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The Alboran Sea (Western Mediterranean) revisited with a view from the Moroccan Margin

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ABSTRACT

This paper presents a new depth to acoustic basement map based on a grid of seismic profiles. In the Western Alboran Basin the basement is more than 10 km deep. An examination of a seismic profile of this region suggests that Late Oligocene–Early Miocene 5 km thick synrift formation has been accumulated in this basin. This West Alboran basin has been formed in an extensional setting, but in a fore-arc position, south of the Balearic Islands then westwards transported behind a subduction zone.

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Introduction

The Alboran Sea, located at the westernmost part of the Mediterranean Sea, is surrounded by the Betics-Rif Arc (Fig. 1B inset). There is a large consensus to place south of the Eurasia Plate, before the Miocene, an original landmass formed by the **Al**boran Realm, the **Ka**bylies, the **Pe**loritan Range in Sicily and **Ca**labria in South Italy (Bouillin, 1986). The Early Miocene southward rollback of this AlKaPeCa domain and the correlative formation of back-arc basins are driven by the north-dipping subduction of the African Plate beneath the Eurasia Plate (Rehault *et al.*, 1984; Jolivet and Faccenna, 2000). The dispersion of the Betics-Rif–Alboran Realm and the Peloritan-Calabria landmass resulted from a westwards (Lonergan and White, 1997; Frizon de Lamotte *et al.*, 2000; Mauffret *et al.*, 2004) and eastwards migration (Doglioni *et al.*, 1997; Guegen *et al.*, 1998), respectively, after the docking of the Kabylies to the Africa Plate. Several hypotheses tried to explain the extensional formation of the Alboran Sea in the core of the Betics-Rif Orogen. One hypothesis (Platt and Vissers, 1989; Platt *et al.*, 2003) favours the extensional collapse of thickened crust and the convective removal of continental lithosphere. Another hypothesis invokes a delamination process (Seber *et al.*, 1996; Calvert *et al.*, 2000). A retreating subduction towards the west is also proposed (Lonergan and White, 1997; Gutscher *et al.*, 2002; Rosenbaum and Lister, 2004) and this hypothesis is strongly supported by the tomographic studies (Faccena *et al.*, 2004; Spakman and Wortel, 2004) and the on-land geology (Frizon de Lamotte *et al.*, 1991; García-Dueñas *et al.*, 1992; Martínez-Martínez and Azañón, 1997; Frizon de Lamotte *et al.*, 2000). The Alboran Sea structure is now very well described after several academic cruises (Willet, 1991; Maldonado *et al.*, 1992; Watts *et al.*, 1993), industrial seismic surveys (Comas *et al.*, 1992), and boreholes (Jurado and Comas, 1992). This multidisciplinary approach culminated in the ODP Leg 161 (Comas *et al.*, 1999). Although several papers (Bourgois *et al.*, 1992; Morley, 1993; Chalouan *et al.*, 1997) presented the main results of the seismic surveys off Morocco we revisited this margin, in particular the Western Alboran Basin, to know precisely its structural evolution.

