18.4.1 Evolution of the Western Mediterranean

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- Introduction

The western Mediterranean is the younger part of the Mediterranean, being a basin formed from late Oligocene to present. The western Mediterranean consists of a series of sub-basins such as the Alboran, Valencia, Provençal, Algerian and Tyrrhenian seas (Fig. 18.4.1.1). These basins have in general a triangular shape and they generally rejuvenate moving from west to east. They are partly floored by oceanic crust (Provençal and Algerian basins, and two smaller areas in the Tyrrhenian Sea). The remaining submarine part of the western Mediterranean basin is made of extensional and transtensional passive continental margins. The continental crust is composed of Paleozoic and pre-Paleozoic rocks deformed by the Caledonian and Variscan orogenic cycles.

In the following the gross structure of the Mediterranean structure and geodynamics is reviewed. In particular, the present day lithosphere-asthenosphere system is analysed using two S-wave velocity cross sections and the lithospheric structure of the area is shown by lithospheric scale cross sections along two regional-scale geotraverses. The lithospheric geological cross sections are based on the result of the TRANSMED project (Cavazza et al., 2004) and the cross sections discussed in this work are a simplification of the TRANSMED II (Roca et al., 2004) and TRANSMED III (Carminati et al., 2004) sections. The S-wave velocity cross sections run parallel to the TRANSMED II and TRANSMED III geotraverses.

Finally the geodynamic evolution of the area in the last 45 Ma is analysed on a series of four evolutionary maps referred to the 45 Ma, 30 Ma, 15 Ma and 0 Ma time steps.

- Lithospheric structure from S-wave velocities

The method

A set of S-wave velocity models of the Mediterranean lithosphere-asthenosphere system is defined along two geotraverses (Fig. 18.4.1.2 and Fig. 18.4.1.3). The group velocity (in the period range 5 s - 150 s) and phase velocity (in the period range 15 s -

150 s) dispersion data for Rayleigh waves, collected in the Mediterranean region are transformed, by tomography, into a 2D (latitude, longitude) group and phase velocity distribution for a properly chosen set of periods. The 2D surface wave tomography provides group and phase velocity distribution and maps indicating the lateral resolution of the used data set.

To ensure sufficient data coverage on the whole Mediterranean area, waveform records of regional-distance earthquakes from broadband and very broadband stations in the region are collected and processed by frequency-time analysis, FTAN (Dziewonski et al., 1969; Levshin et al., 1972, Levshin, 1973; Keilis-Borok, 1989) in order to obtain group velocity dispersion curves of the fundamental Rayleigh wave mode. The records are taken from international and national data centers: IRIS (www.iris.washington.edu), ORFEUS (orfeus.knmi.nl), GEOFON (www.gfz-potsdam.de/geofon), GEOSCOPE (www.geoscope.ipgp.jussieu.fr), MEDNET (mednet.ingv.it), GRNS (www.szgrf.brg.de). The hypocenter, origin time and magnitude of each earthquake are taken from ISC (www.isc.ac.uk) and NEIC (neic.usgs.gov) data centers. The body waves magnitude mb for all events is greater than 4.5.

Phase and group velocities are considered in a complementary way. As a matter of fact one-station group velocity measurements are much less sensitive to source characteristics than phase-velocity measurements which are strongly affected by the apparent initial phase of the source (Panza et al., 1973; Muyzert and Snieder, 1996; Levshin et al., 1999). Short periods of the group dispersion data reach values of 5 s - 10 s for regional distances and 1 s on local distances, while phase velocities are difficult to be measured unambiguously below about 15-20 s, so they cannot be used to obtain the shallow velocity structure at crust level. The period range of the group velocity dispersion curves derived from the records at regional distances (300 km - 4000 km) varies typically between 5 and 80 s, while long period group velocity data from global studies (Ritzwoller and Levshin, 1998; Shapiro and Ritzwoller, 2002) are used to extend the period range to 150 s.

A 2D tomography method developed by Yanovskaya (1984), Ditmar and Yanovskaya (1987) and Yanovskaya and Ditmar (1990) is applied to estimate lateral variations in group and phase velocity dispersion distribution at different periods. The method permits the inversion of the group and phase velocity dispersion data as well as P-wave travel-time residuals data for 2D and 3D inhomogeneous media (Yanovskaya, 1984).

To obtain the shear wave velocity models by non-linear inversion, local dispersion curves are assembled within the considered range of periods, by taking the tomography results at each of the grid points, that define the 1 degree by 1 degree cells crossed by the geotraverses. The method known as "hedgehog" is an optimized Monte Carlo non-linear search of velocity-depth distributions (Valyus et al., 1969; Knopoff and Panza, 1977; Panza, 1981) and it is independent from the starting model since. A priori and independent information, wherever it is available, like seismic, geophysical and geological data, derived from previous studies, in the area of the Mediterranean Sea and surrounding regions, is used in the parametrization of the non-linear inversion in order to optimize the resolving power of the data. The velocity structure is well resolved in the depth range from about 5 to 250 km. A set of solutions in each cell is obtained because of the well-known non-uniqueness of the inverse problem.

To construct the cross-sections along the geotraverses a model should be chosen for each cell. An objective criterion, based on an optimization method imposing lateral

smoothness, is applied to derive a representative solution for each cell. The technique is well known in operational research theory. We use the so called Local Smoothness Optimization (LSO) that fixes the cellular model as the one that has minimal divergence in velocity between neighboring cells. The chosen solutions for each of the cells crossed by TRANSMED II (Fig. 18.4.1.2) and TRANSMED III, (Fig. 18.4.1.3) give a sketch of the lithosphere-asthenosphere system along these profiles. The main features of the two cross sections of Fig. 18.4.1.2 and 18.4.1.3 can be summarised as follows.

Section parallel to TRANSMED II
1. Cells from 1 to 5 have a typical continental structure with the average velocity in the mantle about 4.35 km/s - 4.40 km/s.
2. The structure of cell 6 corresponds to the rift zone in the region. The velocity in the mantle varies slowly (4.25 km/s - 4.35 km/s) down to about 200 km depth.
3. The cell 10 (central Algerian basin) with thin lid (15 km to 30 km) and low velocity asthenosphere (4.10 km/s - 4.15 km/s) has an oceanic signature. Cells 7 in the Valencia trough and 11 in the southern Algerian basin have transitional character, suggesting to be thin continental lithosphere.
4. The lid under Balearic islands is thin and fast (10 km - 15 km and velocity 4.70 km/s - 4.80 km/s).
5. The lithosphere beneath north Africa is faster than in the Mediterranean (4.40 - 4.55 km/s, and its thickness is about 100 km.
6. There is a large contrast in lithospheric thickness (between cells 10 and 11 in north Africa.
7. In the cells 12 - 15 the thick lid overlying a very low velocity zone that evidences the presence of a well developed asthenospheric channel. There is no striking evidence for a continuos slab below northern Africa below 70-100 km.

Section parallel to TRANSMED III
1. Cells 1 and 2 have thick lithosphere with respect to the next cells: about 170 km with the velocity 4.40 km/s - 4.50 km/s.
2. The cells from 1 to 4 have higher velocities in the mantle with respect to the following cells and represent the NW continental part of the geotraverse.
3. It is clearly visible a low velocity channel under all the Mediterranean from cell 5 to cell 17, at depths from ~ 20 km to ~ 140 km.
4. In the cell 16 a low velocity hot mantle reservoir is well visible, potentially interpreted as the shallow asthenosphere in the backarc basin, and sourcing the oceanic crust flooring the Tyrrhenian basin.
5. There is no evidence for deep hot mantle plume under the Tyrrhenian sea.
6. The mantle, consistent with the presence of a continental lithosphere under cells 20, 21, and 22, is quite different from that of North Africa, as seen along TRANSMED II in cells from 11 to 15 (Fig. 18.4.1.2).
7. The cell 23 more or less coincides with local maximum of the heat flow (85 kW/m2 and more). In this cell, there is a relatively low velocity layer centered at about 100 km of depth, very likely a remnant of the Mesozoic rifting.
8. Cells from 24 to 27 have a gentle velocity inversion and together with cell 28 represent a structure of the mantle consistent with a continental structure, where the lithosphere is quite thick.
Seismicity and high velocities correlate quite well in the Apennines and in the Dinaric subduction zone.

- The Western Mediterranean sub-basins

The geological evolution of the western Mediterranean sub-basins and of the intervening continental swells is described in this section from west to east. The main features analysed in the following can be recognised in the simplified tectonic map of Fig. 18.4.1.1 and in the two lithospheric scale cross sections of Fig. 18.4.1.4 which are here only roughly described. For a complete description of TRANSMED II and III transects refer to Roca et al. (2004) and Carminati et al. (2004).

