboran Basin (Maury et al., 2000; Duggen et al., 2004) may be related to a volcanic arc that trends N–S.

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**The Aquitanian stage (23 Ma; Fig. 3A)**

In the western Mediterranean, convergence between the African and European Plates began during the Late Cretaceous (Olivet, 1996) and up to Eocene, convergence mainly being accommodated in the Pyrenees (Roest and Srivastava, 1991). However, a subduction zone was probably also active along the southern margin of the Iberian Plate (Jolivet et al., 2003). The Numidian (Kabyliennes, Sicily) or Algibe (Betics) flyschs were deposited on the top of sedimentary sequences (the Flysch Trough units, Fig. 3A) overlying Tethys oceanic crust, which probably formed an accretionary prism along the southern boundary of the European Plate (Johansson et al., 1998; Stromberg and Bluck, 1998) during the Oligocene – Early Miocene (Aquitanian–Burdigalian). After an Eocene climax of compressional deformation in the Pyrenees, the Alps and the eastern region, extension started in the western Mediterranean realm first in the Gulf of Lion grabens (Oligocene–Aquitanian) then shortly afterward in the Valencia Trough (Aquitanian–Burdigalian). However, the most important grabens are in a fore-arc position between the volcanic arc and the fragments of the European margin (the so-called AlKaPeCa domain; Alboran–Kabylie–Peloritan–Calabria; Bouillin, 1986) particularly along the eastern coast of Corsica and Sardinia (Mauffret et al., 1999; Sartori et al., 2001). A third basin is the West Alboran Basin. This basin and some associated grabens are filled by early Burdigalian syn-rift series (Comas et al., 1992, 1999; Alonso-Chaves and Rodriguez-Vidal, 1998). Extension was orientated N–S in these basins, which were formed east of their present situation (Fig. 1) and then transported westwards (Garcia-Duenas et al., 1992; Martinez-Martinez and Azanon, 1997). In Grande Kabylie (Saadallah and Caby, 1996), in Sicily (Kezirian et al., 1994) and in the La Galite Block (Bouillin et al., 1998) extension expressed by both ductile and brittle normal faulting is dated at about 22–25 Ma. The main basins trend E–W to NE–SW. Some grabens, probably related to a period of transtension, show a NW–SE direction (DSDP Site 372, Sicily). The tight fit is constrained by the distance (200 km; England et al., 2004) between the volcanic arc and the trench.

**Late Burdigalian Stage (16 Ma; Fig. 3B)**

The Provencal Basin opened as a back-arc basin behind the Corsica–
Sardinia Block between 19 Ma and 16 Ma (Speranza et al., 2002). The Sardinia Block migrated along the right-lateral North Balearic Fault Zone and displaced the Balearic Promontory with the coeval rifting in the Valencia Trough (Maillard and Mauffret, 1999). However, Menorca has been involved in the drift of Sardinia more so than Mallorca. Therefore, the Hamilcar magnetic anomaly (Figs 1 and 3B) that marks the scar of the Sardinia drift is aligned parallel with the DSDP Site 122 fault zone that separates Menorca and Mallorca (Maillard and Mauffret, 1999). We have shown that an important period of rifting affected the whole Western Mediterranean Basin but that this rifting in the Valencia Trough and south of the Balearic Promontory were probably sufficient to move the Magrebibian internal blocks (Kabybies) up to collide with Africa without formation of oceanic crust behind the Kabybies. Figure 3(B) demonstrates the widening of the former AlKaPeCa domain. After collision of the Kabybies and Africa during the early Burdigalian, a second extensional episode occurred at 17–15 Ma and this rifting is expressed first in a ductile regime with the exhumation of the deep part of the collisional prism then in a brittle regime that preceded the opening of the Algerian Basin. The Serravallian faults (15.8–11 Ma) in the Betics show a N-S orientation (Garcia-Duenas et al., 1992; Mayoral et al., 1994). The E-W early Miocene trend and the N-S middle Miocene orientation of the faults are characteristic of the chocolate tablet shape of the extensional tectonics in the Betics (Martinez-Martinez and Azanon, 1997). In the Grande Kabybies, a period of brittle extension is recorded and coeval with 15–16 Ma basalts and granodiorites of Delys (Aite and Gérald, 1997). In the Petite Kabybie, a period of ductile extension with a top-to-the-NW sense of shear (Caby et al., 2001) is dated at 16–17 Ma (late Burdigalian; Monié et al., 1992). Therefore, the E-W extension is mainly dated at 16 Ma and preceded the westwards rollback of the Gibraltar Arc (Martinez-Martinez and Azanon, 1997) and formation of oceanic crust behind the arc in the Algerian Basin. In our reconstruction, we have placed the Eastern Alboran Basin and the Western Sardinia margin in a conjugate position. Opening along the Hamilcar magnetic anomaly is transferred to the North Tyrhenian Sea via a fracture zone (Rehault et al., 1984) that probably utilized the previous collision zone between the Sardinia Block and La Galite Block (Tricart et al., 1994; Catalano et al., 2000). The positions of the Eastern Alboran Basin and the Rif during the late Burdigalian are 560 km distant from their present locations. It is likely that the primitive Alboran Sea before the Late Tortonian was wider than it is today (Comas et al., 1992, 1999) and that overthrusting of the internal massifs on the adjacent continents must also be taken into account. In previous estimates, 300 km (Martinez-Martinez and Azanon, 1997) to 250 km (Platt et al., 2003) of displacement are proposed. The rollback of the Gibraltar Arc implies an area of Atlantic oceanic crust where the Alboran Sea is presently located. Motion on the Alboran block is similar to that of the Caribbean Plate (Mann et al., 1995) or Scotia–Sandwich Plates, with left-lateral motion on the Algerian side and right-lateral motion on the Balearic side. The Sierra Norte in Mallorca (Fig. 3B), uplifted during the Langhian (Vergès and Sabat, 1999) could be an expression of this right-lateral transpression, but the most spectacular right-lateral strike-slip fault is the Emile Baudot Escarpment (Mauffret et al., 1992; Acosta et al., 2001) with its horse-tail structure, in its southwest extremity, that characterizes a period of dextral motion.