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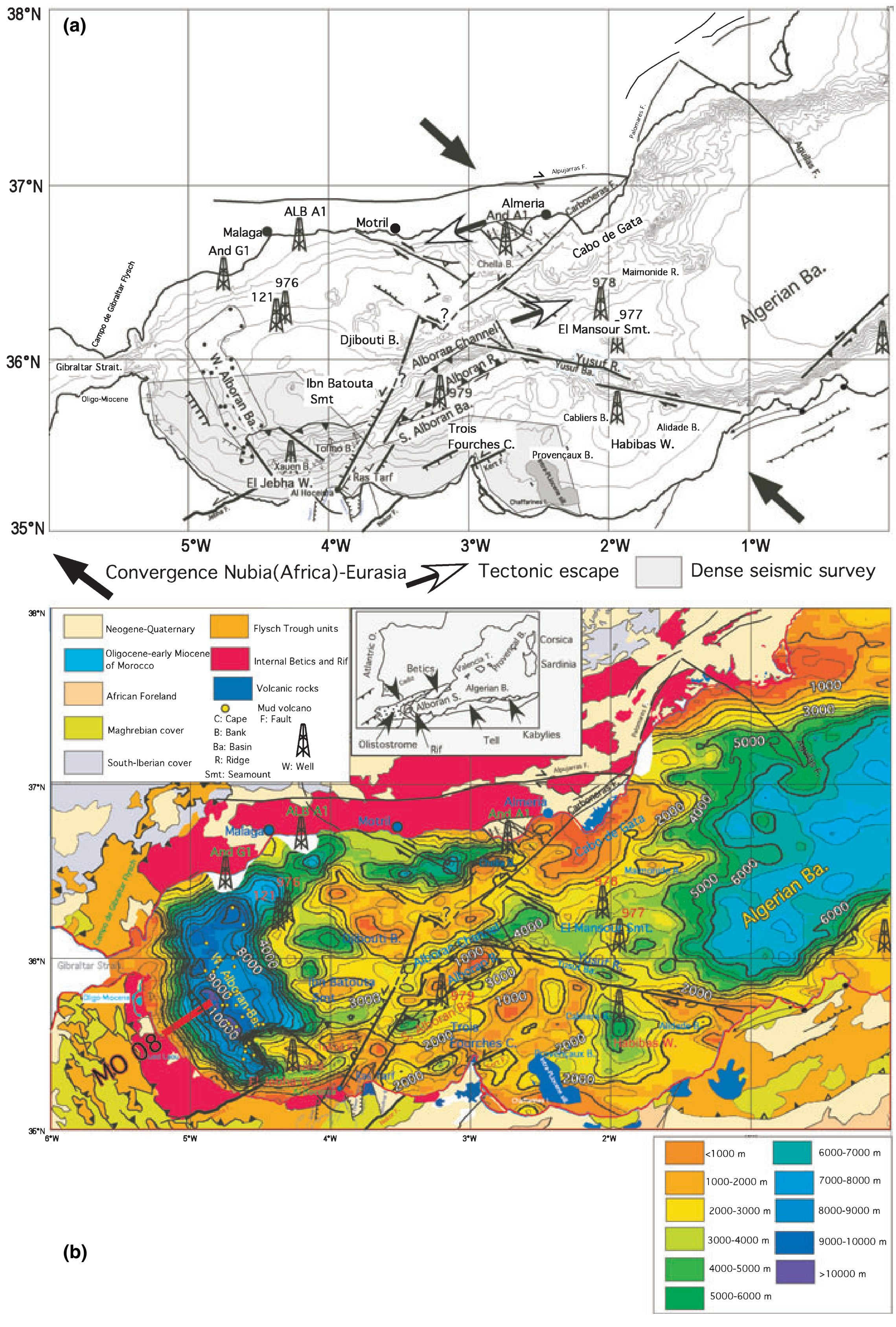
Brief description of the Alboran Sea structure

A very accurate tectonic map has been already presented (Comas *et al.*, 1999) and this sketch is precised here (Fig. 1B) mainly in the southern part of the Alboran Sea. This depth to the acoustic basement map is derived from about 10 000 km of seismic profiles, with a dense seismic grid on the Morocco margin (Fig. 1A), digitized with a value every 250 m. The Alboran Sea is divided into two main domains: the eastern part adjacent to the Algerian oceanic Basin and the western part where lies the Western Alboran Basin. The eastern part is characterized by an abundant cal-alkaline volcanism (Comas *et al.*, 1999; Duggen *et al.*, 2004). The volcanic rocks onshore and sampled offshore by submersible and ODP Sites 977 and 978 are dated between 11.8 and 6.57 Ma (Comas *et al.*, 1999; Duggen *et al.*, 2004). The eastern Alboran Sea is also characterized by large strike–slip faults: Yusuf Fault (Mauffret *et al.*, 1987; Alvarez-Marrón, 1999); Carboneras Fault onshore (Martínez-Díaz and Hernández-Enrile, 2004) and offshore (Rodriguez-Fernandez and Martin-Penela, 1993; Gràcia *et al.*, 2006). This fault seems to be connected to another large left-lateral strike–slip fault that is probably related (Ammar *et al.*, unpublished data) to the large seismicity of the Al Hoceima area (Calvert *et al.*, 1997; El Alami *et al.*, 1998). This fault system forms the western boundary of the East and Central Alboran Sea. The Alboran Ridge is a complex anticline with a left-lateral strike–slip component and a volcanic core (Bourgois *et al.*, 1992; Comas *et al.*, 1999). The Tofino and Xauen banks are offset from the Alboran Ridge (Fig. 1, Ammar *et al.*, unpublished data) and have not the same composition. The seismic