Along the TRANSMED II section the following main domains are crossed (the region analysed in this paper is limited to the central part of this transect):

- At its northern end the transect crosses the Aquitaine basin, which is the northern foreland basin of the Pyrenees. More to the south, the Pyrenees, an orogenic belt which developed between the late Senonian (Late Cretaceous) and the mid-Oligocene times, are crossed together with their southern foreland basin (the Ebro basin).

- The section continues through the Catalan Coastal Ranges, an intraplate belt developed during the Paleogene and crosses the Iberian coast, characterised by continental lithosphere thinning progressively seaward, as seen from Fig. 18.4.1.2.

- The Valencia trough lies on continental lithosphere thinned in the uppermost Oligocene-Lower Miocene.

- The Valencia trough is bordered to the south by the Balearic Promontory, composed of relatively thicker continental lithosphere interested in the Tertiary by both compressional and extensional tectonics.

- South of the Balearic islands the Algerian basin, a Miocene basin floored by oceanic crust, is crossed.

- The African portion of the transect crosses -from north to south- the Tell and the Sahara Atlas (two fold-and-thrust belts), and ends in the northern part of the Saharan platform. The Tell developed on top of a southeasterly retreating northwest dipping slab, which became likely detached and therefore is not drawn in the section. Here the lithosphere sits on a very low velocity asthenospheric channel (Fig. 18.4.1.2) , that seems to be a general feature of the North-Central Pan-African Orogenic block (Hazler et al., 2001).

Along the TRANSMED III section the following main domains are crossed:

- At its northwestern end the section crosses the French Massif Central, made of the continental crust deformed during the Variscan orogeny. The continental lithosphere thins towards the French coast where the transect crosses the Gulf of Lions thinned continental margin.

- The transect continues in the Provençal Basin, floored by Neogene oceanic crust and through the thinned continental lithosphere of the western Sardinia margin.

- Eastward, the profile crosses the continental swell of the Corsica-Sardinia block, which was structured during the Variscan and older orogenic cycles, and was later dissected by Neogene-Quaternary extensional tectonics.

- East of Sardinia the lithosphere thins again reaching its minimum thickness in the Tyrrhenian Basin, which formed mainly from the Tortonian to the Present. Farther east the lithosphere gradually thickens and the thinned continental lithosphere of the Campania continental margin is encountered, as seen also in Fig. 18.4.1.3.

Although the area of interest for this paper is limited to the described northwestern part of the TRANSMED III profile, the geotraverse is shown entirely in Fig. 18.4.1.4 and the following domains are encountered:

- Continental Italy is mostly constituted by the Southern Apennines, a Neogene fold-and-thrust belt, dissected by Late Neogene-Present extensional tectonics, developed in the hangingwall of a west-directed subduction zone.
- The continental lithosphere of the Adria microplate, thinned during the Mesozoic rifting is crossed in the southern Adriatic Sea. In the eastern southern Adriatic Sea the transect crosses the foreland basin of the Albanian Dinarides, an orogen associated with the north-eastward subduction of the Adriatic lithosphere.
- In Albania, Macedonia and Bulgaria, the complex multistage Dinarides-Hellenides orogen is crossed together with its conjugate retrobelt, i.e., the Balkans. The whole region is also affected by Neogene-to-Present extensional tectonics. The transect ends in the Moesian platform, which is the undeformed foreland of the Balkans.

Most of the described domains can be readily identified in the S-wave velocity pattern of fig. 18.4.1.2 and 18.4.1.3.

Valencia Trough

The Valencia Trough is located between the Iberian mainland and the Balearic Islands, at the southwestern prolongation of the Liguro-Provençal basin and has water depths of up to 2200 m. VALSIS and ESCI profiles provide an overall view of the stratigraphy and structure of the basin (Torres et al., 1993; Sàbat et al., 1995). A complete description of the Valencia Trough and a synthesis of its evolution is included in Roca, 2001 and Roca et al., 2004.

The crust of the Valencia Trough is continental. Depth to the Moho decreases from about 32 km in the Iberian mainland to 10-15 km along the axis of the Valencia Trough and increases again towards the Balearic Promontory to 23-25 km (Pascal et al., 1992; Torné et al., 1992; Vidal et al., 1998; Roca, 2001). The minimum thickness of the crust (along the axis of the Valencia Trough) is about 8 km. Lithosphere is also thin, depth to lithosphere - asthenosphere boundary ranges from 50 to 80 km (Ayala et al., 2003), and the mantle lithosphere is characterized by an anomalously low seismic velocity (Pascal et al., 1992), see also Fig. 18.4.1.3.

The continental crust of the Valencia trough was structured during the Variscan orogeny and was extended during the Mesozoic rifting phases (Roca, 2001). The resulting Mesozoic basins were totally or partially inverted during latest Cretaceous-Oligocene as a consequence of the convergence between Iberia and Eurasia (Fernàndez et al. 1995; Roca, 1996; Gaspar-Escribano et al., 2004). This resulted in erosion and in the concomitant development of a major unconformity over the whole area (Stoeckinger, 1976; Martínez del Olmo, 1996), except in minor syn-compressional basins filled by Eocene to Upper Oligocene terrigenous sediments (i.e., Barcelona basin; Roca et al., 1999). Overlying this unconformity, four major packages have been differentiated in the 2 to 6 km thick València trough basin fill (Clavell and Berástegui 1991; Maillard et al. 1992; Martínez del Olmo 1996; Roca et al. 1999). From bottom to top, these are: (1) syn-rift uppermost Oligocene-Lower Miocene continental deposits and terrigenous outer-shelf marine sediments which are restricted to grabens and to the deepest parts of the basin; (2) post-rift Middle and Upper Miocene basinward prograding clastic sequences; (3) Messinian salt deposited in the deeper part of the basin coevally with the development of a major down-cutting unconformity in the shallower part of the basin; (4) Pliocene-Holocene terrigenous sediments deposited in basinward prograding deltas and deep-sea fans.

From the Oligocene-Miocene transition to the early Middle Miocene, the Valencia Trough was characterised by severe extensional tectonics and developed its present

structure.The formation and development of the Valencia Trough were accompanied by widespread volcanic activity which, as in the case of the Liguro-Provençal basin, was (1) calc-alkaline and related to subduction before the Serravallian, and (2) alkaline with a typical intraplate geochemical signature after the Serravallian (Martí et al., 1992).

The upper crustal structure of the Valencia Trough is clearly asymmetric (Fontboté et al. 1990; Sàbat et al., 1995). The Iberian slope (the NW side of the trough) shows a thick-skinned extensional structure characterized by NE-SW striking normal faults. Most of these faults dip to the NW, except the major ones, located along the northwestern basin margin in the proximity of the coast or onshore (Catalan Coastal Ranges), which dip to the SE. These last faults involve partial or total reactivation of Paleogene thrusts faults which, in turn, reactivate Mesozoic extensional faults (Fontboté, 1954; Roca, 2001). In contrast, the Balearic slope is characterised by the presence of basement highs, bounded by both normal and thrust faults; these basement highs have been interpreted as fault-propagation or fault-bend folds associated with SE-dipping thrust faults (Sàbat et al., 1995; Gelabert, 1997).

Finally, inversion of extensional structures and development of contractional structures resulting from the NW propagation of the Balearic fold-and-thrust belt affected the Balearic side of Valencia Trough during the Middle and Late Miocene (Sàbat et al., 1995; Roca et al., 2004).

Balearic Promontory

It is located between the Valencia Trough and the Algerian Basin. There is no clear boundary between the Valencia Trough and the Balearic Promontory; its boundary with the Algerian Basin is outlined by the sharp Emile Baudot Escarpment. Balearic Islands are the emergent areas of the Balearic Promontory. The Balearic Promontory, which forms part of the margin of the Iberian Peninsula, is surrounded by Neogene extensional basins (Valencia Trough and Algerian Basin) and shows a Neogene basin and range structure. Data are obtained from very good outcrops and seismic lines (most of them off-shore).

Crust in the Balearic Promontory reveals a continental character; it is thin (at most around 25 km) but thicker than in surrounding areas.

The basement is of Variscan type and outcrops extensively in Menorca. Sedimentary cover comprises Mesozoic calcareous sedimentary series similar to those in the Catalan Coastal Ranges and the Iberian Range: Triassic Germanic facies, Early Jurassic shallow water limestones, Middle and Late Jurassic pelagic marls and limestones and Cretaceous pelagic marls and shallow water limestones (Rangheard, 1984; Bourrouilh, 1983; Alvaro et al., 1989). A stratigraphic gap embraces part of the Cretaceous and the Paleogene (Ramos-Guerrero et al., 1989). Paleogene sediments are thin and discontinuous on the Balearic Promontory. They are made up of Eocene shallow water limestones (only present in the southeast) and Oligocene continental conglomerates and other detritic rocks. Neogene sediments are widespread on the Promontory. They show both syn-orogenic and post-orogenic character. Syn-orogenic Neogene sediments consist of conglomerates, calcareous sandstone, peri-recifal limestones and calcareous turbidites. Post-orogenic Neogene sediments comprise recifal limestones, calcareous sandstone and marls.

