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Lithospheric-scale geodynamic context of the Messinian salinity crisis

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Abstract

Comparison between the tectonic and palaeogeographic evolution of the Mediterranean region from the Oligocene to the Present shows that the isolation of the Mediterranean region is a progressive process that started some 30–35 Ma ago. This comparison also emphasizes the importance of back-arc post-orogenic extension in forming the present-day tectonic setting and delaying the Messinian salinity crisis. It further shows that the Messinian salinity crisis occurred after the end of late-orogenic extension in the Alboran domain and the onset of a new stage of N–S compression in the westernmost Mediterranean. The crisis thus did not result from the slab retreat process that has instead probably favoured the connections between the Atlantic and Mediterranean waters during most of the Miocene. Conversely, without slab retreat and consequent extension, compression and crustal thickening would have instead closed the connections earlier and provoked an earlier salinity crisis. The slab retreat thus has probably delayed significantly the salinity crisis that is a logical consequence of the progressive isolation of the Mediterranean basin during convergence.

Keywords: Messinian; Slab retreat; Collision; Mediterranean; Salinity

1. Introduction

The concept of a Messinian salinity crisis in the Mediterranean region dates back to the seventies (Hsü et al., 1973a,b) after coring in the deep basins (DSDP Leg 13) the top of a thick Messinian evaporitic layer that was suggested by Ryan (1969) based on the observation of seismic profiles. Immediately, this catastrophic event was considered at its true scale and the large evaporitic deposits already known onland in Sicily (Ogniben, 1957), in Algeria (Perrodon, 1957), in the Betics (Rios, 1968), in Crete (Freudenthal, 1969) or in Cyprus (Gass and Cockbain, 1961) for instance were understood as uplifted segments of the evaporite formations in regions of active Plio-Pleistocene tectonics. This discovery triggered an active discussion between tenants of the various possible models (shallow or deep basin, deep or shallow sedimentation…) (Hsü et al., 1973a,b; Nesteroff, 1973; Fabricius and Hieke, 1976; Hsü, 1977; Cita et al., 1978; Busson, 1979; Rouchy, 1980). A relative consensus has been reached today and the "deep basin –

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shallow water" model has been widely accepted although various modalities of the exact scenario are still discussed (Rouchy and Saint Martin, 1992; Clauzon et al., 1996; Krijgsman et al., 1999a; 1999c).

The recent years have brought a wealth of new data on the exact timing of the onset and end of the crisis that are really synchronous over the whole Mediterranean (Gautier et al., 1994; Clauzon et al., 1996; Krijgsman et al., 1999a,c, 2002). Available scenarios differ by the velocity and intensity of dessication [one stage (Krijgsman et al., 1999c), two stages (Clauzon et al., 1996)] or several repeated stages of water levels drawdown (Rouchy, 1980)], on the significance and timing of the marginal basins compared to deep ones (as well as on the timing of the corresponding erosion), and on the significance of one of the most actively studied evaporitic deposit, the Sicilian Basin [deep basin (Rouchy, 1980; Krijgsman, 2002) or marginal basin (Clauzon et al., 1996)]. Little work has been done on the interaction between tectonic and evaporitic sedimentation except for the Sicilian and Apenninic basins where correlations between sedimentation of evaporites and the development of thrust and folds have been documented (Butler et al., 1995, 1999; Roveri et al., 2001; Matano et al., 2005).

Since the early 70s a tectonic cause has been felt behind the salinity crisis. Early papers already put forward the fact that the closure of the gates of Gibraltar and the connection with the Atlantic Ocean has a tectonic origin (Hsü et al., 1973b; Hsü, 1977). It has been also suggested that global sea level changes may have played a role whose importance is still debated to tune the main dessication events in the Mediterranean region (see Adams et al., 1977; Hodell et al., 1986, Thunell et al., 1987; Kastens, 1992; Rouchy and Saint Martin, 1992; Aharon et al., 1993; Clauzon et al., 1996).

The ultimate tectonic cause of the crisis has not yet been clearly identified. In a very broad sense it appears clearly that the progressive closure of the Western Mediterranean domain is responsible for the limitation of water exchanges with the Atlantic. However, drastically different scenarios were proposed. The reduced width of the gates of Gibraltar can merely result from the N–S shortening that has been active throughout the Tertiary leading to the shortening and final closure of the Betic and Rifian corridors shortly before the salinity crisis, respectively between 7.6 Ma and 6.8 Ma (Martín et al., 2000; Garcés et al., 2001) and between at least 6.0 Ma and 5.65–5.4 Ma (Krijgsman et al., 1996, 1999b; Warny et al., 2003). It could also be instead envisaged as the result of the westward retreat of the Betic slab (Duggen et al., 2003, 2004). This retreat and resulting removal of the base of the continental lithosphere would have resulted in a generalized uplift of the Alboran domain and closure of the Mediterranean–Atlantic connection. The first scenario involves compression, the second ones involves uplift, slab retreat and upper-plate extension.

In order to discuss the deep causes of the salinity crisis we present a set of paleogeodynamic and paleogeographic reconstructions, and we review the available data that constrain the timing of tectonic, magmatic and sedimentological events. We show that the geodynamic evolution is mainly controlled by back-arc extension and slab retreat from the late Oligocene until the present in the Eastern and Central Mediterranean. Conversely in the Western Mediterranean extension has stopped some 8 Ma ago and was replaced by N–S compression that is still active today (Comas et al., 1999). The closure of the connection with the Atlantic Ocean thus cannot be the result of slab retreat but rather of a renewal of N–S compression, and slab retreat seems instead to have delayed the salinity crisis which is a logical consequence of the convergence process and progressive isolation of the Mediterranean. We also discuss the connection with the Paratethys in the east and the propagation of the North Anatolian Fault in the Aegean domain.

2. Mediterranean geodynamics and basin formation

The geodynamic evolution of the Mediterranean region during the Cenozoic (Fig. 1) is under the control of two contemporaneous phenomena: (1) Africa–Eurasia convergence and northward subduction of the African plate and (2) slab retreat and coeval back-arc basin opening (Le Pichon and Angelier, 1981; Horvath and Berkhemer, 1982; Dercourt et al., 1986; Malinverno and Ryan, 1986; Dewey, 1988; Le Pichon et al., 1988; Royden, 1993; Carminati et al., 1998; Jolivet and Faccenna, 2000; Wortel and Spakman, 2000). The subduction and ensuing collision processes have led to the formation of mountain belts from the Betic Cordillera and the Rif in the west, to the Hellenides and Taurides in the east. Since the late Oligocene and the collision between Africa and Eurasia in the west and east, the Mediterranean subduction has been locked between two collision zones and slab retreat became predominant (Jolivet and Faccenna, 2000), leading to the formation of four major back-arc basins: the Alboran Sea, the Liguro–Provençal Basin, the Tyrrhenian Sea and the Aegean Sea (Fig. 1A and B). This second period has shaped

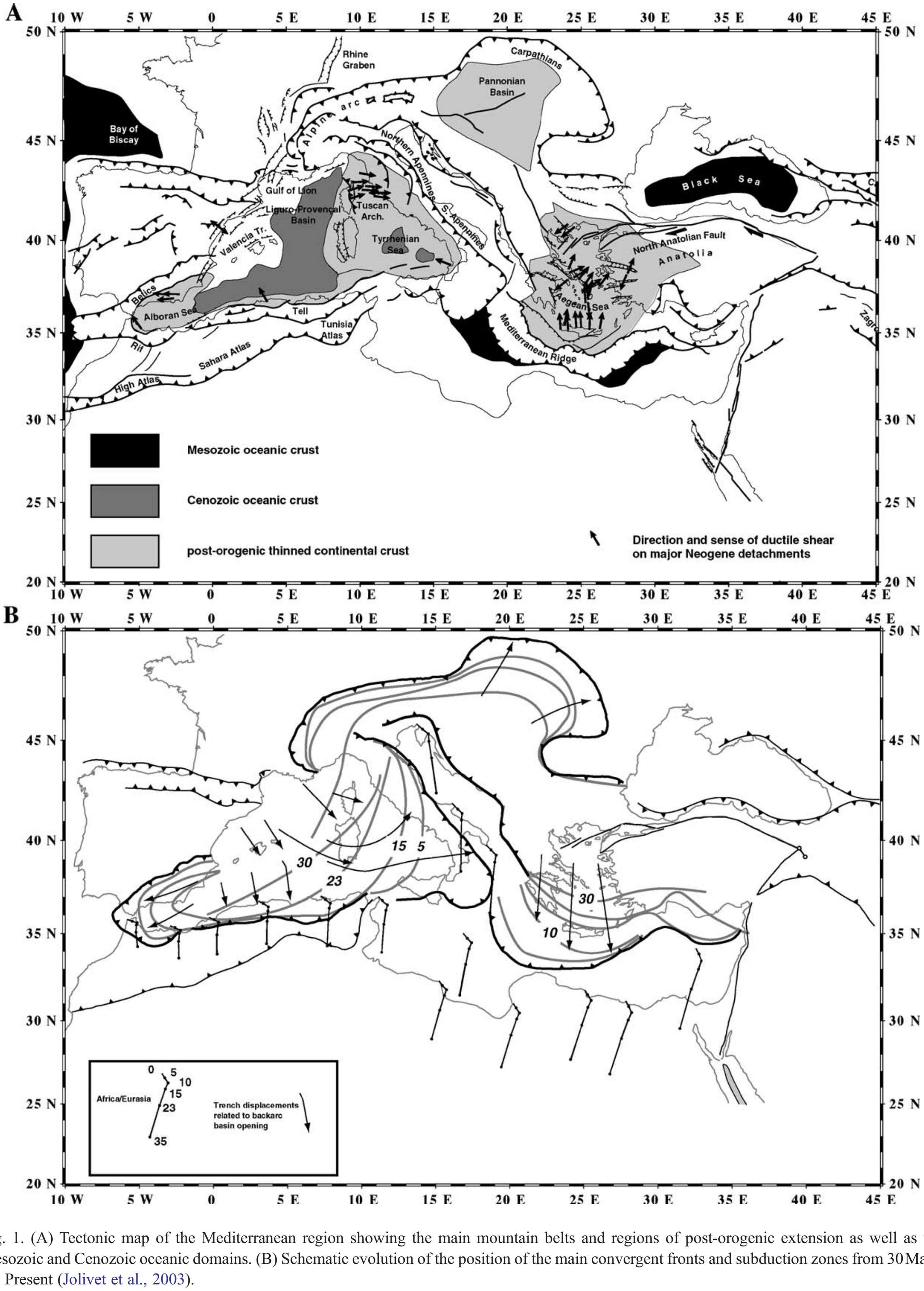


Fig. 1. (A) Tectonic map of the Mediterranean region showing the main mountain belts and regions of post-orogenic extension as well as the Mesozoic and Cenozoic oceanic domains. (B) Schematic evolution of the position of the main convergent fronts and subduction zones from 30 Ma to the Present (Jolivet et al., 2003).

the present-day physiography of the Mediterranean basins (Fig. 1A).