(a)
38°N
37°N
36°N
35°N
5°W
4°W
3°W
2°W
1°W
Almeria
And A1
ALB A1
Motril
Malaga
And G1
Cabo de Gata
Maimonide R.
978
977
El Mansour Smt.
976
121
Djibouti B.
Alboran Channel
Alboran R.
979
Yusuf R.
Algerian Ba.
Gibraltar Strait.
Oligo-Miocene
W. Alboran Ba.
Ibn Batouta
Smt
S. Alboran Ba.
Trois
Fourches C.
Cabliers B.
Habibas W.
Alidade B.
Provençaux B.
Xauen B.
Tofiño B.
Ras Tarf
El Jebha W.
Al Hoceima
Chella B.
Campo de Gibraltar Flysch
Alpujarras F.
Carboneras F.
Palomares F.
Aguilas F.
Convergence Nubia(Africa)-Eurasia
Tectonic escape
Dense seismic survey
Neogene-Quaternary
Flysch Trough units
Oligocene-early Miocene of Morocco
Internal Betics and Rif
African Foreland
Volcanic rocks
Mud volcano
C: Cape
B: Bank
Ba: Basin
R: Ridge
Smt: Seamount
F: Fault
W: Well
Maghrebian cover
South-Iberian cover
Atlantic O.
Betics
Valencia T.
Provençal B.
Corsica
Sardinia
Algerian B.
Alboran S.
Olistostrome
Rif
Tell
Kabylies
MO 08
(b)
<1000 m
1000-2000 m
2000-3000 m
3000-4000 m
4000-5000 m
5000-6000 m
6000-7000 m
7000-8000 m
8000-9000 m
9000-10000 m
>10000 m

profiles and the Jebha well suggest (Gensous *et al.*, 1986; Bourgois *et al.*, 1992; Morley, 1993; Chalouan *et al.*, 1997) that these banks have been formed by the compression and uplift of sedimentary layers (Early Miocene reworked and Langhian to Pliocene in the well) that should have been previously deposited in the deep Western Alboran Basin.

Western Alboran Basin

This basin is bounded (Fig. 1) by the Moroccan Margin (Internal Rif), the Gibraltar Strait (Campo de Gibraltar Flysch) and the Spanish Margin (Internal Betics). With the Motril, and Malaga basins to the north (Watts *et al.*, 1993) and the Tofino-Xauen banks area to the south, the Western Alboran Basin shows a horseshoe shape that mimics the shape of Rif-Betics Arc. This shape can be also observed along the Ibn Batouta Seamount and the DSDP 121–ODP 976 horst (Chalouan *et al.*, 1997; Comas *et al.*, 1999). The Ibn Batouta Seamount shows a magnetic signature (Willet, 1991) that suggests a volcanic composition, although the deep basement is unknown and is probably made of the same metamorphic rocks (Alpujarride) as the ODP 976 ridge (Comas *et al.*, 1999). The bottom of the Western Alboran basin is deeper than 10 km in depth to the acoustic basement map (Fig. 1B). A recent seismic survey consulted by one of the authors (H. Jabour) indicates that the depth is greater than 12.5 km. With the Moho 18 km deep (Torné *et al.*, 2000), the crust is only 5.5 km thick and almost oceanic. The curves generated by automatic contouring are smoothed, and the centre of the basin is occupied by mud diapirs that disturb the sedimentary layers (Mulder and Parry, 1977; Campillo *et al.*, 1992; Pérez-Belzuz *et al.*, 1997; Talkuder *et al.*, 2003) and conceal the deep part of the seismic profiles.

The depth to the acoustic basement map (Fig. 1B) has been obtained by the analysis and digitization of old seismic profiles (Fig. 1A) which have been partly already published (Morley, 1993; Chalouan *et al.*, 1997). The apparent absence of extensional fault suggested to Morley (1993) a sag basin although these faults, facing the continent, are described along the DSDP 121–ODP 976 horst (Comas *et al.*, 1992) which acts as the footwall of the Malaga and Motril basins. Low angle normal faults dipping towards the sea are also described on-land in Spain (García-Dueñas *et al.*, 1992) and Morocco (Chalouan *et al.*, 1995).