Only two outcrops of Cenozoic volcanic rocks are known in the Balearic Islands (Martí et al., 1992). They are made up of calc-alkaline volcanic rocks Early Miocene

in age. There is an extensive submarine volcanic field in the SE margin of the Promontory (between Formentera and Cabrera islands) (Acosta et al., 2001).

The structure of the Promontory is typical of those existing in the basin and range areas:

a) Ranges consist of folded and thrusted series where the major detachment is the Triassic Keuper facies (gypsum and lutites). Shortening is high (50%) and increases to the SE (Gelabert et al., 1992; Gelabert, 1997). In the southeastern areas of the islands of Eivissa and Mallorca, large overturned folds are present. Most of the contractional features strike NE-SW, but NW-SE structures are also present (Sàbat et al., 1988; Freeman et al.,1989). Deduced overall shortening is NW-SE (Sàbat et al., 1988; Gelabert et al., 1992). In Mallorca, the oldest contractional structures are Late Oligocene in age and the youngest are Langhian (Ramos-Guerrero et al., 1989). In the other islands, contractional structures seem to be of similar age, but contractional structures in off-shore areas to the NW of the islands are Middle-Late Miocene in age (Sàbat el al., 1995). Contraction shows an overall SE to NW propagation.

b) Basins have both an extensional and contractional origin. Although normal faults are key structural elements of the basins, transversal contractional highs play an important role. Few normal faults are Oligocene in age but most of them are Serravallian. The oldest sediments are Early or Middle Miocene in some basins although they are Late Miocene in most of the basins.

The Balearic Promontory underwent tectonic quiescence during the Paleogene (accompanied by erosion and shallow marine and sub-aerial sedimentation) and during the Neogene "sensu latu" (Late Oligocene, Early and Middle Miocene) contraction took place (accompanied by sedimentation in shallow marine and talus environments). This contraction was followed by the Middle and Late Miocene (mostly Serravallian) extension (Cespedes et al., 2001). Following a period of quiescence, contraction resumed during the Plio-Quaternary (Gimenez, 2003) (accompanied by shallow marine and sub-aerial sedimentation). Paleomagnetic data suggest a clockwise rotation of the Balearic Promontory during the Neogene (Parés et al., 1992).

The lithosphere, about 30 km thick sits on a very low velocity asthenosphere (Fig. 18.4.1.2).

Algerian Basin

This is a deep basin (around 3000 m) located between the Balearic Promontory and the North Africa margin. The boundary between the Balearic Promontory and the Algerian Basin is sharp (the Emille Baudot Escarpment). There are few data on the Algerian Basin; in addition to overall pictures from gravity and magnetic maps, the only data available are a few seismic refraction data (Hinz, 1973), short segments of seismic reflection profiles (Sàbat et al., 1995; Mauffret et al., 2004) and several boreholes - DSDP 124, 371 and 975 (several references in Mauffret et al., 2004).

The crust in the Algerian Basin seems to have an oceanic character. In most of the basin the Moho depth is less then 14 km and the crust is 4 to 6 km thick (Hinz, 1973), being comparable to the oceanic crust (about 5 km thick) of the Liguro-Provençal basin (Pascal et al. 1993). In the East Alboran basin the heat flow is very high (125 mWm^{-2}; Polyak et al. 1996), suggesting a shallow Moho (10 -11.5 km) and a thin lithosphere (45 km) (Torné et al., 2000). A slightly thicker lithosphere (50-60 km) has been proposed (Marillier and Mueller, 1985; Ayala et al., 2003) for the Algerian

basin. In Fig. 18.4.1.2 the lithosphere is not more than 40 km thick and sits on a well developed asthenospheric low velocity channel.

The scarcity of deep drilling in the Algerian basin hampers the determination of the characteristics and the age of its sedimentary fill. Thus, the stratigraphy must be inferred by correlation from better known areas in the proximity (Mauffret et al. 1973; Sans and Sàbat, 1993). Boreholes only indicate that the Pliocene-Quaternary sedimentary layer is very thin (0.5 km) and that the Messinian evaporites (1.2 km thick) are highly disturbed by diapirism. Beneath the Messinian evaporites, there is a 2-3 km thick non reflective pre-Messinian layer cut by some oil wells close to the Algerian coast; these wells indicate the presence of Langhian to Tortonian deposits (Cope, 2003).

The ESCI seismic profile crosses an area (South of Mallorca) whose sea bottom is flat but affected by a number of normal faults with small displacements, where Messinian evaporites form spectacular diapirs (Sàbat et al., 1995). Seismic data from the area located to the North of the Great Kabylie also reveal normal faults; the displacements produced by these faults in the Messinian evaporites are smaller than those produced at the top of the basement (Cope, 2003; Mauffret et al., 2004). According to the map representing the depth to the basement, it has been suggested that these normal faults strike NNE-SSW (Mauffret et al., 2004).

In addition to the aforementioned faults, very little else is known about the internal structure of the Algerian Basin. Nevertheless, the three following features should be pointed out: a) The Emille Baudot escarpment, b) the Hannibal Ridge and c) the Hamilcar magnetic anomaly.

a) The Emille Baudot escarpment (Mauffret et al., 1992) is the surface expression of the fault system related to the boundary between the continental crust of the Balearic Promontory and the thin oceanic crust of the Algerian Basin; displacement of the synthetic faults shows a clear normal component (Sàbat et al., 1995). The Emille Baudot escarpment strikes NE-SW and at its SW end shows a horse-tail configuration with individual faults striking ENE-WSW turning to E-W (Acosta et al., 2001). Further W a sharp escarpment reappears, it strikes NE-SW and finally (further W) it runs E-W (Mazarron escarpment). Thus, the overall trace of this prominent fault system shows a zigzag geometry. It has been suggested that this fault system, at present inactive, had behaved as a right-lateral strike-slip system (Mauffret et al., 1992; Acosta et al., 2001), but no conclusive arguments are available. Moreover, the geometry described above is more consistent with a NNW-SSE extension (this extension is sub-parallel to the one in the Valencia Trough).

b) The Hannibal ridge is located to the North of the Great Kabylie and is parallel to the aforementioned NNE-SSW normal faults; it has been suggested that this is a Late Miocene volcanic feature related to E-W extension (Mauffret et al., 2004).

c) The Hamilcar magnetic anomaly is located at the boundary between the oceanic areas of the Liguro-Provençal and Algerian basins. It consists of a set of fan shaped anomalies striking NW-SE and converging towards the NW (Galdeano and Rossignol, 1977). Its geometry is consistent with a local NE-SW extension (see Fig 5 in Gelabert et al., 2002). This extension should be coeval with the formation of the oceanic crust in the Algerian basin and in the Liguro-Provençal Basin. This crust has been dated as Burdigalian in the latter basin.

No volcanic rocks have been sampled from the Algerian Basin although they are visible in the seismic profiles. Nevertheless, volcanism recorded in the North African margin shows that the area was affected by a calc-alkaline volcanic event Aquitanian

to early Serravallian in age. This was followed by alkaline volcanism until the present (Wilson and Bianchini, 1999; Maury et al., 2000; Coulon et al. 2002).

The age of the Algerian Basin is unknown but must be much older than the dated Messinian deposits, as these deposits overlie a 2-3 thick km sequence of undated basin fill sediments. In this regard, the unloaded depth of the basement gives an apparent age of 20 Ma (Cavazza et al., 2004; Roca et al., 2004) whereas the maximum heat flow of 132 - 125 mWm^{-2} corresponds to an apparent age of 16 Ma for the youngest areas in the basin. Tomographic studies (Carminati et al. 1998a) and magmatic history (Maury et al. 2000) suggest an age ranging between 15 Ma and 10 Ma; and, finally, partial restorations of balanced cross-sections point to an age ranging between 25 to 15 Ma (Vergés and Sàbat, 1999) or 20 and 13 Ma (Roca et al., 2004). These last ages agree with the ages deduced from a) the theoretical thickness-age relationships $y_L=2(kt)^{1/2}$ (where y_L is the lithospheric thickness, t is the lithosphere age and k is the thermal diffusivity; Turcotte and Schubert, 1982) – assuming $k=1$ mm^2 s-^{1}, a thickness of 40 km, as imaged in Fig. 18.4.1.2, yields an age of 12.7 Ma; and b) the suggestion that Algerian and Liguro-Provençal basins are similar in age (21-16 Ma). This suggestion is supported by the fact that these two basins are similar in depth, crustal thickness and layer thickness ratios of the pre-Messinian, Messinian and post-Messinian sediments (Mauffret et al., 2004). Thus, the ages of the two basins should be similar but the sedimentation rate in the Algerian basin must be different (much lower) than that in the Liguro-Provençal Basin.