The formation of the Alps and the Carpathians is under the control of the subduction of the European plate below Apulia, an emissary of the African plate. The retreat of the European slab allowed the formation of the Pannonian Basin in the east, although its dynamics cannot be compared simply to the true back-arc basins of the Mediterranean region (Linzer, 1996; Horvath and Tari, 1999; Wortel and Spakman, 2000; Cloething et al., 2004).

The Mediterranean basin was isolated from the Tethys and the Indian Ocean in the east by the collision between Africa–Arabia and Eurasia, progressively from the Eocene to the Miocene, and the connections definitely closed during the Langhian and Serravalian (Rögl and Steininger, 1983; Popov et al., 2004). The deep oceanic domain in the future Betic–Rif and Bitlis collision zones disappeared in the Oligocene but some sea connection remained opened until the middle–late Miocene (Dercourt et al., 1986, 1993; Popov et al., 2004). North–South shortening due to the progressive collision between the Arabian plate and Eurasia is responsible for the disconnection. The passageway between the Mediterranean and the India Ocean was still open 10 Ma ago and progressively closed until 7–8 Ma (Popov et al., 2004). This disconnection participated in the isolation of the Mediterranean and somehow prepared the Messinian salinity crisis but the final closure occurred later, and this happened in the west. However, the amount of evaporites deposited in the deep basins cannot be so far explained by the amount of salt and gypsum contained in the Mediterranean seawater, thus implying that some connection must have remained open, although considerably reduced (Rouchy, 1982; Rouchy and Saint Martin, 1992; Blanc, 2002; Clauzon et al., 2005).

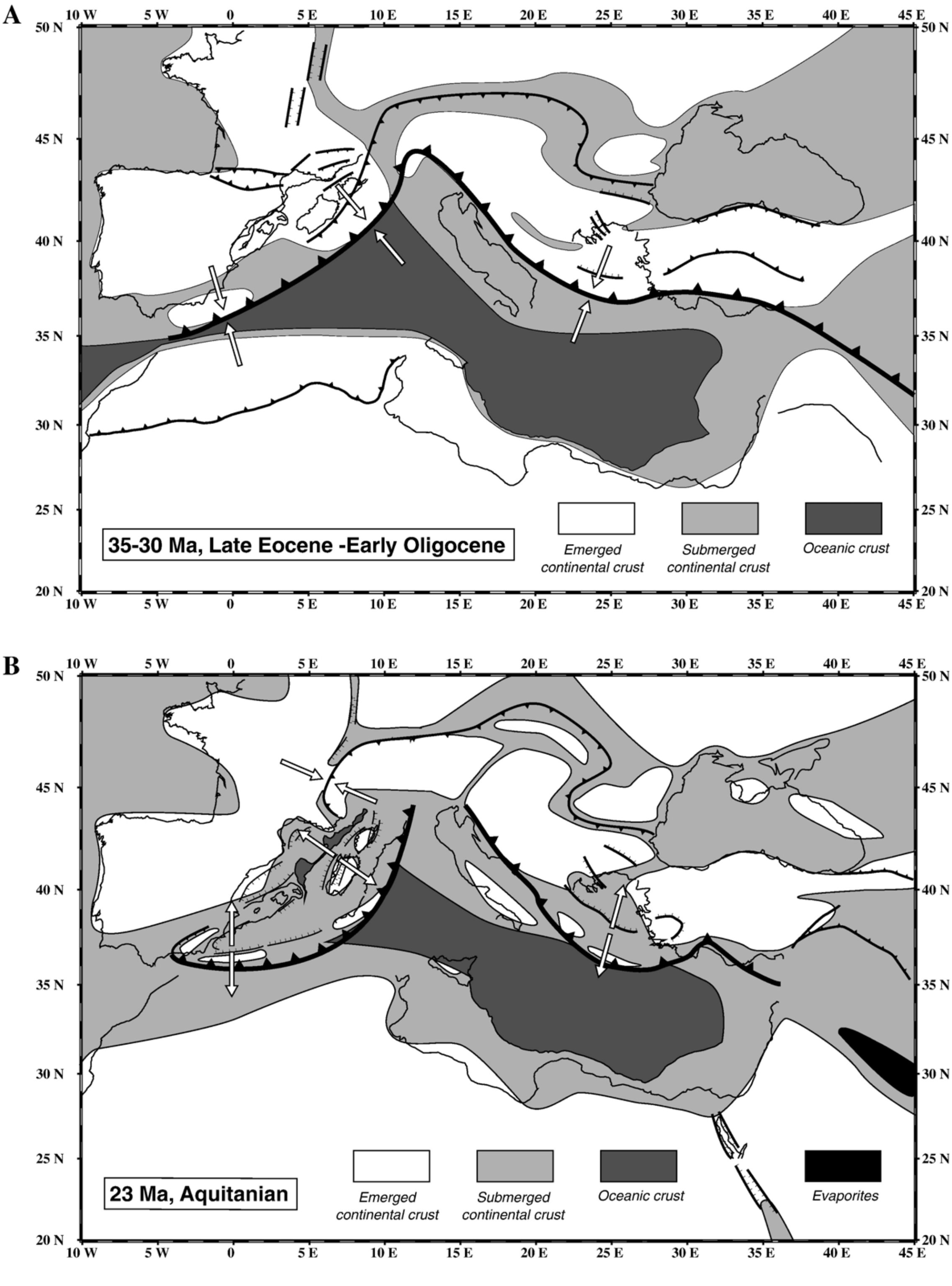
During this evolution, the nature of the deep basins changed through time. In the Eocene and the Oligocene, deep oceanic domains were confined to the old Mesogean oceanic crust trapped south of Apulia. After the Oligocene, new deep basins formed as back-arc basins above the subduction zones and the old oceanic crust was progressively subducted. The formation of the Western Mediterranean Basin (Liguro-Provençal, Tyrrhenian and Alboran) was achieved at the expense of the Mesogean lithosphere and new deep basins replaced the old ones. The Oligocene situation was characterized by one large deep basin. The present shows three deep oceanic basins and the Aegean Sea which is floored by thinned continental crust. The geometry and kinematics of opening of these new basins is a direct consequence of the slab behaviour at depth (Faccenna et al., 2001, 2003, 2004).