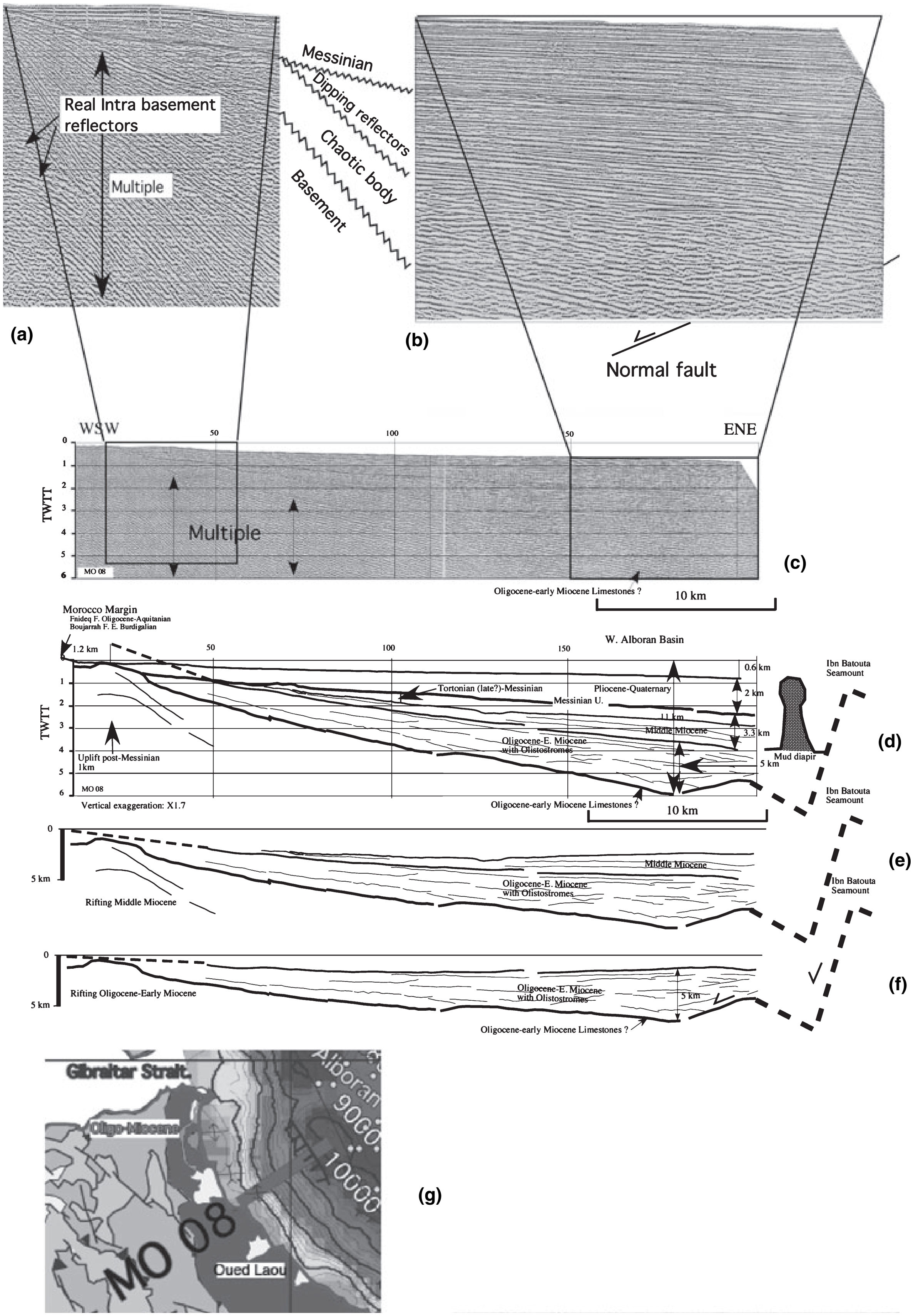
Seismic profile crossing the Moroccan margin

A change of dip in the basement observed at the eastern extremity of the profile (Fig. 2B) maybe interpreted as a normal fault facing the continent with a low angle (Chalouan *et al.*, 1997). However, the diapir field begins just after the eastern end of the profile (Fig. 1) and the faults maybe concealed by these diapirs. Nevertheless, the Ibn Batouta–DSDP 121–ODP 976 horst is the main footwall of the Western Alboran Basin. It is obvious that a very important tilting towards the Mediterranean Sea (Fig. 2) occurred along the margin of Morocco. A post-Messinian 1-km uplift is inferred from the seismic geometries and field data. In the Oued Laou valley a Pliocene ria infilled a Messinian canyon and Pliocene oysters have been recovered 500–600 m above the present sea level (Wildi and Wernli, 1977; Wernli, 1988). If an extensional regime is described during the Pliocene–Quaternary (Morel, 1989; Saji and Chalouan, 1995) those extensional faults may be related to a transtension. Indeed, the NW–SE stress field related to the convergence of the Eurasian and African Plates did not change during this period (Martínez-Díaz and Hernández-Enrile, 2004), and several evidences of E–W folds, that are probably responsible of the uplift, are observed on the seismic profiles (Fig. 2A). A strong erosion occurred during the Messinian and 1.2 km thick sedimentary layer, including the deepest formation has been removed from the upper margin and probably the onshore (Fig. 2D). Before the tilting due to the uplift of the continent, the middle Miocene series showed a fan-shaped configuration that suggests a rifting (Fig. 2E). This rifting can be dated middle Miocene (Langhian–Serravallian–Early Tortonian?) (Comas *et al.*, 1992; Mayoral *et al.*, 1994; Chalouan *et al.*, 1997) but the Tortonian (late?)–Messinian is not affected by the tilting (Fig. 2D). The main rifting episode is recorded in the deepest chaotic sedimentary sequence, which is as thick as 5 km (Fig. 2E). This formation is very well known from wells (Jurado and Comas, 1992), in the sediments transported to the surface by mud diapirs (Sautkin *et al.*, 2003) and land geology. This Late Oligocene–Early Miocene olistostrome unit is made of a mixture of marls, sandstones, carbonaceous material and reworked Late Cretaceous–Palaeogene clasts. The chaotic nature of this formation is well illustrated by the seismic profiles (Comas *et al.*, 1992; Chalouan *et al.*, 1997). Figure 2F suggests that this lower formation extended originally on-land, probably connected to the Fnideq and Boujarrah Late Oligocene–Early Miocene formations (Feinberg *et al.*, 1990; Maaté *et al.*, 1995) and has been eroded during the Messinian.