Liguro-Provençal Basin

This basin comprises the present day Ligurian Sea, the Gulf of Lions as well as the portion of the Mediterranean Sea located West of Corsica and Sardinia, and East of Menorca. It is the oldest Western Mediterranean basin and today has a maximum water depth of 2800 m. The ECORS-CROP seismic section provides an overall view of the stratigraphy and structure of the basin (De Voogd et al., 1991). A description of the Liguro-Provençal basin and a synthesis of its evolution is included in Roca (2001).

The lithosphere beneath the Liguro-Provençal basin is thin (minimum depth to the lithosphere-asthenosphere boundary is less than 30 km, as shown in Fig. 18.4.1.3) and seismic and gravity data show that the crust decreases in thickness from about 32 km beneath the Eurasian mainland and the Corsica-Sardinia Block to about 5 km beneath the central part of the basin where it is oceanic (Pascal et al., 1993; Chamot-Rooke et al., 1997). This oceanic crust occupies the central part of the basin and delineates a 150 km wide string oriented NE-SW bounded by a broad transitional crust. The oceanic crust in the Liguro-Provençal basin has been dated as late Aquitanian to late Burdigalian-early Serravallian on the basis of heat flow, tristanite in samples, paleomagnetism and subsidence analyses (Vigliotti and Langenheim, 1995).

The Liguro-Provençal basin fill comprises a fairly complete succession of lower or upper Oligocene to Holocene deposits that can be subdivided into syn-rift and post-rift series (Gorini et al., 1993). The syn-rift series, Late Eocene to Aquitanian in age, are restricted to grabens and to the deepest parts of the basin, and are made up of coarse grained alluvial to lacustrine successions which grade laterally into marine deposits. Post-rift series form the bulk of the sedimentary fill of the Liguro-Provençal basin and are late Aquitanian to Holocene in age. They consist of marine deposits mainly sedimented in terrigenous shelf, delta and deep sea fan environments. The rapid Messinian drop in the level of the Mediterranean sea is recorded by the

development of a major down-cutting unconformity in the shallower parts of the basin and by the accumulation of thick evaporites in its deeper part.

The formation and development of the Liguro-Provençal basin were accompanied by widespread magmatic activity with volcanic centres located both offshore (i.e., Ligurian Sea) and onshore (south France, Corsica and Sardinia islands). The geochemical signature of these volcanic rocks clearly shows a magmatic evolution with two-well differentiated cycles (Roca, 2001): the first cycle, Late Oligocene-early Serravallian in age, is calc-alkaline and mainly located along the Corsica-Sardinia Block, Ligurian Sea and neighbouring southern Alps. The second one is alkaline, spans from Tortonian to Holocene times and affects the entire basin with extrusive centres clustered onshore in the Languedoc-Montaigne Noire (southern France) and Sardinia.

The structure of the Liguro-Provençal basin reflects its extensional origin. It consists of 1) a broad northwestern extensional margin with ENE-oriented grabens and horsts mainly bounded by SE-dipping faults; the detachment of the faults is shallow and located in the Triassic evaporites to the NW and ramps down seawards into the basement where it gives way to a thick-skinned extensional system; 2) a central and flat oceanic part with Pliocene diapirs cored by Messinian salt; and finally 3) the narrow and abrupt Corsica-Sardinia margin whose internal structure is not clear (Gorini et al., 1993; Séranne et al., 1995; Roca, 2001).

Rifting in the Provençal-Ligurian basin started during latest Eocene-Early Oligocene (34-28 Ma) and ended in the middle Aquitanian (21 Ma) (Séranne, 1999). Subsequently, the central oceanic portion of the basin was generated between the late Aquitanian and late Burdigalian (21-16 Ma) associated with the counter clockwise rotation of the Corsica-Sardinia Block (Vigliotti and Langenheim, 1995; Speranza et al., 2002). Before drifting, this block of the Iberian plate was located close to the Provençal coast and the present-day, Gulf of Lions.

The Liguro-Provençal Basin is considered to be a back-arc basin generated from the southeastward roll-back of the Apennines-Maghrebides subduction (Rehault et al., 1984; Doglioni et al., 1997).

The Corsica-Sardinia block

The Sardinia block is characterized by a crust about 20 km thick that is getting thicker as we move towards East; on the West (Sardinia trough) the lid is thin but fast. Therefore, the lithosphere reaches about 40 km on the West side, while is it about 70 km thick on the East (Fig. 18.4.1.3).

In Sardinia abundant traces of Paleozoic tectonics (both compressional and extensional) ascribed to the Variscan cycle and of late Paleozoic magmatism are recognised (Carmignani et al., 1994). The Variscan tectono-metamorphic edifice of Sardinia is composed by several tectonic units. Their emplacement time ranges from Devonian to Carboniferous (Carmignani et al. 1994) and the direction of transport varied through time from top to the south to top to the west (Conti et al. 2001). In a later stage, N-S shortening resumed. The Variscan Belt was dissected by normal faults pertaining to the late Variscan stages, which were accompanied by late Carboniferous igneous activity and by LP/HT metamorphism.

During the Permo-Triassic Sardinia experienced continental rifting and, during the Jurassic, it was located along the western margin of the Alpine Tethys. Being until the Oligocene attached to the Iberian plate, the Oligocene-Aquitanian NE-SW trending normal faults, that dissected previously deformed areas of the Pyrenees and of the

Iberian chain in the Camargue and Gulf of Lions rifts, affected also Sardinia. These faults cut through pre-Alpine basement in the Sardinia rift (Burrus 1989). The rift is bounded by west and east dipping normal faults and it is filled by up to 2 km of syn- and post-rift deposits, as constrained by seismic lines (Casula et al. 2001). Syn-rift Upper Oligocene deposits mainly consist of non-marine sediments followed by fluvial-deltaic sediments, marls and silts. Post-rift sediments are Burdigalian marine marls-siltstones. Marine sedimentation continued until the Messinian regression (accompanied by an erosional stage). Pliocene-to-recent sedimentation was mainly localised in the Campidano graben (SW Sardinia).

Corsica can be divided in two distinct geological domains (Durand Delga, 1978). The western two thirds of the island (Variscan Corsica) consist mainly of a large late Variscan (Carboniferous to Permian; 340-260 Ma Cocherie et al., 1984) granite batholith, which intruded Precambrian and Palaeozoic country rocks. Scattered portions of this Variscan metamorphic basement crop out within the late Carboniferous granitoids. These metamorphic outcrops are characterized by poly-metamorphic histories (Menot and Orsini,1990), and different metamorphic grade (from granulite and eclogite to anchizonal facies (Lardeaux et al., 1990; Thevoux-Chabuel et al., 1995). The Variscan basement is cut by major late Variscan left-lateral strike-slip faults mainly oriented NE-SW.

The northeastern part of Corsica (Alpine Corsica) consists of a complex stack of unmetamorphosed to eclogite facies (Brunet et al., 2000; Malavieille et al., 1998) tectonic units of Alpine age of both oceanic and continental origin, interpreted as the southern continuation of the Penninic domain of the Western Alps. The Alpine Corsica units were thrust westward onto the Variscan basement of western Corsica (Durand Delga, 1978; Mattauer et al., 1981). The oceanic rocks are interpreted as remnants of the Liguro–Piemontese ocean, a portion of the western Tethys ocean which was located between the European and Adriatic continental margins and was subducted since the Cretaceous. The subduction process was accompanied by top-to-the-west thrusting and development of HP/LT metamorphism (Mattauer et al., 1981) and was followed by continental collision and underthrusting of continental crust during Early Tertiary time.

During the Oligocene, the sense of shear reversed and wideperead top-to-the-east extension dissected the previous Alpine edifice (Daniel et al., 1996; Jolivet et al., 1990). Early Miocene rapid denudation and cooling accompanied the extensional collapse of the Alpine belt (Brunet et al., 2000; Cavazza et al., 2001).

Extensional tectonics led to the formation of strongly subsiding basins (e.g., Saint Florent and Aleria basins), filled with up to 2 km sediments deposited since the Burdigalian (Ferrandini et al., 1996) and continuous with the offshore Corsica basin.

Late Oligocene-early Serravallian calc-alkaline volcanism is widespread along the Corsica-Sardinia Block (Savelli 1988; Beccaluva et al. 1989). A later phase of volcanism started in Tortonian-Messinian times and is of alkaline affinity.

The present day position of the Corsica-Sardinia block is the result of the above mentioned Oligocene-Aquitanian rifting phase. These extensional tectonics began to separate Corsica and Sardinia from the Iberia peninsula. The subsequent oceanization of the Provençal Basin was accompanied by an anticlockwise rotation of the Corsica-Sardinia block during Early Miocene (Vigliotti and Langenheim 1995; Speranza et al. 2002). Such a rotation determined the displacement of the Variscan Belt cropping out in Sardinia from its counterpart cropping out in Iberia and in the French Massif Central.