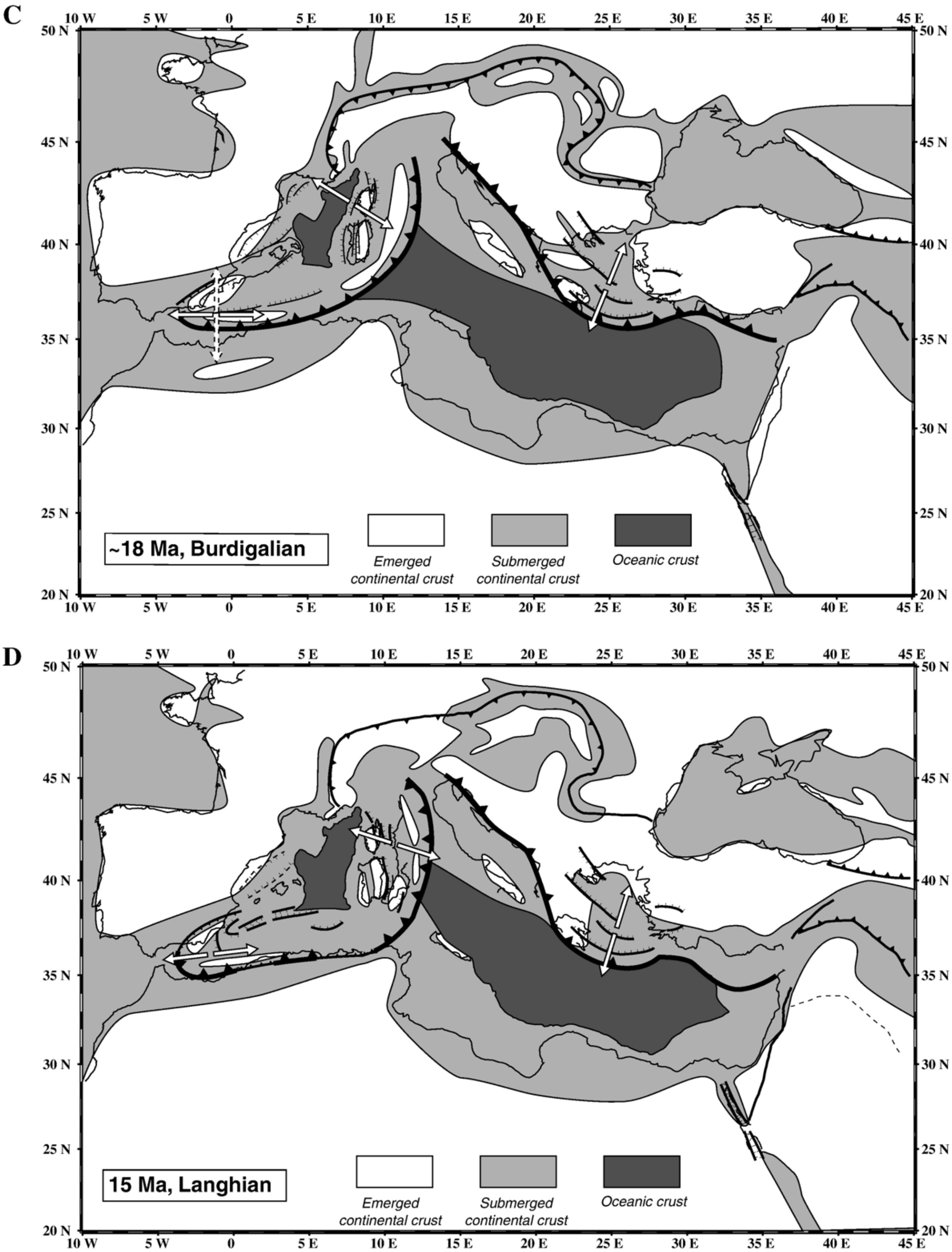
The following reconstructions (Fig. 2) were created from the paleokinematic reconstructions of Jolivet et al. (2003) and Lacombe and Jolivet (2004) which are based on the kinematic parameters of Dewey et al. (1989) and Rosenbaum et al. (2002) for the large plates motions as well as geological data for the internal kinematics of the back-arc basins. We complete the reconstructions with paleo-coastlines taken from various sources (Gvirtzman and Buchbinder, 1978; Rögl and Steininger, 1983; Boccaletti et al., 1990; Clauzon et al., 1990, 1995, in press; Benson et al., 1991; Dercourt et al., 1993; Clauzon, 1996; Rubino and Clauzon, 1996; Rögl, 1999; Görür et al., 2000; Meulenkamp and Sissingh, 2003; Poisson et al., 2003; Popov et al., 2004; Gorini et al., 2005; Lofi et al., 2005; Çagatay et al., this volume). These reconstructions are only intended to show the general evolution of basins and should not be taken as quantitatively precise. For example, the regional consequences of some transgressions or regressions, being forced by tectonics or by glacio-eustasy, have not been considered. If the kinematic parameters are self-consistent and each stage represents a given geodynamic stage significantly different from the preceding one and is thus illustrative of the geodynamics of a quite long period, the coastlines are probably more instantaneous images and thus details are not important. This is why we tried to show the general evolution and why the original maps have been sometimes simplified.

In a first approach the maps show a maximum of extension of the sea in the Burdigalian. Afterward we see a general reduction of the surface occupied by sea

Fig. 2. Paleogeographic and paleotectonic maps of the Mediterranean region based on Dewey et al. (1989), Jolivet et al. (2003) and Lacombe and Jolivet (2004) for plate displacements and internal deformation, and Gvirtzman and Buchbinder (1978), Rögl and Steininger (1983), Boccaletti et al. (1990), Clauzon et al. (1990), Benson et al. (1991), Dercourt et al. (1993), Clauzon et al. (1995), Clauzon (1996), Rubino and Clauzon (1996), Rögl (1999), Görür et al. (2000), Meulenkamp and Sissingh (2003), Poisson et al. (2003), Popov et al. (2004), Clauzon et al. (2005), Çagatay et al. (this volume), Gorini et al. (2005) and Lofi et al. (2005) for the position of coastlines and Rouchy and Caruso (2006-this issue) for the extension of Messinian evaporites. NB: oceanic crust is shown only in the Mediterranean domain, not in the Atlantic Ocean and Black Sea.

and a reduction of the connections between the Mediterranean and the Paratethys. The Messinian event is thus the climax of a process that started long before and culminated when the Gibraltar sill was closed. In the following we present each successive stage from the Eocene to the Present.





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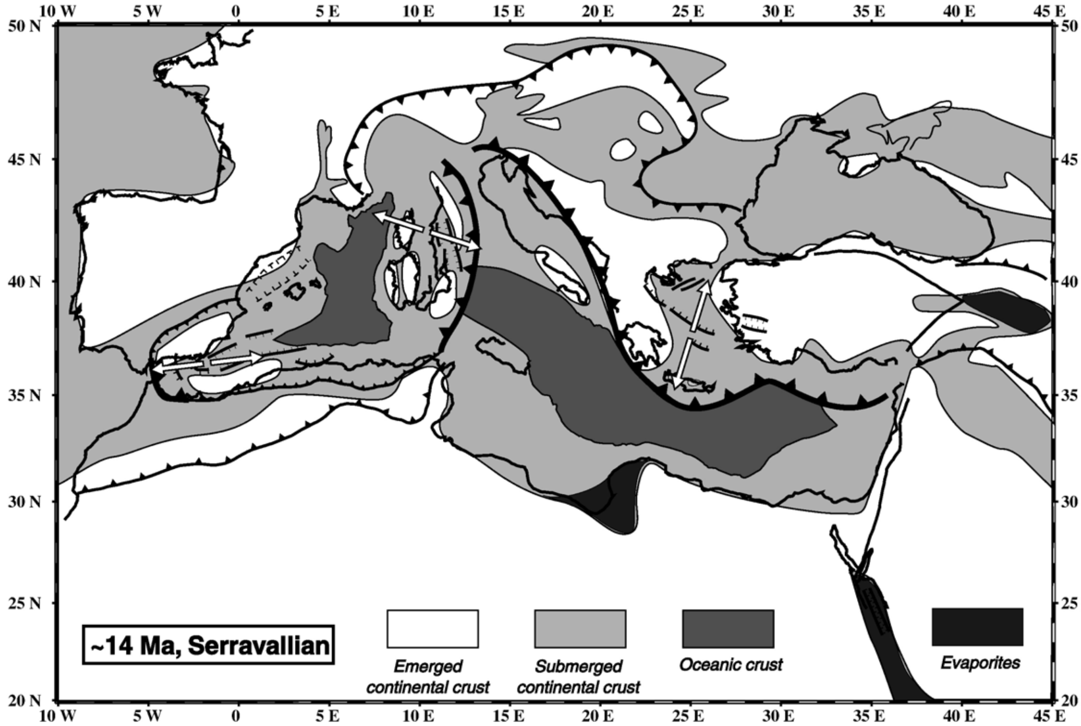

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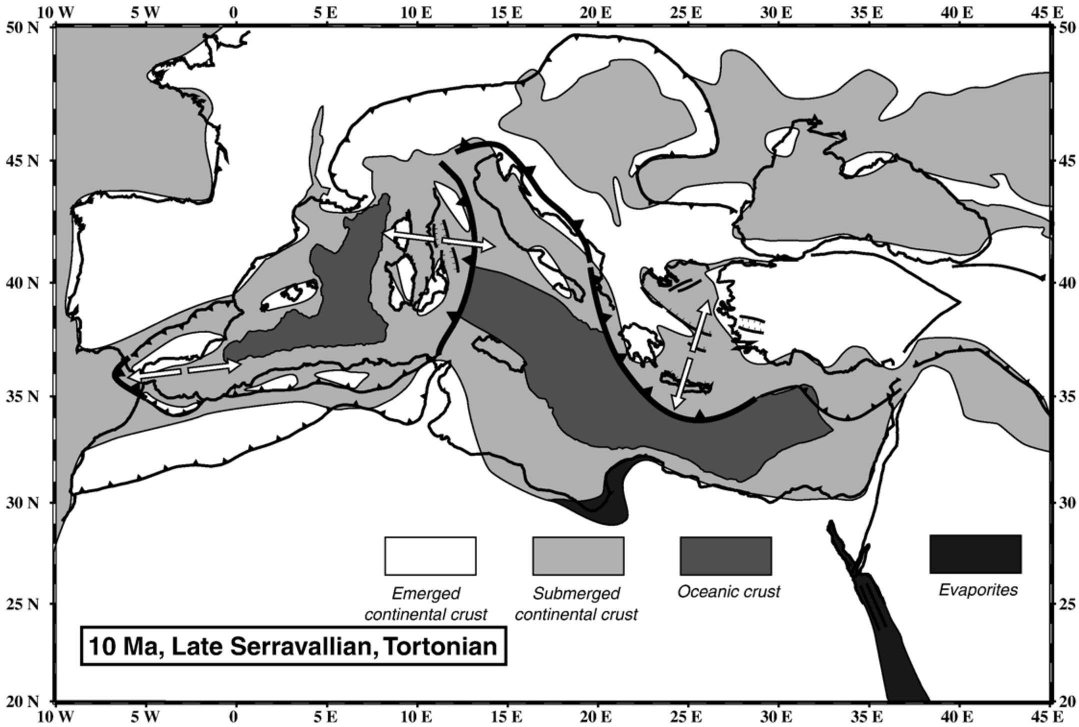