Discussion

There is a general consensus to consider the Provençal Basin as a back-arc basin with a rifting during the Late Oligocene–Aquitanian (Rehault *et al.*, 1984; Gorini *et al.*, 1993), followed by a drifting during the Burdigalian

Fig. 1 (a) Main tectonic features superimposed on a bathymetric map (Intergovernmental Oceanographic Commission, 1981). The dense seismic survey of the Morocco Margin has been performed by several oil companies. Black arrows show present plate convergence direction (McClusky *et al.*, 2003). White arrows suggest the tectonic escape not only to the west (Martínez-Díaz and Hernández-Enrile, 2004) but also to the east. (b) Depth to acoustic basement map of the Alboran Sea. Contour interval: 0.5 km. On-land geology from El Alami *et al.* (1998), Comas *et al.* (1999), Duggen *et al.* (2004) and Crespo-Blanc and Frizon de Lamotte (2006). Offshore geology from Gensous *et al.* (1986), Mauffret *et al.* (1987), Bourgois *et al.* (1992), Rodriguez-Fernandez and Martin-Penela (1993), Chalouan *et al.* (1997), Alvarez-Marrón (1999), Comas *et al.* (1999) and Gràcia *et al.* (2006).


Real Intra basement reflectors
Multiple
Messinian
Dipping reflectors
Chaotic body
Basement
(a)
(b)
Normal fault
WSW
ENE
50
100
150
TWTT
MO 08
Multiple
(c)
Oligocene-early Miocene Limestones ?
10 km
Morocco Margin
Fnideq F. Oligocene-Aquitanian
Boujarrah F. E. Burdigalian
1.2 km
W. Alboran Basin
0.6 km
Ibn Batouta Seamount
Tortonian (late?)-Messinian
Pliocene-Quaternary
2 km
Messinian U.
11 km
Middle Miocene
3.3 km
Oligocene-E. Miocene with Olistostromes
(d)
Uplift post-Messinian 1km
5 km
Mud diapir
Vertical exaggeration: X1.7
(e)
Rifting Middle Miocene
(f)
Rifting Oligocene-Early Miocene
Gibraltar Strait.
Oligo-Miocene
Alboran
MO 08
Oued Laou
(g)