Late Tortonian Stage (8 Ma; Fig. 4) We have presented the probable relationship between heat flow (Erickson, 1978), the low velocity illustrated by the tomographic map at 50 km depth (Piromallo and Morelli, 2003) and the Hannibal Ridge that consider to be a spreading centre that ended its activity at 8 Ma. However, some volcanic and faulting activity persisted locally, for example the seamount revealed at DSDP 371 (Fig. 2C,F). We consider that westwards motion ended during the late Tortonian because a period of compression orientated NW–SSE is documented in the Betics (Comas et al., 1992, 1999). However, subduction may have persisted at the westernmost end of the system beneath the present-day Gibraltar Arc (Gutscher et al., 2002), although a slab detachment (between the surface and 50 km depth) beneath the Western Alboran Sea is also possible because seismicity at intermediate depth, probably related to an almost vertical slab, is active only between 50 and 150 km (Lopez Casado et al., 2001). The general N-S trend of the 12–6 Ma calc-alkaline to transitional volcanic rocks (Maury et al., 2000; Duggen et al., 2004) supports the presence of a late Miocene volcanic arc crossing the East Alboran Basin (Fig. 4).

From our map (Fig. 4), it appears that 400 km of oceanic crust has been created on the western side and only 160 km on the eastern side of the Hannibal Ridge. These correspond to rates of 5 cm yr⁻¹ and 2 cm yr⁻¹, respectively. Such difference in the same setting have been observed elsewhere. For instance, the East Scotia Ridge is a spreading centre located between the Scotia Plate in the west and the Sandwich Plate in the east. In an absolute reference frame (Giner-Robles et al., 2003) the Sandwich Plate, which is adjacent to the Sandwich Trench, is moving more rapidly (5.5 cm yr⁻¹) than the Scotia Plate (2.2 cm yr⁻¹). A value of 5 cm yr⁻¹ is a high rate of opening but is similar to the rate observed for the Tyrrenhian Sea (4 cm yr⁻¹; Gueguen et al., 1998; Doglioni et al., 1999). It is clear that the opening in both cases is passive and related to the rollback of the trench and its pull effect (Lonergan and White, 1997; Malinverno and Ryan, 1986; Royden, 1993). After the late Tortonian, the African Plate moved 65 km to the north and the margin became transpressional (Auzende et al., 1975; El Robrini, 1986), as shown by the present compressional seismicity (Argus et al., 1989; Meghraoui et al., 1996). The southern part of the Hannibal Ridge may be subducted beneath the Africa Plate.

Conclusions We propose that the Algerian Basin opened when widening of the Provençal Basin stopped (16 Ma) and that the opening finished when the
The Tyrrhenian Basin began to open (8 Ma; Gueguen et al., 1998; Doglioni et al., 1999; Sartori et al., 2001). However, some activity may have persisted until as late as the Messinian (6 Ma). Formation of the Tyrrhenian Sea and Alboran Sea is not directly related to the slow N-S convergence of the two large plates but resulted from the fast rollback of two arcs migrating east and west, respectively (Lonergan and White, 1997; Frizon de Lamotte et al., 2000) towards the free faces (Ionian Sea and Atlantic Ocean, respectively) (Gelabet et al., 2002). Here we present a new hypothesis, with E-W opening of the Algerian Basin behind the Gibraltar migrating arc. Traction of the Apennines Arc in the North Tyrrhenian region may have had a minor influence on the rifting and opening in the south.

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References


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