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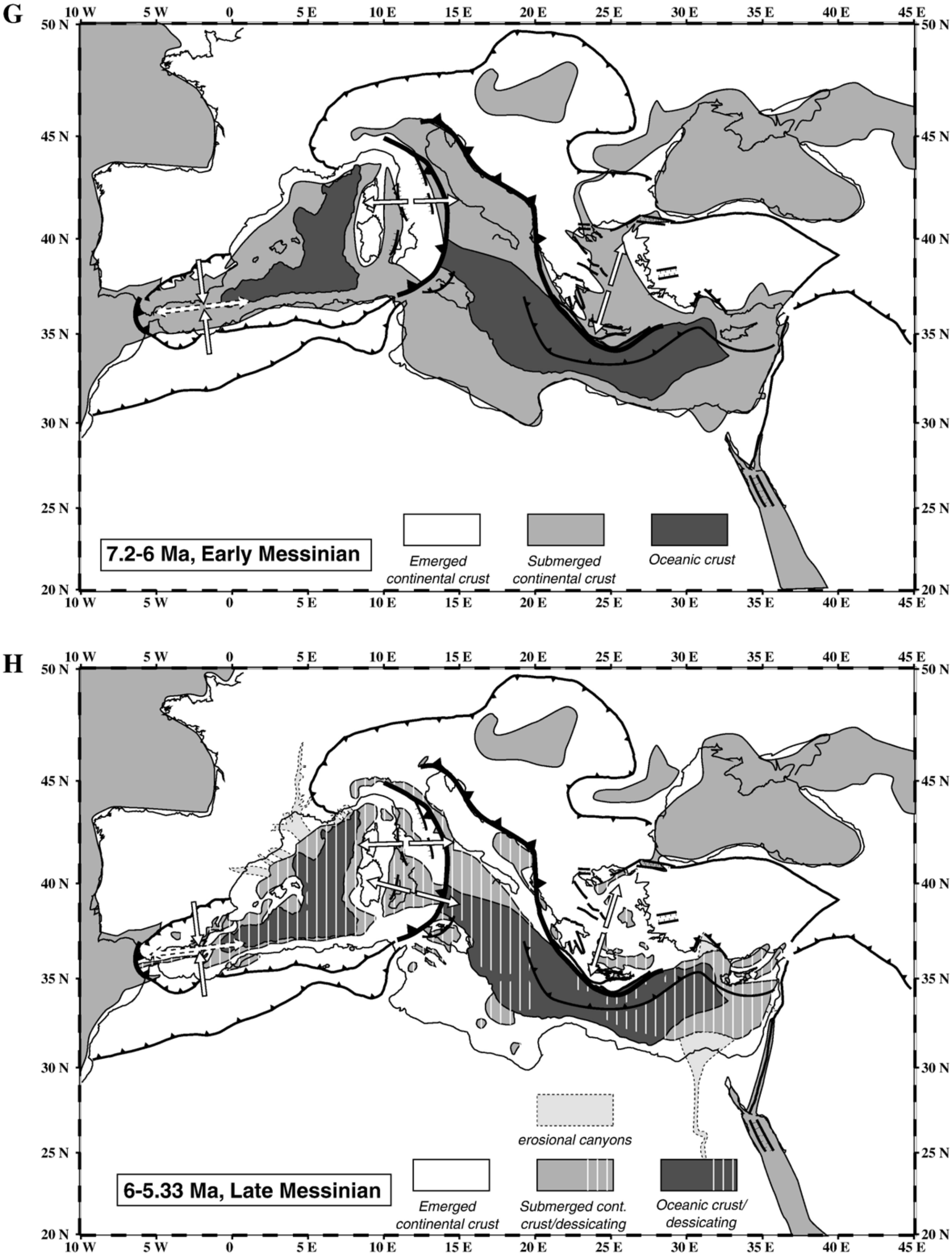

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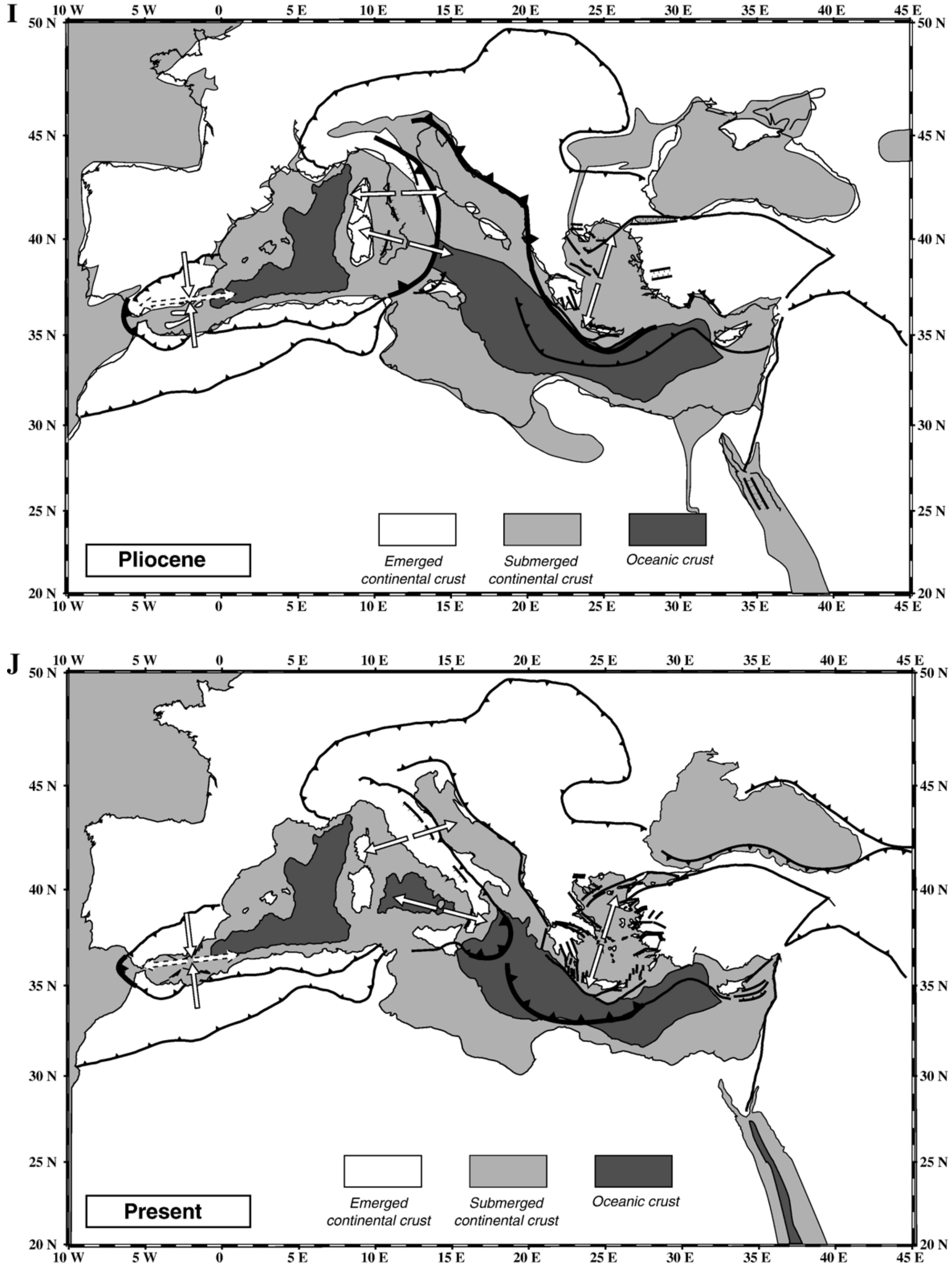

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2.1. Late Eocene – early Oligocene (Fig. 2A)

This period is a turning point in the evolution of the Mediterranean (Le Pichon et al., 1988; Jolivet and Faccenna, 2000). The Mesogean oceanic lithosphere is progressively locked between two collision zones. The first one, in the east, is the collision between the future Arabian plate and Eurasia. The second one is the collision between the Iberian Peninsula and the northern margin of the African plate which occurs progressively from 35 to 30 Ma. Far-field effects of this early event leads to a first intracontinental shortening along the Atlas range quite far inland from the plate boundary (Frizon de Lamotte et al., 2000, 2004). Mountain belts are under construction above the north-dipping subduction of the Mesogean lithosphere. The Betic Cordillera forms by the underthrusting the African margin below the southern margin of Iberia as shown by the occurrence of high-pressure-and-low-temperature metamorphic rocks of Paleocene–Eocene age in the Alboran domain. The subduction of continental fragments is attested by blueschists and eclogites dated from 48 to 30 Ma in the Betic Cordillera (Augier, 2004; Augier et al., 2005a; Platt et al., 2005). Further east, the subduction is well engaged as well, and a calc-alkaline volcanic arc develops in Sardinia and southern Provence. This period sees the end of compressional deformation in Alpine Corsica and Pyrenees–Provence in the internal zones of the orogen (Arthaud and Séguret, 1981; Choukroune et al., 1990; Seranne et al., 1995; Vergés et al., 2002). Subduction of the Apulian lithosphere induces the formation of the Dinarides and Hellenides passing to the Taurides. In the north the Paratethys is continuous from the Alpine–Carpathian basin to the Black Sea area and some connection with the Mediterranean can be observed in the southern part of the Alps or, periodically, in the Aegean region through the Meso-Hellenic trough (Ferrière et al., 2004).

During the time interval the subduction regime changed to extensional with a significant amount of slab retreat from 30 Ma and the Present Mediterranean dynamics starts to develop (Jolivet and Faccenna, 2000). At the same time, the tectonic context changed also around the future Arabian plate as the collision with Eurasia in the north and the slab pull below the Iranian block in the future Zagros range induced the formation of two major rifts, the Aden and Red Sea rifts (Jolivet and Faccenna, 2000; Bellahsen et al., 2003).