(Rosenbaum *et al.*, 2002; Speranza *et al.*, 2002). The Valencia Trough rifting is coeval to the Provençal rifting or slightly younger but there the extensional process aborted and there is no evidence of the generation of the oceanic crust (Maillard and Mauffret, 1999). The north-west dipping subduction zone was located between the internal (AlKaPeCa; Bouillin, 1986) massifs and the Tethys oceanic crust covered by the Flysch Trough units (Guerrera *et al.*, 2005). The volcanic arc was placed in the southern boundary of the back-arc behind the Balearic Islands (Martí *et al.*, 1992), on the western part of Sardinia (Savelli, 2002) and west of Corsica (Rossi *et al.*, 1998). The Corsica and Sardinia Basins (Sartori *et al.*, 2001), the Late Oligocene–Aquitanian arkosic basins of Algeria, Fnideq Unit in Morocco (Feinberg *et al.*, 1990; Ouazani-Touhami and Chalouan 1995); Ciudad Granada Formation in Spain (Guerrera *et al.*, 2005) and the Western Alboran Basin, before its migration towards the west, were clearly placed in a fore-arc position between the volcanic arc and the subduction zone (Mauffret *et al.*, 2004). However, the Western Alboran Basin and the related terranes in Spain and Morocco underwent an evident extensional episode coeval to the rifting of the Provençal Basin. The slab rollback is the probable cause of this extension (Yamaji, 2003) and the segmentation of the fore-arcs (Guarnieri, 2006). Although the extension in fore-arc basins has not been frequently described, a volcanism related to the rifting may occur (Macdonald *et al.*, 1999). It is obvious that the Western Alboran Basin had on-land extensions in Morocco and Spain that have been obliterated by the Tortonian to Present compression (Comas *et al.*, 1992). The dikes described in the Malaga region are contemporaneous with the rifting and dated 22–23 ma (Aquitanian) by K–Ar method (Torres-Roldan *et al.*, 1986) and 17.4–19.8 Ma by $^{40}Ar/^{39}Ar$ (Duggen *et al.*, 2004). The Aquitanian rifting episode and the correlative crustal thinning are coeval (21–23 Ma) to a thermal peak and HT/LP metamorphism (Platt *et al.*, 2003). Moreover, the exhumation of the HP/LT metamorphic Alpujarride–Sebtide complex is recorded at the Aquitanian–Burdigalian boundary (20 Ma) by the presence of boulders and breccias, related to a large normal faulting, of Alpujarride rocks in the Burdigalian Vinuela formation (zoom Fig. 3D). In the older formations (Ciudad Granada and Fnideq, Fig. 3C) only Ghomaride–Malaguide non-metamorphic clasts are present in a shallower environment than the Burdigalian formations (Durand-Delga *et al.*, 2003). In these clasts the presence of granitic boulders with an Alpine metamorphic imprint suggest a Lost Realm originally placed at the proximity of the Galite Block (Bouillin *et al.*, 1999) and Sicily (Martín-Algarra *et al.*, 2000). The rising of the HP rocks (Alpujarride–Sebtide) may occur along the subduction channel (Burov *et al.*, 2001; Chemanda *et al.*, 2001; Jolivet *et al.*, 2003) that may have been opened by the rollback. In Fig. 3A, the West Alboran Basin trends NE–SW, parallel to the Valencia trough. However, the initial configuration has been completely disrupted by the subsequent westwards motion, the rotation of blocks (Lonergan and White, 1997), and the partial integration of the basin to the continent. The initial rifting of the fore-arc and back-arc basins are coeval and probably related to the rollback but the rifting persisted during the lower Burdigalian in the fore-arc and maybe in the Valencia Trough and stopped in the Mid Burdigalian (18 Ma) when the Kabylies docked to the Africa Plate and the Maghrebian Flysch Basin closed (Maaté *et al.*, 1995).

At the end of the Burdigalian (16 Ma) a complete reorganization of the plate motions (Guegen *et al.*, 1998; Gelabert *et al.*, 2002) occurred with a westwards migration of an Alboran Plate (Frizon de Lamotte *et al.*, 2000). This motion is corroborated by on-land geology (Frizon de Lamotte *et al.*, 1991; García-Dueñas *et al.*, 1992), tomography (Faccena *et al.*, 2004; Rosenbaum and Lister, 2004; Spakman and Wortel, 2004) and plate tectonics that implies a three-plate configuration with a slow convergence of the Africa and Eurasia Plates (Dewey *et al.*, 1989) and a fast westwards migration of an Alboran microplate (Mauffret *et al.*, 2004). The Alboran volcanic Arc (Duggen *et al.*, 2004) was located behind a narrow slab rolling back westwards and the West Alboran Basin had again a fore-arc position and underwent a middle Miocene rifting with a N–S trend at this time (Martínez-Martínez and Azañón, 1997). Behind the Alboran Arc, the Algerian Basin opened in a back-arc position with formation of oceanic crust (Mauffret *et al.*, 2004). We agree with Gutscher *et al.* (2002) to correlate the giant olistostrome located in the Gulf of Cadiz and the Atlantic Ocean (Fig. 4) with a narrow slab dipping eastwards beneath the Alboran Sea. However, this subduction is probably not active at Present because a gap is recorded between 0 and 50 km in the intermediate seismicity (50–150 km) located beneath the western Alboran Basin (Buforn *et al.*, 1997). Moreover, the NNW–SSE convergence between the Eurasia and Africa Plates is again perceptible since the Tortonian in the Alboran Region with compression, strike–slip faults and normal faults trending parallel to the vector of convergence (Stich *et al.*, 2003; Martínez-Díaz and Hernández-Enrile, 2004). This normal faulting is probably not only related to a westwards tectonic escape (Martínez-Díaz and Hernández-Enrile, 2004) but also eastwards (Fig. 1A). The end of the westwards motion and the beginning of the convergence with

Fig. 2 Seismic profile (2.4 km long streamer, 24 folds, stack) that trends ENE–WSW crossing the Morocco margin and the western side of the West Alboran Basin. The low vertical exaggeration (1.7) suggests a scale 1/1 at depth. (a) Zoom showing the Messinian erosion (1.2 km) and a fold. (b) Zoom of the chaotic body and dipping reflectors. Observe the normal fault. (c) Seismic profile. (d) Line drawing. (e) Rifting middle Miocene. The top of the Middle Miocene has been placed in horizontal position. (f) Rifting Oligocene–Early Miocene. The top of the Early Miocene has been placed in horizontal position. (g) Zoom of the depth to basement map. Note the outcrops on-land of the Oligo-Miocene (Fnideq-Boujarrah) formations (Feinberg *et al.*, 1990; Maaté *et al.*, 1995).