The wide eastern Sardinia and Corsica continental margin marks the transition from

the Corsica-Sardinia block to the Tyrrhenian basin and its extensional tectonics is related to the opening of the Tyrrhenian, rather than to the Gulf of Lions and Provencal basins, as observed in the basins of Sardinia. It is characterised by irregularly spaced blocks dissected by listric normal faults (with basal detachment in the lower crust) dipping both landward and seaward. Resulting half grabens are filled by Late Tortonian to Messinian syn-rift sediments (including thick Messinian evaporites) and sutured by Pliocene-Quaternary post-rift sequences. Serravallian-Tortonian pre-rift sediments (mainly clastics) occur in the uppermost Sardinia margin. The passive margin basement consists of metamorphic rocks structured during the Variscan and Alpine cycles (including Tethyan ophiolites and their cover sequences) covered by syn-orogenic clastic sediments.

The Tyrrhenian Sea

The Tyrrhenian basin is the easternmost sub-basin of the wider western Mediterranean backarc basin, developed since the Late Oligocene in the hangingwall of the Apennines-Maghrebides "west"-directed subduction zone, which generated the arc running from northwest Italy throughout the Italian peninsula, Sicily and the north-western margin of Africa, from Tunisia to Morocco (Rehault et al., 1984; Gueguen et alii, 1998). The Tyrrhenian Sea, characterised by a triangular shape and by water depths in excess of 3500 m, represents the recentmost sub-basin, developed from Middle Miocene to Present (e.g., Scandone, 1980; Malinverno and Ryan, 1986). A detailed discussion of the geological features and of the geodynamic setting of this sub-basin can be found in chapter 17.3.3.2, while in the following only a brief description of its main characteristics will be provided.

Positive Bouger anomalies (>250 mGal; Mongelli et al., 1975; Consiglio Nazionale delle Ricerche, 1992) and very high heat flow values (>200 mW/m^2; Cataldi et al., 1995; Della Vedova et al., 2001) have been measured in the Tyrrhenian Sea. Moho maps (e.g., Scarascia et al., 1994: Nicolich, 2001) show depths shallower than 15-20 km below the bathyal plane, and two minima of 10 km centered on the Vavilov and Marsili basins. All these data are consistent, as can be seen from Fig. 18.4.1.3, with the presence of a thin lithospheric mantle and of a shallow asthenosphere with relatively low S-wave velocity of 4.00-4.30 km/s (e.g., Panza and Calcagnile, 1979; Panza et al., 2003) as also confirmed by mantle tomography and by Q values (Doglioni, 1991; Peccerillo and Panza, 1999; Piromallo and Morelli, 2003; Mele et al., 1997). The coincidence between the highest values of the heat flow and domains characterised by a thin crust (e.g., Vavilov and Marsili basins) and their position, shifted in the south eastern side of the basin, highlight the asymmetry of the rift and possibly of the underlying mantle.

The onset of the rifting processes on the upper Sardinia slope, responsible for the separation of the Calabrian and Peloritan massifs from the Sardinia block, can be referred to the late Burdigalian (Gueguen et al., 1998), although the syn-rift evolution is generally ascribed to the Upper Tortonian-Messinian (e.g. Moussat et al., 1986; Mascle and Rehault, 1990). Rifting process and the related magmatism migrated in time from west to east but in the south-eastern part it deviated to SE since late Pliocene (e.g., Kastens et al., 1988; Sartori, 1989; Doglioni, 1991; Savelli, 2002). The extensional processes generated oceanization in the Vavilov (7-3.5 Ma) and Marsili (1.7-1.2 Ma) sub-basins (Bigi et al., 1989).

Stretching in the Tyrrhenian Sea decreases from south to north, and there seems not to be a linear relation between total extension and heat flow. However there rather

appears an evident correlation between active magmatism and heat flow.

The Italian and Sardinia-Corsica continental margins are dissected by mainly listric (dipping both basinward and landward) normal faults rooted in the lower crust.

The listric faults disappear across the oceanized sectors. In the Vavilov Basin 4 Ma old MORB basalts were drilled by ODP Site 655 below the Pliocene-Quaternary sequence (Beccaluva et al. 1990; Serri et al. 2001). Moreover, a section of sub-crustal foliated and serpentinized upper mantle occurs in the eastern Vavilov Basin, as shown by ODP Site 651 drillings (Kastens et al. 1987; Bonatti et al. 1990). The peridotites are capped by 2.6 Ma old pillow lavas.

The bottom of the Tyrrhenian basin is expected to be floored with remnants of the Variscan and Alpine belts dissected by the Tertiary extensional tectonics. For example, along the steep escarpment of the Flavio Gioia Seamount, Variscan basement rocks, tectonically associated with Mesozoic-Tertiary sedimentary rocks (likely involved in the Calabride-Kabylide fold-and-thrust belt), were dredged (Dal Piaz et al. 1983; Sartori 1986).

- Geodynamic evolution of the Western Mediterranean area

Restoring the Corsica-Sardinia block to its Middle Eocene (45 Ma; Fig. 18.4.1.5) position prior to rotation, the Alps were probably linked to the Betics through Alpine Corsica and the Balearic promontory to form a double vergent belt (Doglioni et al. 1999). The development of the Alpine-Betic chain is related to a south-eastward subduction occurring at that time in the Tethyan area, as testified by HP mineral assemblages known in the Alps, in Corsica, in the Alboran domain and possibly in the Kabylies domain. The Alpine chain development was synchronous to the Pyrenees shortening (that continue until ~24.7 Ma according to a magnetostratigraphic study; Meigs et al. 1996) resulting from the complete inversion of a basin intervening between Iberia and Eurasia.

Some 30 Ma ago (Fig. 18.4.1.6), the west directed Apennines-Maghrebides subduction started. As proposed by Doglioni et al. (1999), it possibly nucleated along the Alps-Betics retrobelt and was possibly triggered by the occurrence, in the foreland east of the Alpine belt, of oceanic or thinned continental lithosphere. Following the same authors, a period of contemporaneity between Alpine and Apennines subductions possibly occurred from Late Oligocene to Early Miocene.

The Apennines-Maghrebides subduction zone experienced a fast radial (towards the north-east, south-east and the south) rollback, testified to by the migration of the subduction related calkalcaline volcanism. The rollback of the subduction hinge determined the onset of widespread extensional tectonics in the backarc (in the Gulf of Lions and Provençal Basin, in the Catalan Coastal Ranges area, in Sardinia, still attached to Iberia, in the Valencia trough and in the Algerian Basin). Regional thinning was accompanied by alkaline volcanism and, in the Provençal and in the Algerian Basins, continental crust stretching evolved into oceanization. In summary: the Valencia Trough, together with the Liguro-Provençal basin, records the onset of the southeastward roll-back of the Apennines-Maghrebides subduction, which generated the opening of both basins as back-arc basins and the coeval development of the Balearic fold-and-thrust belt (Doglioni et al., 1997; Vergés and Sàbat, 1999; Roca, 2001). The active extension and oceanization in the Liguro-Provençal basin is coeval with the counterclockwise rotation of the Corsica-Sardinia block.

The geodynamic origin of the Algerian basin is less constrained. It has been related to the roll-back of the Apennines-Maghrebides subduction, but as regards the direction

of this roll-back, two different kinematic scenarios have been proposed: 1) south to southeastward retreat of the subduction generating a NNW-SSE extension; in this case, Kabylies must have moved away from the Balearic Promontory (Verges and Sàbat, 1999; Roca et al., 2004), 2) westwards retreat of the subduction resulting in a ENE-WSW extension; in this case, the Alboran block was originally located as far as the present location of the Algerian Basin and must have moved a considerable distance to its present position (Martinez-Martinez and Azañon, 1997; Platt et al., 2003; Mauffret et al., 2004; Spakman and Wortel, 2004). Most of the data supporting these two hypotheses are derived from the areas around the Algerian Basin and very few come from the basin itself. In the Mediterranean Sea, which has been comprehensively studied, there are still blank areas on the map, and the Algerian Basin is one of these. The first hypothesis is here preferred, i.e. between 20 and 15 Ma a NNW-SSE large extension associated with SSE roll-back of the Maghrebides subduction led to the opening of the Algerian Basin.

The part of the Alps-Betics orogen which was located in the area of the Apennines-Maghrebides backarc basin was disarticulated and spread-out into the western Mediterranean (e.g., the metamorphic slices of Kabylie in N-Algeria, and Calabria in S-Italy).

The Apennines and Maghrebides fold-and-thrust belt developed on top of the retreating subduction and the deformation front migrated to the east in the Apennines and to the South in the Maghrebides, following the slab roll-back (Malinverno and Ryan 1986; Patacca et al. 1990; Doglioni 1991). The Balearic Promontory experienced NW-SE contraction coevally with the NW-SE extension in the Valencia Trough and probably also with the extension in the Algerian Basin (Vergés and Sàbat, 1999). The Balearic Promontory resembles an extensional crustal boudin, but underwent coeval (and syntaxial) horizontal shortening.