2.2. Aquitanian (Fig 2B)

Slab retreat has now been active during ~7 Ma and rifts have formed in the back-arc domains of the Western and Eastern Mediterranean subductions. In the west, the first oceanic crust has formed in the Liguro–Provençal Basin and extension is recorded in Corsica, Sardinia, the Valencia Trough, the Alboran domain and the Aegean Sea (Cerchi and Montadert, 1982; Gautier and Brun, 1994; Geel, 1995; Jolivet et al., 1998; Gautier et al., 1999). Post-orogenic extension leads to the formation of metamorphic core complexes and subsequent exhumation of metamorphic rocks in all these regions (Lister et al., 1984; Platt and Vissers, 1989; García-Dueñas et al., 1992; Jolivet et al., 1998). The oceanic domain of the eastern Mediterranean has started to subduct below Crete and the Aegean region. Compressional deformation and formation of HP–LT metamorphic rocks however still continue near the front of the subduction zones in Crete, Peloponnese and the internal Apennines (Tuscan archipelago) (Jolivet et al., 1996, 1998; Rossetti et al., 1999a, in press; Trotet, 2000). Nappes are emplaced in the frontal zones of the Betics and the Rif around the Gibraltar arc (Morley, 1993; Lonergan and White, 1997; Platt et al., 2003). A marine gulf extends from the Liguro–Provençal basin northward along the front of the Alps but it is not yet connected to the Helvetic molassic basin (Demarcq, 1970). Some connections are opened between the Paratethys and the Mediterranean region through the Aegean Sea and the Meso-Hellenic trough. A seaway is still open with the Indian Ocean. The exact extension of sea in the Aegean region is difficult to determine because Miocene sediments are rare on Aegean islands and little is known about the offshore stratigraphy, such as in the Thermaikos Gulf where deposits from Aquitanian to Serravallian have been recorded (borehole Nireas 1: Driavaliari, 1993).

2.3. Burdigalian (Fig. 2C)

A large part of the Liguro–Provençal basin is opened and extension is active in the future Algerian and Alboran Basins (Chamot-Rooke et al., 1999; Gattacceca, 2000; Mauffret et al., 2004). Metamorphic core complexes continue to form in the Aegean region and in the Alboran domain. In Corsica, extension controls the deposition of small shallow basins (Saint Florent) (Daniel et al., 1996; Jolivet et al., 1998). The marine gulf in the Southern Alps is now connected with the forearc basin of the central and eastern Alps and further east with the Carpathians and the Paratethys. The thrust front progresses westward within the molasse basin. The front of subduction progressively advances toward the Mesogean oceanic crust in the Central and Eastern Mediterranean and toward the continental domain of Northern Africa in the Western Mediterranean. The Apulian continental lithosphere is subducting below the advancing Northern Apennines accretionary wedge while oceanic lithosphere subducts below the Southern Apennines (Malinverno and Ryan, 1986; Lavecchia, 1988). The Paratethys is connected with the Mediterranean in the Alps and through the Aegean region and seaways are opened with the Indian Ocean and with the newly opened Red Sea. The connection with the Atlantic Ocean is widely opened on either sides of the Betic-Rif orogen (Popov et al., 2004).

2.4. Langhian – Serravallian (Fig. 2D and E)

Back-arc extension in the Western Mediterranean is under progress and extension tends to localize in the future Algerian Basin where a new oceanic crust will form (Mauffret et al., 2004). Extension progressively stops in the Liguro–Provençal Basin and migrates east of Corsica and Sardinia in the back-arc domain of the Apennines (Bartole, 1995; Jolivet et al., 1998; Mauffret and Contrucci, 1999). The oceanic domain that is subducting below the southern Apennines is getting narrower with time. The length of the Western Mediterranean subduction zone is increasing with time and the slab probably has been ruptured laterally in several pieces already at that time (Faccenna et al., 2004; Spakman and Wortel, 2004). The timing of slab tearing will be discussed longer later in this paper but it is sure that in the Langhian the direction of slab retreat below the Betics is westward (Lonergan and White, 1997; Augier et al., 2005b). This time-window is characterized by significant fluctuations in temperature (Miller et al., 1991; Jiménez-Moreno et al., 2005) and resulting glacio-eustatic changes in global sea level (Haq et al., 1987). Therefore, the geographic transformations in the Mediterranean realm are the consequence of both global eustasy and regional tectonics. The Paratethys is no longer continuous from the Alps to the Carpathians and a disconnection is noted between the Black Sea and the Carpathian–Pannonian Basin in the Langhian but the connection is established again in the Serravallian. A marine gulf is observed in the forearc region of the Southern Alps connected to the Liguro–Provençal Basin (Demarcq, 1970; Rögl and Steininger, 1983). Some narrow connections still exist with the Mediterranean across the Aegean or in the eastern Alps. In the Serravallian, the connections with the Indian Ocean closed and evaporitic deposits formed in the Red Sea, the Mesopotamian basin and locally along the northern margin of Africa in the Sirt basin (Gvirtzman and Buchbinder, 1978; Rögl and Steininger, 1983; Rouchy, 1986; Orszag-Sperber et al., 1998).

2.5. Late Serravallian–Tortonian (Fig. 2F)

The opening of the Western Mediterranean basin is now complete, except for the southern Tyrrhenian Sea which is still narrow. Extension is still active in the internal Apennines and it migrates eastward. In the Alboran region, the Tortonian is a turning point. The early Tortonian is characterized by the prosecution of extensional tectonics and basin subsidence (Montenat and Ott d´Estevou, 1990; Augier et al., submitted for publication). Local restriction of water exchanges led to the first deposition of evaporites in the eastern Betics in the Lorca and Fortuna basins (Rouchy et al., 1998; Dinarès-Turell et al., 1999; Krijgsman et al., 2000). From the late Tortonian, an uplift of the basin basement is recorded and extensional basins are partly inverted in the internal Betics (Weijermars et al., 1985; Vissers et al., 1995). N–S compression is active from ~8 Ma in this region (Augier et al., submitted for publication). Local restriction of water exchanges led to the first deposition of evaporites in the eastern Betics in the Lorca and Fortuna basins in the late Tortonian to early Messinian (Rouchy et al., 1998; Dinarès-Turell et al., 1999; Krijgsman et al., 2000). N–S extension still goes on in the Aegean region and the Aegean Sea and Cretan Sea widen. A narrow seaway is still observed across the Arabia–Eurasia collision zone. A narrow basin is observed along the trace of the North Anatolian Fault in the Northern Aegean (Çagatay et al., this volume). Added to the observation of a short compressional event in the eastern Aegean, some 9 Ma ago (Angelier, 1979; Weidmann et al., 1984; Ring et al., 1999), this suggests that the North Anatolian Fault was actually entering the Aegean domain for the first time. Otherwise the connection with the Paratethys seems closed or was only temporarily opened in the Tortonian, as attested by sporadic brief incursions of Mediterranean nannoplankton into the Eastern Paratethys (Marunteanu and Papaianopol, 1998). Evaporites form along the northern margin of Africa and in the Red Sea (El-Heiny and Martini, 1981; Rögl and Steininger, 1983; Rouchy, 1986; Orszag-Sperber et al., 1998).

2.6. Messinian (Fig. 2G and H)

The Tyrrhenian Sea is opening but is still quite narrow (Sartori et al., 1987; Kastens et al., 1988; Mascle et al., 1988). Extension is active in the internal Apennines and northern Tyrrhenian Sea (Jolivet et al., 1998; Rossetti et al., 1999b). The Betics and the Rif are subjected to a N–S compression and the Betic slab is still subducting eastward (Comas et al., 1999; Gutscher et al., 2002). Back-arc extension in the Alboran domain seems to have mostly stopped (Krijgsman and Garcés, 2004). The frontal thrust of the southern Apennines reaches Sicily and the Messinian deposits are progressively incorporated in the orogenic wedge (Butler et al., 1995, 1999). The connection with the Atlantic will be partly closed during the Early Messinian. In the Betics, the transition between marine and continental sedimentation in the Guadix Basin occurred at about 7.3 Ma ago (Soria et al., 1999), at about 6.8 Ma in the Guadalhorce corridor (Martín et al., 2000), recording the final closure between the Atlantic and Mediterranean in the Betic corridor. In Morocco, the Rifian corridor, linking the Mediterranean Taza-Guercif Basin to the Atlantic Ocean, was initiated at ~8 Ma and deep marine sediments were deposited. At 7.4–7.2 Ma, near the Tortonian–Messinian boundary, a fast shallowing-up is observed (closure of the Taounate corridor: Barhoun, 2000) and the Guercif area was finally emergent before 6.0 Ma whereas very reduced relationships existed through the Touahar pass up to 5.65–5.4 Ma (Warny et al., 2003). Some connection with oceanic water input must remain open during the whole Messinian to explain the amount of evaporites deposited in the deep basins. This situation will lead in a very short time laps to a partial dessication of the Mediterranean and deposition of evaporites in the remnant sea. The water level drop induced a very active erosion of the margins and formation of deep fluvial incisions far inland (Clauzon et al., 1990) such as the Rhone and Nile canyons. The Aegean region is under extension and a thick accretionary wedge forms south of Crete: the Mediterranean ridge (Lallemant et al., 1994; Chaumillon et al., 1996; Le Pichon et al., 2002). The Marmara Sea is forming as a pull-apart basin along the North Anatolian Fault (Armijo et al., 1999, 2002). Direct connection with the Indian Ocean is definitively closed and the question of the connection with the Red Sea might be discussed but the presence of Messinian evaporites in the Red Sea (Stoffers and Ross, 1974) suggests that the absolute sea level was very low and thus makes the possibility of a passage to the Mediterranean unlikely (Rouchy, 1986). Temporary connections with the Eastern Paratethys are proposed through the northern Aegean (Clauzon et al., 2005). Resulting from the evidence of two successive influxes of Mediterranean nannoplankton into the Eastern Paratethys (Semenenko and Olejnik, 1995; Marunteanu and Papaianopol, 1998; Snel et al., in press), a new concept of the Lago Mare facies is proposed by Clauzon et al. (2005) as exchanges between the two realms during two brief high sea-level phases, just before and just after the entire desiccation of the Mediterranean.