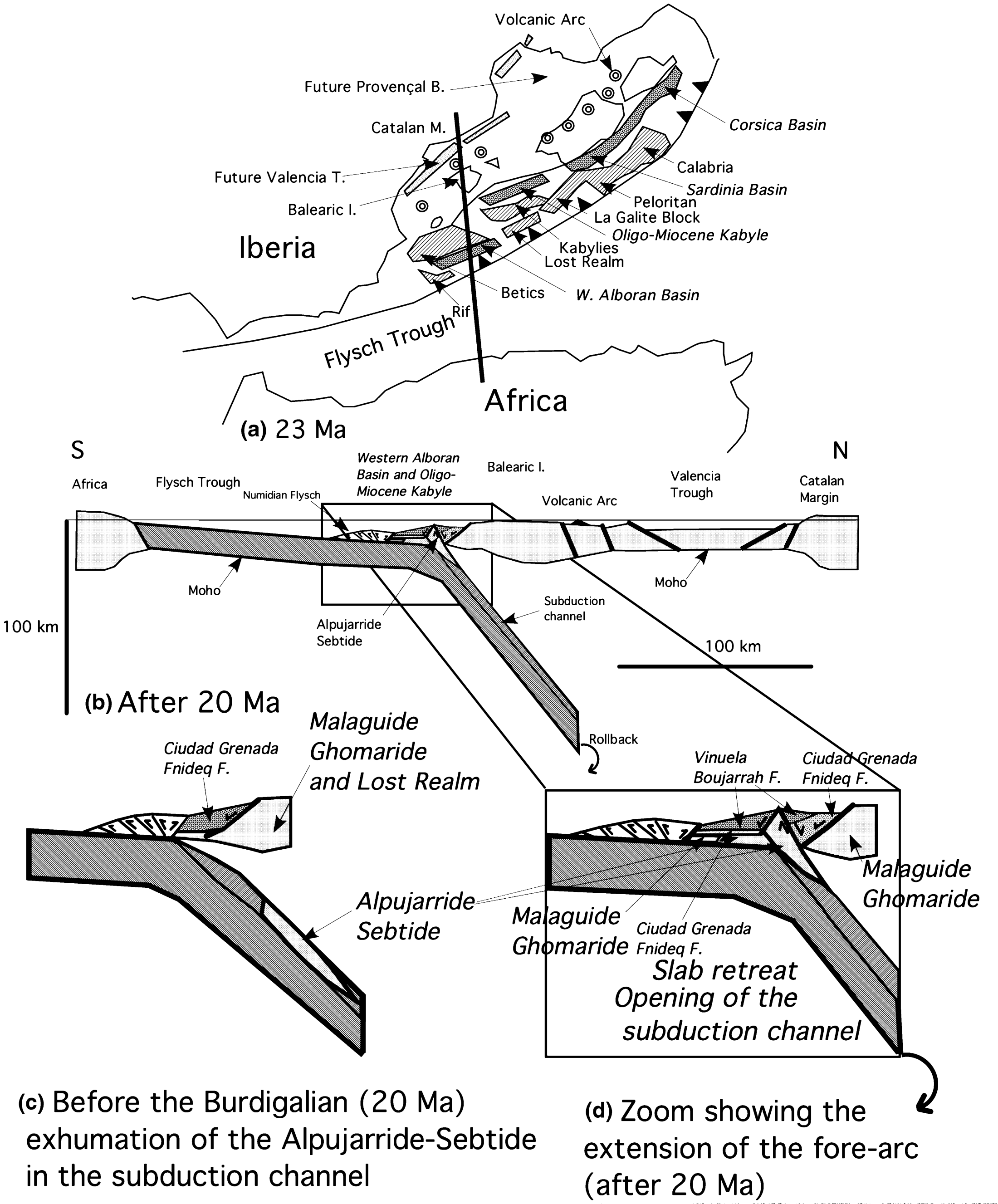


Fig. 3 (a) Tectonic reconstruction at 23 Ma (Mauffret *et al.*, 2004; modified). The Corsica and Sardinia Basins, (Sartori *et al.* 2001), the Algerian Oligo-Miocene basins and the West Alboran Basin are placed in a fore-arc position between the volcanic arc and the subduction zone. Note the presence of a Lost Realm (Martín-Algarra *et al.* 2000) near the Galite Block (Bouillin *et al.* 1999). (b) Cross–section from the Catalan Margin to Africa (after 20 Ma). (c) Situation before 20 Ma. (d) Zoom.The extension in the back-arc (Provençal Basin and Valencia Trough but also in the fore arcs is related to the rollback of the retreating slab. This rollback may provoke also the rising of the HP metamorphic rocks (Alpujarride and Sebtide) in the subduction channel (Burov *et al.* 2001, Chemanda *et al.* 2001, Jolivet *et al.* 2003).