A shift of active extension from west to east of Sardinia occurred in the Langhian (ca. 15 Ma; Fig. 18.4.1.7) and produced the Middle Miocene to Present opening of the Tyrrhenian Basin. At the same time, a major tectonic and magmatic reorganization occurred along the northern African margin (Carminati et al., 1998a). The continental collision between the Kabylies and north Africa led to the end of the oceanic accretion in the Algerian Basin. Subsequently, compressive stresses built-up in the whole area, inducing compressive and transpressive reactivation of previous faults (Vergés and Sàbat 1999; Frizon de Lamotte et al. 2000). It has been proposed that continental collision was followed by slab breakoff and consequent lithosphere detachment along the northern African margin (Carminati et al., 1998a). The occurrence of a slab detachment process is supported by tomographic models showing an interruption in the slab subducting beneath northern Africa, by the S-wave velocity model of Fig. 18.4.1.2 that shows no evidence for a subducting slab, by the occurrence of bimodal volcanism and by uplift of the African margin (Carminati et al., 1998a; Coulon et al., 2002; Roca et al., 2004). The breakoff possibly begun in Central Eastern Algeria and subsequently propagated eastwards and westwards (Maury et al., 2000; Coulon et al. 2002). The contractional scenario following the continental collision continues to the present and the convergence between the European and African plates is mainly accommodated along the southern margin of the Algerian Basin (Vannucci et al., 2004), where several ENE-WSW-trending reverse faults such as the fault responsible for the 2003 Boumerdès earthquake are active. This could suggest that a new north-verging accretionary prism will form along the African margin (Roca et al., 2004).

Several explanations for the Langhian eastward jump of extension from west to east

of the Corsica-Sarinia blocks have been proposed. In a first model (Gueguen et al., 1997) it is emphasized that the western Mediterranean backarc setting is comparable with Atlantic and western Pacific backarc basins that show similar large-scale lithospheric boudinage, where parts of earlier orogens have been scattered in the backarc area, like the Central America Cordillera relics that are dispersed in the Caribbean domain. In other words, boudinage is a natural feature in backarc basins (Doglioni et al., 1999) and therefore no major geodynamic, kinematic or tectonic changes are to be sought to explain such structures.

In a second model, the shift of extension is genetically linked to the detachment of the Maghrebian portion of the retreating Apennines-Maghrebides slab (Carminati et al., 1998a, 1998b). According to this interpretation the detachment of part of the slab caused a change in the rollback geometry from radial to eastward directed. This hypothesis is supported by the contemporaneity of the two processes and by the results of numerical models showing that the suggested slab detachment could have driven to a major stress change in the western mediterranean area.

A further model suggests that the eastward jump of extension was induced by the interaction between the subducting slab and the 660 km discontinuity which came into contact in the Langhian (Faccenna et al., 2001). This contact possibly brought to a sudden decrease of the subduction rate and trench migration between 16 and 10 Ma ago and to an eastward jump of extension when fast slab retreat resumed.

Whatever the cause of the shift of extension, the continued rollback of the Apennines slab brought to the formation of the Tyrrhenian basin from the Langhian to the Present (Fig. 18.4.1.8).

The Tyrrhenian architecture and magmatism seem to be directly correlated to the composition, thickness, and the subduction rate of the downgoing subducting lithosphere beneath the Apennines (i.e., continental in the Adriatic and oceanic in the Ionian). Since the volume left by the slab retreat is necessarily filled by the upper mantle, the slab rollback of the Apennines subduction kinematically requires an eastward mantle flow in the Tyrrhenian area (and in western Mediterranean in general), as also revealed by shear wave splitting (Margheriti et al., 2003).

Regardless this mantle flow is the cause or a consequence of the slab rollback, it can account for the progressive eastward rejuvenation and boudinage of the western Mediterranean basins, e.g., from the Provençal to the Tyrrhenian (Gueguen et al., 1997).

The boudins and necks are also asymmetric: the base of the crust and of the lithosphere is in fact shifted several tens of km eastward relative to the topography of the basins and swells (Cella et al., 1998), coherently with a shear between lithosphere and underlying mantle (see also Fig. 18.4.1.3).