2.7. Pliocene to Present (Fig. 2I and J)

The geodynamic situation in the early Pliocene is the same but the re-opening of connections through the Gibraltar gate leads to a refill of the Mediterranean (Hsü et al., 1973b) in a very short time whose duration is difficult to estimate. From a hydrological modelling, Blanc (2002) considered that the basin may have been completely refilled in a few tens of years. Seawater of the long sea-level high-stand (starting in the Late Messinian, Haq et al., 1987) invades the canyons deeply dissected by the Messinian regressive erosion (Clauzon et al., 1990, 1995; Clauzon, 1996). The connections with the Paratethys were opened through the Dardanelles strait only very recently in the Quaternary (Çagatay et al., this volume). The major connection with the Paratethys is proposed to have been across the Balkans (Clauzon et al., 2005). Connections with the Red Sea are not documented. The recent evolution sees the fast opening of the southern Tyrrhenian Sea and thus the widening of the Western Mediterranean deep basin and the construction of the thick Mediterranean Ridge accretionary prism. The Anatolian plate moves fast toward the west along the North Anatolian Fault (McClusky et al., 2000; Le Pichon et al., 2001). In the Alps it seems that crustal thickening has now stopped and gravitational spreading has started (Delacou et al., 2004). In the Aegean region, extension is still active at about the same rate as in the Miocene but the deformation is more localized in western Turkey and around the Volos and Corinth Rifts as well as around Crete and in the Peloponese (Jolivet, 2001; Kreemer and Chamot-Rooke, 2004; Kreemer et al., 2004). No major tectonic reorganisation is observed between the Messinian and the Pliocene and there is no need for it to reopen Gibraltar 's gates. From a morphological study of the offshore area it was proposed that the Gibraltar strait is in fact a large Messinian canyon flowing toward the east which progressively eroded backward and reopened the connection in a catastrophic way (Blanc, 2002).

3. Closing the gates: tectonic and geodynamic evolution of the peri-Alboran region

Because of a narrower space available for slab retreat, the peri-Alboran domain evolved quite differently from the Tyrrhenian or Aegean regions and N-S compression has been active since the late Miocene (~8 Ma). We now describe the magmato-tectonic evolution of the Betic Cordillera and discuss the causes of the closure of the Mediterranean–Atlantic connections.

The magmatic evolution is described in details by Duggen et al. (2003, 2004). Cenozoic magmatic rocks are found onshore and offshore along an oblique NE–SW trending line that crosses the Alboran Sea as well as in the western Betics in the form of a large basaltic dyke swarm around Malaga. As shown in Fig. 3, the Malaga basalts were emplaced with ages ranging from 38 to 17 Ma and their chemistry suggests a normal subduction process of sediments or hydrous oceanic crust. Around 20 Ma, cordierite-bearing igneous rocks and leucogranites suggest an episode of crustal anatexy. Early Miocene ages for some of the Malaga dykes are probably due to a thermal resetting of older dykes. The middle and late Miocene are characterized by calc-alkaline and tholeiitic rocks in the whole region. They are interpreted as resulting from the subduction of an hydrated oceanic crust or sediments. The end of the late Miocene and the Pliocene shows the transition toward shoshonites and alkali basalts that is interpreted by Duggen et al. (2003, 2004) as an indication of convective removal of the lower lithosphere from beneath the Alboran region. This evolution of the composition of magmatic rocks led Duggen et al. (2003, 2004) to propose a scenario favouring convective removal near the end of westward rollback of the slab as the cause of the Messinian salinity crisis. This interpretation conflicts with the timing of extension and/or convective removal proposed earlier (Platt and Vissers, 1989; Lonergan and White, 1997; Platt et al., 2003; Faccenna et al., 2004).

The tectonic evolution of the internal Betics (Fig. 4) indeed favours, as explained below, a scenario of slab rollback (Lonergan and White, 1997; Faccenna et al., 2004; Spakman and Wortel, 2004); however, this evolution seems to have ended before the Messinian salinity crisis. In the following we summarize the main observations and propose a kinematic scenario involving southward-then-westward rollback.

3.1. Onshore basins

The onshore Neogene basins of the Alboran domain developed in three stages. The oldest deposits date back to the Burdigalian (Comas et al., 1992, 1999). They are restricted to small basins controlled by an episode of N–S extension and the development of a first set of low-angle normal faults (Crespo Blanc, 1995). The second episode started in the Serravallian some 12 Ma ago [the Filabres System (Martínez-Martínez et al., 2002)]. It formed the main basins seen today in the eastern Betics (Montenat and Ott d´Estevou, 1990; Crespo Blanc, 1995; Augier et al., submitted for publication). These basins (Huercal-Overa and Tabernas basins, Fig. 4) are extensional and the main direction of syn-sedimentary extension is again N–S (Montenat and Ott d´Estevou, 1990; Sanz de Galdeano and Vera, 1992; Mora, 1993; Vissers et al., 1995; Augier et al., submitted for publication). This episode occurred contemporary with the late exhumation of the main metamorphic domes, the Sierra Nevada – Sierra de los Filabres and Sierra Alhamilla (Johnson et al., 1997; Martínez-Martínez et al., 2002; Augier et al., 2005a,b). The third episode (from ~8 Ma to the present) corresponds to the uplift of the basin floor and tectonic inversion associated with N–S compression in the east while the Granada Basin in the west still records NE–SW extension (Weijermars et al., 1985; Vissers et al., 1995; Galindo-Zaldivar et al., 1999).

3.2. Metamorphic core complexes

The exhumed metamorphic rocks record the evolution of the deep crust. The Alboran basement is made of three main groups of nappes (Blumenthal, 1927; Van Bemmelen, 1927; Egeler and Simon, 1969). From top to bottom (Fig. 4): (1) the Malaguides (or Ghomarides in the Rif) are made of a Palaeozoic basement which did not record any Alpine high-pressure and low-temperature metamorphism. They are separated from the (2) Alpujarrides (Sebtides in the Rif) by a large-scale detachment, the Malaguide–Alpujarride Contact (MAC) (Lonergan and Platt, 1995). The Alpujarrides are made of a stack of nappes that widely contain HP–LT parageneses indicating pressure conditions as high as 12 kbar for rather low temperatures around 450 °C (Goffé et al., 1989; Azañon and Goffé, 1997). This HP event occurred during the Eocene (Platt et al., 2005). A second metamorphic event occurred some 20 Ma ago. It corresponds to the quite sudden thermal event that reset most radiometric systems (Zeck et al., 1989, 1992; Monié et al., 1991, 1994; Platt and England, 1994; Platt and Whitehouse, 1999). High temperature parageneses overprint the HP–LT ones in some of the Alpujarride units. The deformation within the Alpujarrides is characterized by a N–S to NE–SW direction of stretching and a consistent top-to-the-

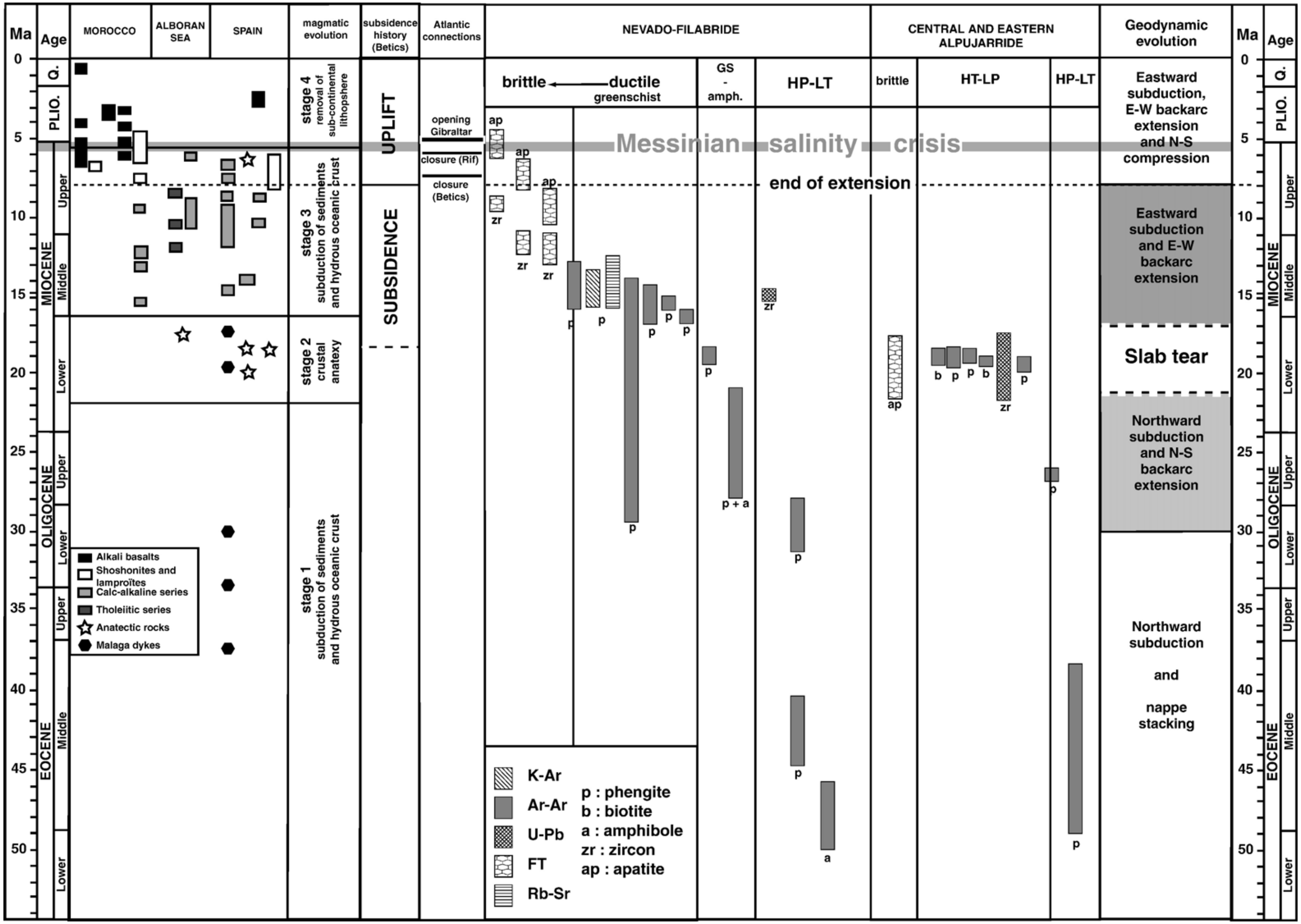

Fig. 3. Timing of metamorphic, magmatic sedimentary events in the Mediterranean region. Data after Duggen et al. (2003, 2004), Augier et al. (2005a), Platt et al. (2005) and Negro et al. (2005).