a cannibalization of the Alboran Sea by the adjacent land by uplift and folding could be Late Tortonian (Comas *et al.*, 1992, 1999). However, some motion may persist up to the Messinian because the shoshonites of the Cabo de Gata, that indicate the end of the subduction, are 6.57 Ma old (Duggen *et al.*, 2004) and some activity is yet recorded on the Hannibal Spreading Centre (Mauffret *et al.*, 2004).

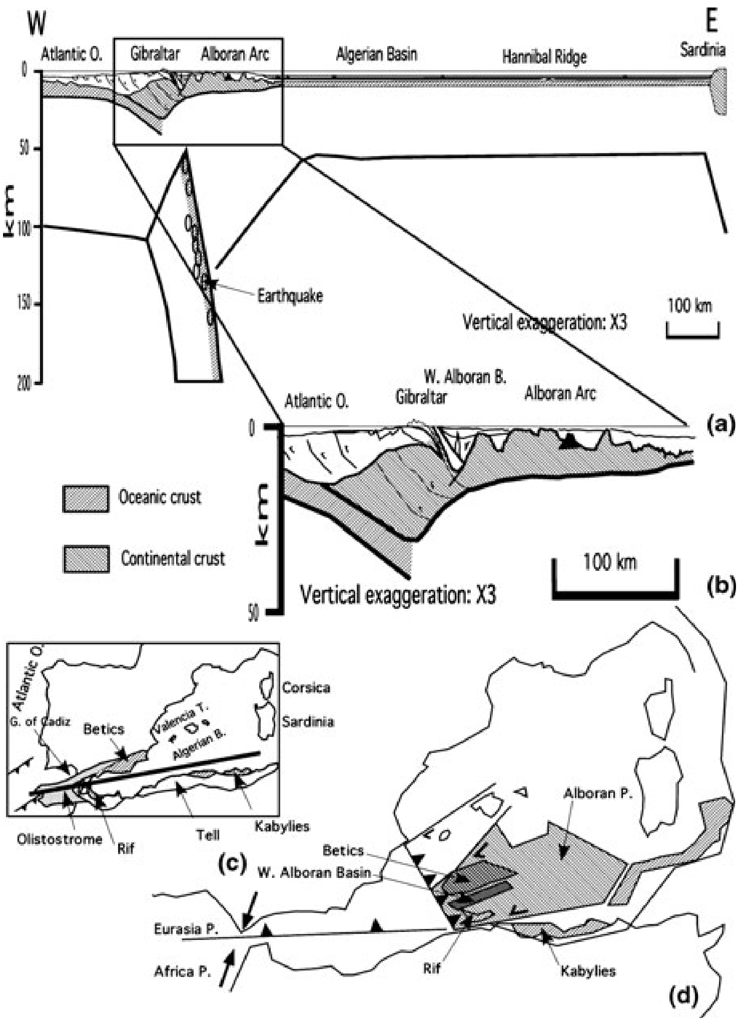


Fig. 4 (a) Present E–W cross-section from Sardinia to the Atlantic Ocean. (b) Zoom. (c) Location of the cross-section. The olistostrome area in the Atlantic Ocean is reconstructed from deep penetration seismic profiles and OBS (Gutscher *et al*., 2002; Medialdea *et al*., 2004). The Alboran continental thinned crust is drawn from gravity modelling (Torné *et al*., 2000) and seismic profiles (this work). The deep structure of the lithosphere is inferred from gravity modelling (Torné *et al*., 2000) and tomographic studies (Faccena *et al*., 2004; Rosenbaum and Lister, 2004; Spakman and Wortel, 2004). Earthquakes located at intermediate depth 50–150 km from Buforn *et al*. (1997). (d) Tectonic reconstruction at 16 Ma (Mauffret *et al*., 2004; modified). Vector of convergence Africa–Eurasia from Dewey *et al*. (1989).

Conclusions

The geophysical and geological data favour a westwards transport of the Western Alboran Basin behind a Middle–Late Miocene retreating subduction. This basin was placed in a fore-arc position south of the Balearic volcanic arc and close to a granitic source. The fore-arc position may explain the interactions between compression, extension and the exhumation of high-pressure metamorphic terranes.

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