Finally, it is worthwhile to emphasize that, paradoxically, the extension affecting most of the western Mediterranean developed in a context of relative convergence between Africa and Europe. However that the maximum amount of North-South Africa/Europe relative motion in the last 23 Ma was about 135 km at the Tunisia longitude, more than five times smaller than the eastward migration of the Apennines arc, which moved eastward more than 700 km during the same time span (Fig. Fig. 18.4.1.6-18.4.1.8). Therefore the eastward migration of the Apennines-Maghrebides arc cannot be considered as a consequence of the relative N-S relative convergence between Africa and Europe, but it is rather a consequence of the Apennines-Maghrebides subduction rollback.
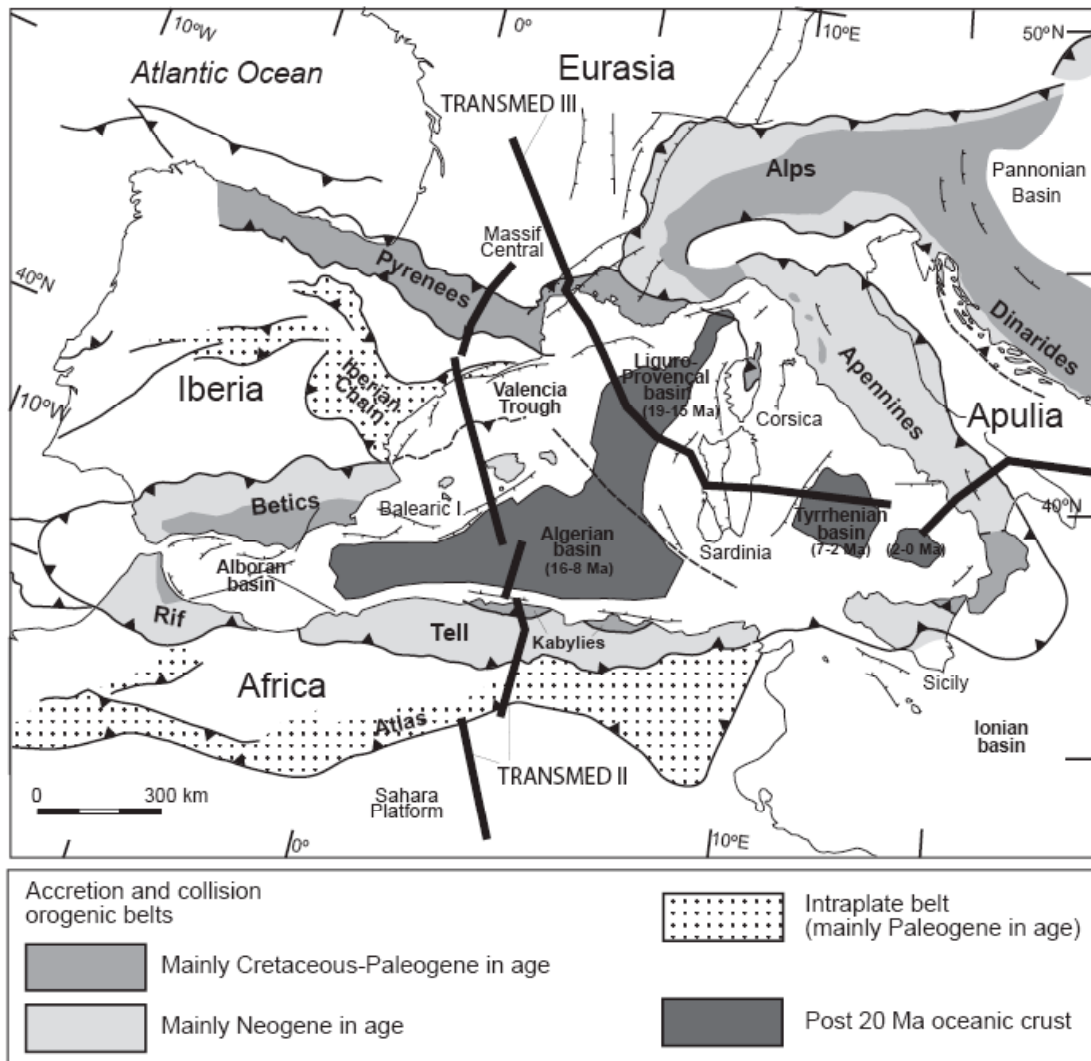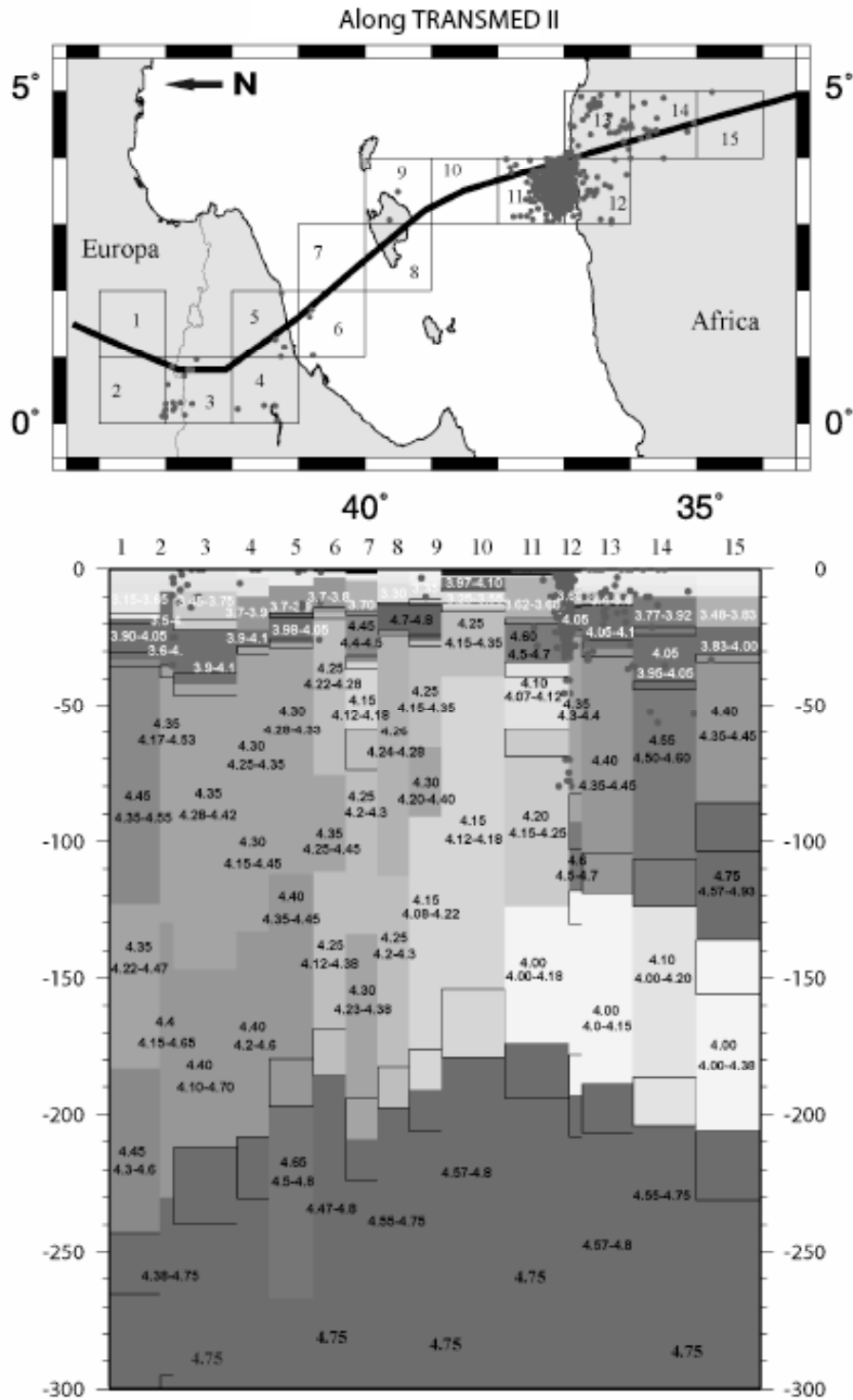
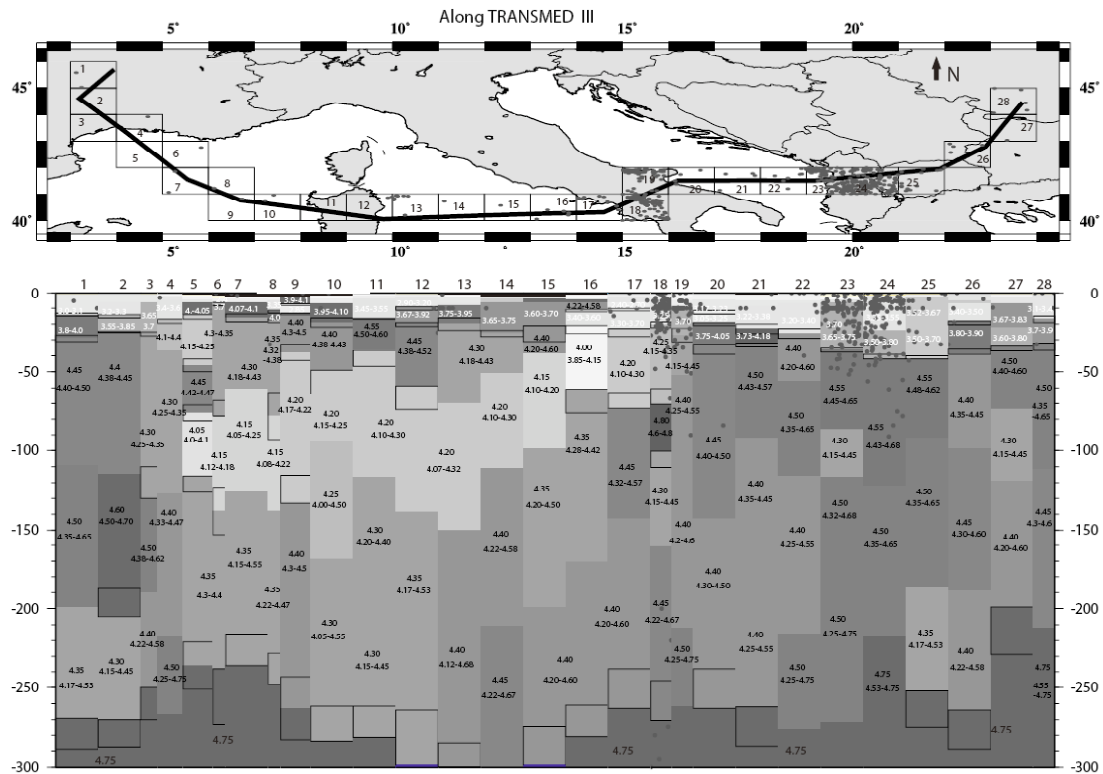Figure 18.4.1.1 - Simplified tectonic map of the western Mediterranean.

Figure
18.4.1.2 - Lithosphere-astenosphere system for geotraverse II: upper part - location of the studied profile; lower part - the cross section obtained from selected solutions and related seismicity (body waves magnitude greater or equal to 3.0). The chosen shear velocity and its range of variability in km/s is printed on each layer. The rectangles outline the range of variability of the layers thicknesses. The velocity of the crustal layers is denoted by white numbers. The hypocenters are denoted by dots.

Figure 18.4.1.3 - Lithosphere-astenosphere system for geotraverse III: upper part - location of the studied profile; lower part - the cross section obtained from selected solutions and related   seismicity (body waves magnitude greater or equal to 3.0). The chosen shear-wave velocity and its range of variability in km/s is printed on each layer. The rectangles outline the range of variability of the layers thicknesses. The velocity of the crustal layers is denoted by white numbers. The hypocenters are denoted by dots.
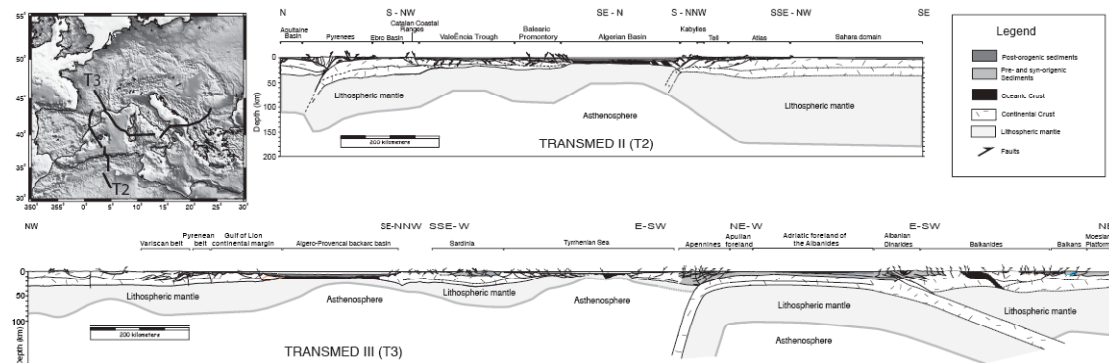