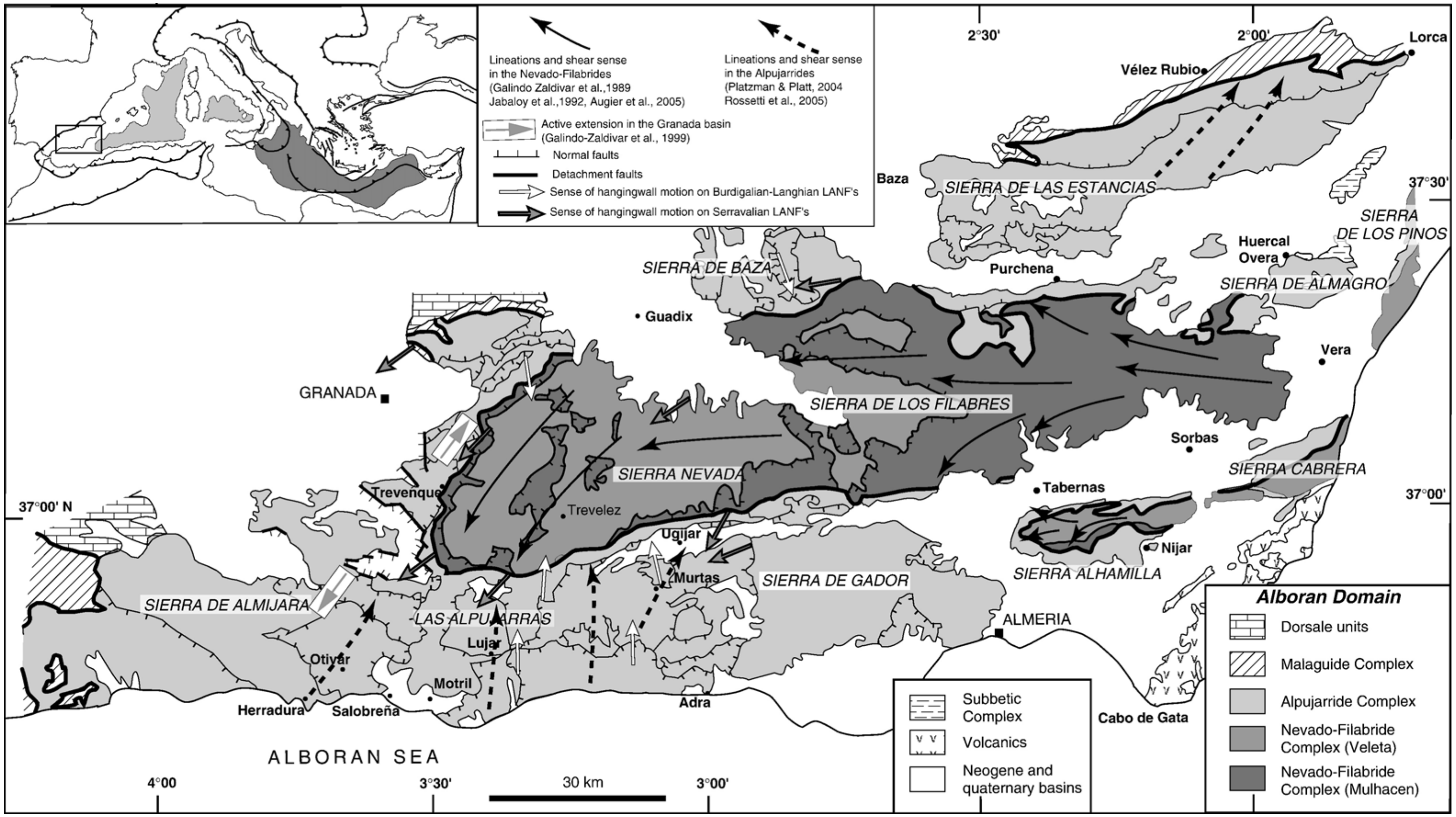

Fig. 4. Tectonic map of the eastern part of the internal parts of the Betic Cordillera showing the main detachment and structural units as well the directions and sense of shear in the exhumed metamorphic complexes and direction of extension in the sedimentary basins (García-Dueñas et al., 1992; Mora, 1993; Crespo Blanc et al., 1994; Crespo Blanc, 1995; Martínez-Martinez and Azañon, 1997, 2002; Galindo-Zaldivar et al., 1999; Augier et al., 2005b; Augier et al., submitted for publication; Rossetti et al., 2005; Platzman and Platt, 2004).

north or -northeast sense of shear and an evolution from ductile to brittle during exhumation (García-Dueñas et al., 1992; Crespo Blanc et al., 1994; Rossetti et al., 2005; Platzman and Platt, 2004). The first set of Burdigalian–Langhian low-angle normal faults developed at the end of this episode (García-Dueñas et al., 1992; Crespo Blanc et al., 1994; Crespo Blanc, 1995). A second set of west-dipping low-angle normal faults developed in the Serravallian. The Alpujarrides are separated from the (3) underlying Nevado–Filabrides by a second detachment, the Filabres detachment (Martínez-Martínez et al., 1997; Martínez-Martinez and Azañon, 1997, 2002; Martínez-Martínez et al., 2002). The Nevado–Filabrides have no equivalent in the Rif. They are made of two main units. The upper unit (Mulhacen) reached the conditions of the eclogites facies (20 kbar for some 600 °C) some 48 Ma ago. It rests above the Veleta unit, the deepest part of the Alboran domain, where maximum pressure conditions are of the order of 12–13 kbar (Booth-Rea et al., 2003) confirming that the contact with the Mulhacen unit was originally a thrust (Gonzales Cassado et al., 1995; Augier et al., 2005a). The oldest ages for the high-pressure event in the Veleta are around 30 Ma (Augier et al., 2005a).

The deformation records mainly the exhumation history. Some NW-verging folds, preserved from later deformation may correspond to the crustal thickening and nappe-stacking episode (Langenberg, 1972; Vissers, 1981). The main foliation formed between 30 and 20 Ma while a gradient of retrograde shearing deformation is observed toward the top of the Nevado–Filabrides. The most obvious stretching is E–W associated with a very consistent top-to-the-west shear sense. This episode furthermore shows an evolution from ductile to brittle from 20 Ma to ~14 Ma (Augier et al., 2005a). The greenschist-facies stretching directions show diverging patterns in both the Sierra de los Filabres and Sierra Alhamilla domes, suggesting that the domes were already present at that time (Augier et al., 2005a,b). The deformation near the Filabres detachment evolves with time with a progressive rotation of the direction of extension from E–W ductile stretching to N–S when reaching the brittle field. This late N–S extension controls the deposition of the Serravallian–Tortonian sediments in overlying basins. Fission-track ages along the strike of the Sierra de los Filabres and Sierra Nevada dome suggest a progressive exhumation of the Nevado–Filabrides from east to west, compatible with the observed ductile shear sense from 12 to 9 Ma, that is during the formation of the basins (Johnson et al., 1997). The N–S extensional regime is thus restricted to the basins and did not affect the whole crust. It has been interpreted as a result of doming (Augier et al., 2005b).

3.3. A tectonic scenario

The tectonic evolution of the internal Betics can thus be described in five main episodes (Fig. 5):

(1) Crustal thickening and nappe stacking around 40–50 Ma (Eocene). HP–LT metamorphic conditions prevailed during this early stage. This episode is associated with ~N–S compression and northward subduction of the African plate.

(2) From 30 to ~20 Ma (Oligocene and early Miocene) the first extension is recorded in the Alboran Sea. N–S extension leads to the formation of low-angle normal faults in the Alpujarrides that follows a N–S ductile stretching. This episode can be related to a first southward slab retreat episode. In the external zones the thrust front is advancing.

(3) 20 Ma ago (early Miocene), a sudden event induces a widespread thermal overprint and local crustal anatexy (Platt and Whitehouse, 1999; Zeck and Whitehouse, 1999). During this period, N–S extension is still active in the Alpujarrides and E–W extension in the underlying Nevado–Filabrides. It can be interpreted as the transition from southward to westward slab retreat. Tearing of the slab, as suggested by tomographic images (Spakman and Wortel, 2004) would explain the sudden increase of temperature.

(4) After this kinematic change, westward slab retreat is active until ~8 Ma (upper Miocene, Late Tortonian) and the basins form, partly controlled by the exhumation of large metamorphic domes. Paleomagnetic data show that no rotation occurred after the Late Tortonian in the Betics suggesting that deformation related to slab retreat had indeed mostly stopped by that time (Krijgsman and Garcés, 2004).

(5) From 8 Ma to the Present, tectonic inversion and basin uplift is active in the eastern Betics, while NE–SW extension is recorded in the Granada Basin.

The Messinian salinity crisis thus occurred during the late compressional episode after the cessation of westward slab retreat. We can now discuss the interpretation of Duggen et al (2003, 2004).