Figure 18.4.1.4 – Lithospheric scale cross sections simplified and redrawn from the Transmed II and III geotraverses (Roca et al., 2004; Carminati et al., 2004)
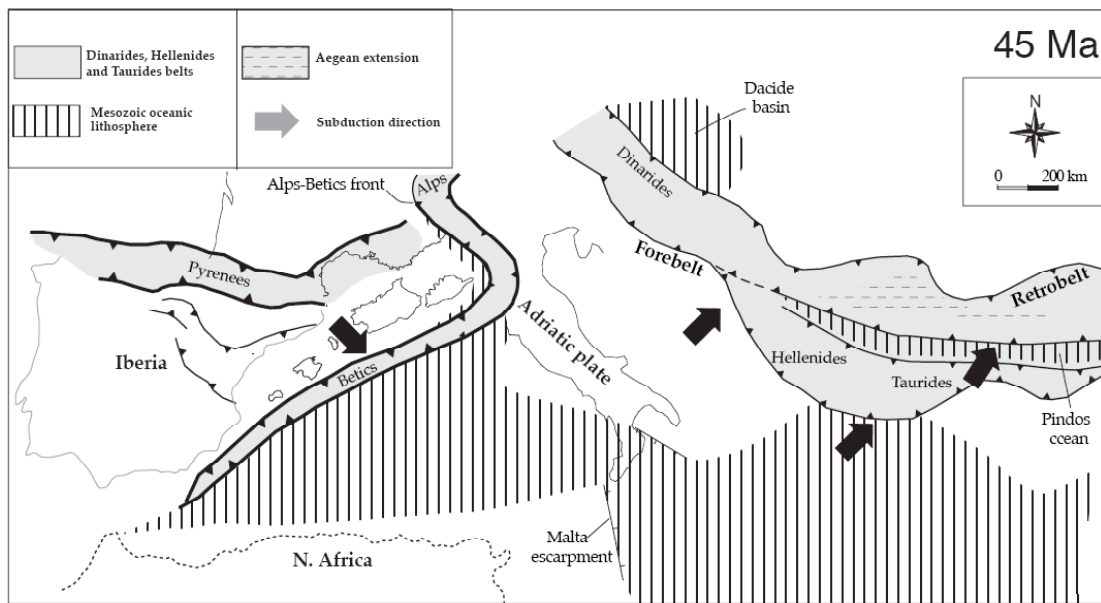
Figure 18.4.1.5 - Mediterranean paleotectonic evolution map at 45 Ma. Modified after Carminati and Doglioni (2004).
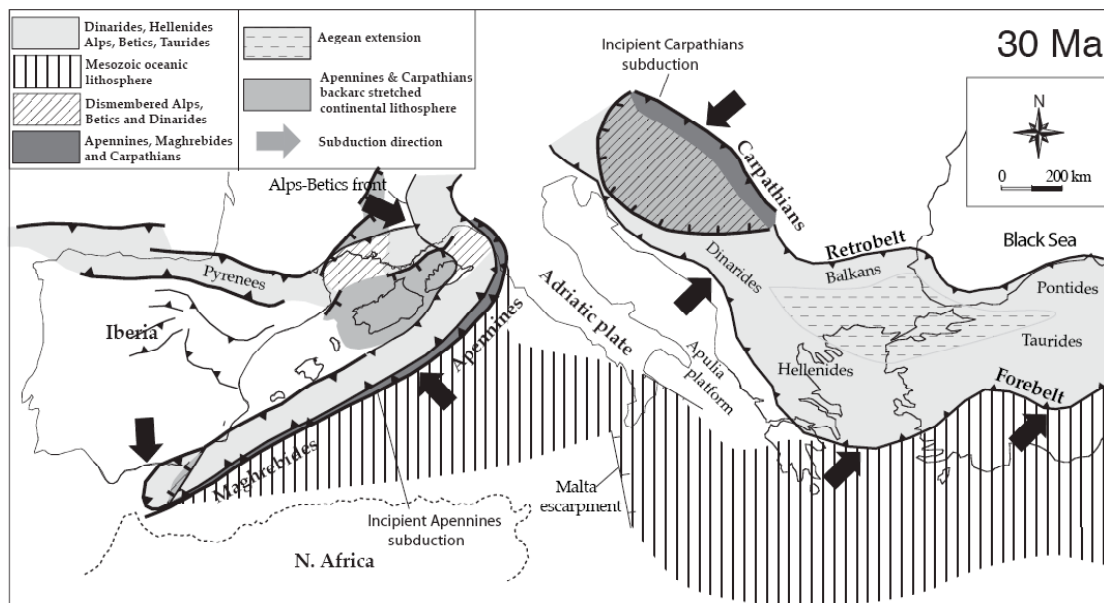

Figure 18.4.1.6 - Mediterranean paleotectonic evolution map at 30 Ma. Modified after Carminati and Doglioni (2004).
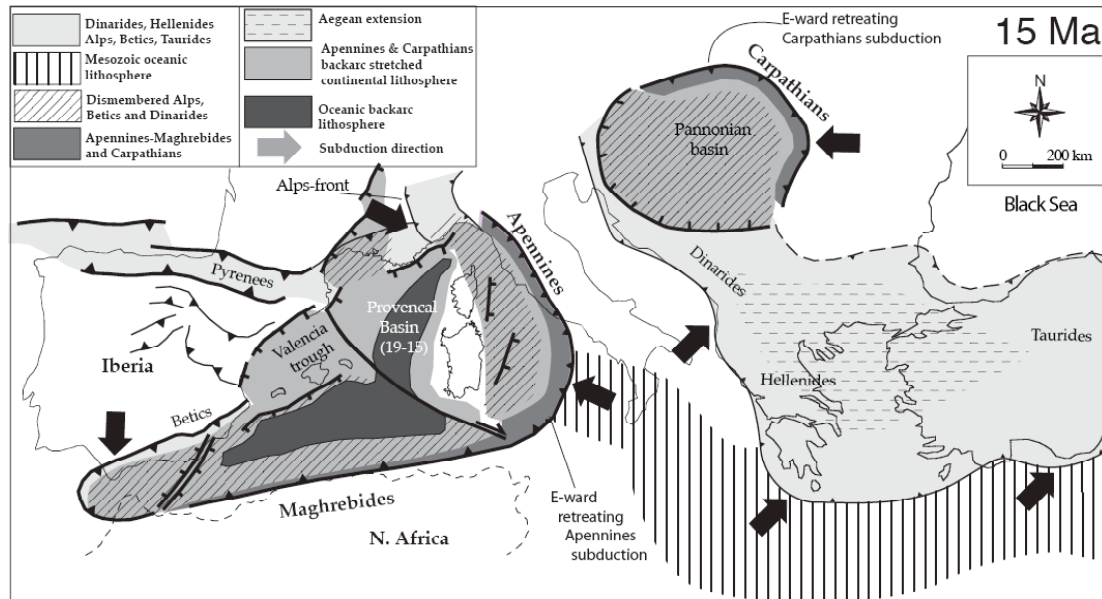
Figure 18.4.1.7 - Mediterranean paleotectonic evolution map at 15 Ma. Modified after Carminati and Doglioni (2004).
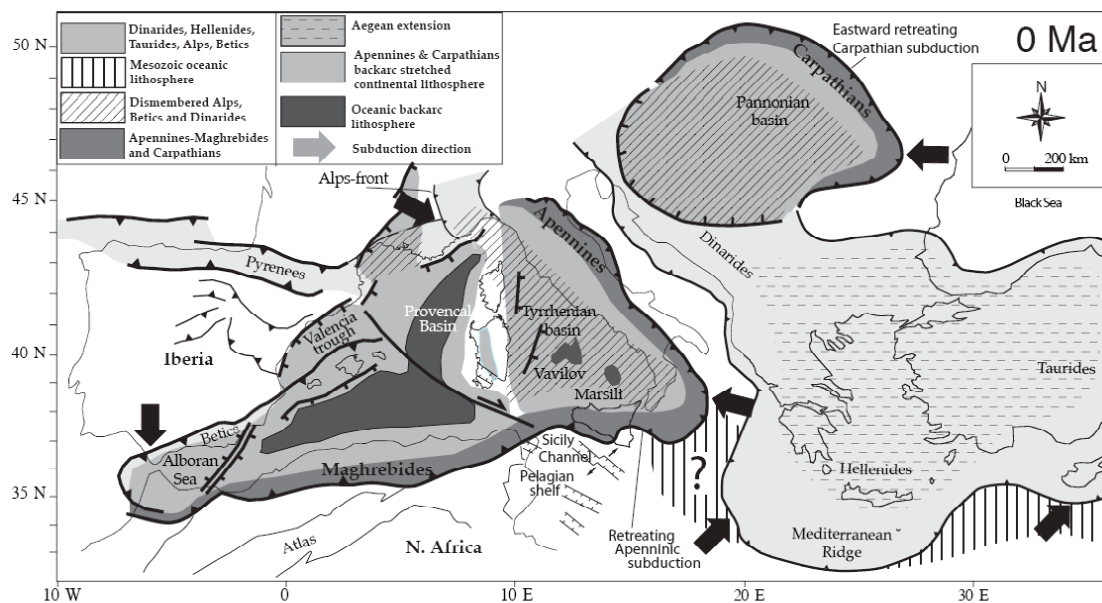

Figure 18.4.1.8 - Mediterranean paleotectonic evolution map present. Modified after Carminati and Doglioni (2004).

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