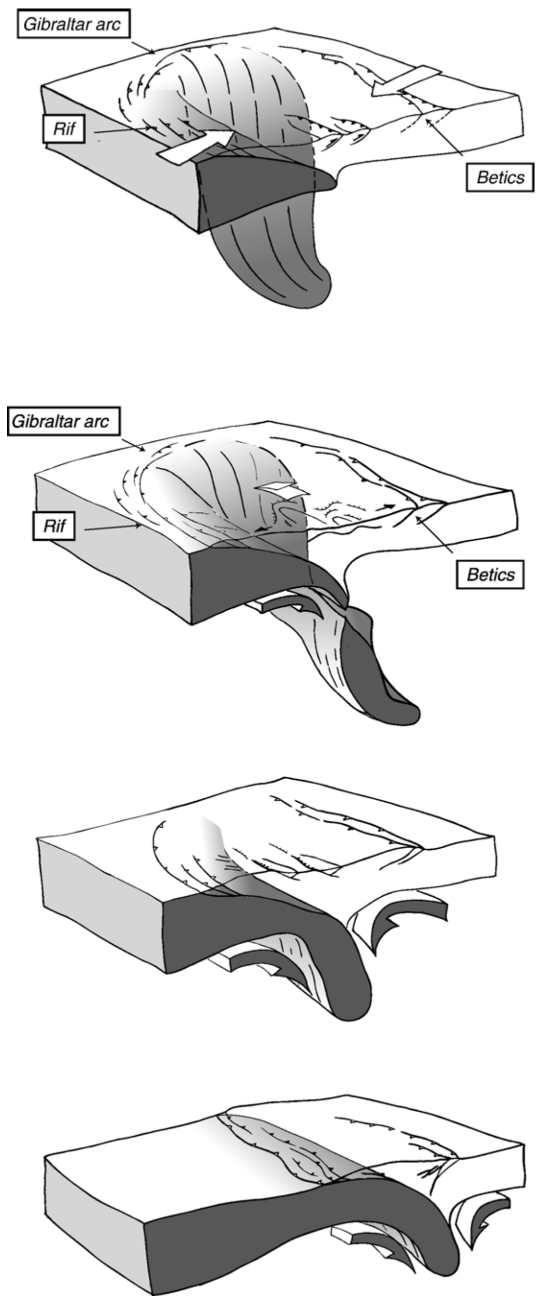

Fig. 5. Schematic evolution of the Gibraltar arc from the Eocene to the late Miocene.

In their model, the closure of the Betic and Rifian corridors are due to the regional uplift consequent to the lithospheric removal and asthenospheric upwelling during slab retreat. This model is different from the more classical model of convective removal (Platt and Vissers, 1989) because of a more progressive removal of the upper mantle from below the chain but the effect in terms of vertical displacement is globally the same. A numerical model shows that a 1 km uplift can be predicted that would have closed the connections between the Atlantic and Mediterranean waters. Actually even a much more limited uplift would be sufficient to close these connections (100–200 m). The large uplift postulated in that model would have considerably increased the potential energy stored in the Alboran lithosphere and would thus have favored extension instead of compression. The geological history of the Betics shows that extension had stopped 2 Ma before the salinity crisis happened. We then conclude that the closure of the Rifian and Betic corridors is a consequence of renewed compression after gravitational potential energy has been released, rather than slab retreat. Slab retreat and crustal collapse have instead favoured the formation of a deep basin by crustal thinning throughout the Miocene and thus favoured connections between the two water masses, delaying the salinity crisis. A simple continuous shortening from the Eocene to the Present would have closed the connections much earlier and the salinity crisis would have happened long before the Messinian. The effect of this late compressional episode could be two-folds: formation of compressional structures as recorded in the internal Betics (amplification of the metamorphic domes and formation of reverse faults during the Messinian) and regional buckling that could have induced a more regional uplift, thus closing the corridors.

Subsidence history in the Alboran region shows a variety of behaviours. While the onshore region shows a fast uplift in the Early Messinian, as attested by the basins of the eastern Betics or by the drill holes in the Granada Basin, the offshore domain does not show this obvious uplift at the same period (Comas et al., 1999; Rodriguez-Fernandez et al., 1999). A regional uplift due to convective removal should be felt over a large domain and especially offshore where the lithosphere is the thinnest. A regional buckling would have instead amplified the existing topography and uplifted already elevated regions.

4. Communications between the Eastern and Western Mediterranean

One of the important questions about the Messinian salinity crisis is the existence of two separate basins, the Western and Eastern Mediterranean which communicate mainly through the channel between Sicily and Tunisia. The presence of a shallow sill there that would separate two basins with independent evolutions in terms of rainfall, runoff and evaporation has consequences on the crisis scenario (Blanc, 2000). The paleogeographic maps suggest the establishment of a narrow strait between the Tortonian and the early Messinian. Later the main event is the opening of the southern Tyrrhenian Sea during the Pliocene. The two main basins have been present since the Burdigalian, the western basin growing progressively with time at the expense of the eastern basin. Except for the southern Tyrrhenian Sea the Messinian configuration was quite similar to the present one (Blanc, 2000). But we have no precise idea of the water depth and its evolution through time in the Sicily–Tunisia channel. The eastern basin, floored by a Mesozoic lithosphere, has most likely been always deeper than the western one with its Cenozoic lithosphere, and an abrupt deepening is probable across the channel. Because of the progressive rollback of the Apennine–Maghrebide subduction and the compressional deformation at the front (Rosenbaum and Lister, 2004), the sill can only be shallower at present than it was in Messinian times but it is difficult to be more conclusive.

5. Connections with the Paratethys and formation of the North Anatolian Fault

A connection between the Aegean Sea and the Paratethys was present episodically between the Balkans and Anatolia. This connection was present in the Aquitanian and Burdigalian; it was severed in the Langhian and re-established in the early Serravallian. In the late Serravallian and Tortonian no clear connection existed but a narrow gulf extended the Aegean Sea inside Anatolia parallel to the present trace of the North Anatolian Fault. In the early Messinian, a connection was established through the Balkans. After the Messinian salinity crisis and the Pliocene reflooding this connection was still present until recent times. How was this connection through the Balkans first established is not clear.

The migration of the North Anatolian Fault in the Aegean domain is classically dated to the Messinian–Pliocene boundary and a possible interaction with the end of the Messinian salinity crisis should be discussed. The age of this event is debated and ages from 16 Ma to 3 Ma have been proposed in the literature (see a review in Bozkurt, 2001). Extrapolating recent rates of motion (McClusky et al., 2000) to the early Pliocene gives a total lateral displacement of 75–125 km that is in good agreement with the geological estimate (Bozkurt, 2001) suggesting that most of the displacement occurred from ~5 Ma to the Present. However, the possibility that the fault was already present in the Dardanelles region earlier, with however a slower rate, must be considered. Armijo et al. (1999) used the unconformity between the Alçitepe (which they attributed to the Pliocene) and the Kirazli (Messinian) formations to date the formation of a broad anticline offset by the North Anatolian fault on both sides of the Dardanelles Strait. Recent data on these formations suggest that they could be older than postulated by Armijo et al. (Çagatay et al., this volume): the Alçitepe formation would be Messinian and the Kirazli formation early to mid Miocene. An additional important observation is that a marine incursion is recognized during the Late Serravallian along the trace of the western part of the North Anatolian Fault (Çagatay et al., this volume) once again suggesting an earlier formation of the North Anatolian fault in this region. Furthermore, a short compressional episode is recorded in the eastern part of the Aegean Sea in the lacustrine deposits of the island of Samos some 9 Ma ago (Angelier, 1979; Weidmann et al., 1984). This compressional episode could be the result of the propagation of the North Anatolian Fault following a mechanism proposed for the more recent evolution of the Aegean region (Armijo et al., 2003). On either sides of the tip of a propagating shear crack, two lobes of extension and compression form. This compressional deformation would be only short lived because of the regional extension prevailing in the Aegean domain. It is thus possible that the North Anatolia Fault reached the eastern Aegean region earlier than classically postulated. The Marmara Sea formed progressively as a pull-apart basin from the Messinian to the Present. A connection between the Black Sea and the Mediterranean was opened only in the latest Pliocene in the area of the Bosphorus (Görür et al., 2000) and some connections have existed before temporarily either through the Balkans or across the Marmara Sea (Marunteanu and Papaianopol, 1998; Gillet, 2004). From new chronological investigations in the Carpathian foredeep, Vasiliev et al. (2004, in press) concludes to the absence of relations between the Mediterranean and the Paratethys during the late Messinian. The propagation of the North Anatolian Fault in the Aegean domain is thus not strictly contemporaneous with the end of the Messinian salinity crisis, whatever the true age of propagation of the North Anatolian Fault in the Aegean region (~5 Ma or earlier). It is possible that fault propagated early in the Aegean region but started its fast motion later around the Messinian–Pliocene boundary. Depending upon the engine controlling the initiation of the North Anatolian Fault there could be a link or not with the Messinian salinity crisis. For instance if the model recently proposed of a link between slab detachment below eastern Turkey and the formation of the North Anatolian Fault (Faccenna et al., 2006) holds, the consequences could be more important than simply kinematic ones (vertical motion). This question thus remains open.

6. Conclusion

A comparison between the geodynamic and palaeogeographic evolution of the Mediterranean region from the Oligocene to the Present shows that the isolation of the Mediterranean region is a progressive process that started some 30–35 Ma ago. The consequent change in subduction regime with predominant slab retreat has considerably changed the shape of the basin. The large deep basin in the Oligocene is now confined in the Eastern Mediterranean south of Crete. In the meantime, several large deep basins formed in the Western Mediterranean by the rollback of the Apennines–Calabrian–Betic slab progressively separating the Eastern and Western Mediterranean domains. The acme of the restriction that resulted in the sudden Messinian salinity crisis is a consequence of this evolution but it has probably been delayed by the slab retreat and subsequent back-arc extension.

Slab retreat, crustal collapse and thinning shaped the Alboran Sea and Gibraltar arc. The tectonic timing shows that the Messinian salinity crisis occurred after the end of extension and the beginning of a new stage of N–S compression. The crisis thus did not result from the slab retreat process that has instead probably favoured the connections between the Atlantic and Mediterranean waters during most of the Miocene but rather from a renewal of N–S compression in the Alboran domain after the gravitational potential energy stored during slab retreat and delamination has been released. A simple compression and crustal thickening would have instead closed the connections and provoked an earlier salinity crisis. The slab retreat thus probably has delayed significantly the salinity crisis that is a logical consequence of the progressive isolation of the Mediterranean Basin during convergence.

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