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The Mediterranean region—a geological primer

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The last twenty-five years of geological investigation of the Mediterranean region have disproved the traditional notion that the Alpine-Himalayan mountain ranges originated from the closure of a single, albeit complex, oceanic domain—the Tethys. Instead, the present-day geological configuration of the Mediterranean region is the result of the creation and ensuing consumption of two major oceanic basins—the Paleotethys and the Neotethys—and of additional smaller oceanic basins within an overall regime of prolonged interaction between the Eurasian and the African-Arabian plates. In greater detail, there is still some debate about exactly what Tethys existed at what time. A consensus exists as to the presence of (i) a mainly Paleozoic paleotethyan ocean north of the Cimmerian continent(s); (ii) a younger late Paleozoic-Mesozoic neotethyan ocean located south of this continent, and finally; (iii) a middle Jurassic ocean, the Alpine Tethys-Valais, an extension of the central Atlantic ocean in the western Tethyan domain. Additional late Paleozoic to Mesozoic back-arc marginal basins along the active Eurasian margin complicated somewhat this simple picture. The closure of these heterogeneous oceanic domains produced a system of connected yet discrete orogenic belts which vary in terms of timing, tectonic setting and internal architecture, and cannot be interpreted as the end product of a single "Alpine" orogenic cycle.

In Neogene time, following prolonged indentation along the Alpine front, a number of small continental microterranes (Kabylies, Balearic Islands, Sardinia-Corsica, Calabria) rifted off the European-Iberian continental margin and drifted toward south or southeast, leaving in their wake areas of thinned continental crust (e.g. Valencia Trough) or small oceanic basins (Algerian, Provençal and Tyrrhenian basins). The E Mediterranean is similarly characterized by widespread Neogene extensional tectonism, as indicated by thinning of continental crust along low-angle detachment faults in the Aegean Sea and the periaegean regions. Overall, Neogene extension in the Mediterranean can be explained as the result of roll-back of the N-dipping subducting slab along the Ionian-E Mediterranean subduction zones. The complex Neogene geologic scenario of the Mediterranean is complicated further by the deposition of widespread evaporites during Messinian (late Miocene) time.

Introduction

Many important ideas and influential geological models have been developed based on research undertaken in the Mediterranean region. For example, the Alps are the most studied orogen in the world, their structure has been elucidated in great detail for the most part and has served as an orogenic model applied to other collisional orogens. Ophiolites and olistostromes were defined and studied for the first time in this region. The Mediterranean Sea has possibly the highest density of DSDP/ODP sites in the world, and extensive research on its Messinian deposits and on their on-land counterparts has provided a spectacular example for the generation of widespread basinal evaporites. Other portions of this region are less well understood and are now the focus of much international attention.

The Mediterranean domain as a whole provides a present-day geodynamic analog for the final stages of a continent-continent collisional orogeny. Over this area, the oceanic lithospheric domains originally present between the Eurasian and African-Arabian plates have been subducted and partially obducted, except for the Ionian basin and the southeastern Mediterranean. The array of interconnected, yet discrete, Mediterranean orogens have been traditionally considered collectively as the result of an "Alpine" orogeny, when instead they are the result of diverse tectonic events spanning some 250 Ma, from the late Triassic to the Quaternary. To further complicate the picture, throughout the prolonged history of convergence between the two plates, new oceanic domains have been formed as back-arc basins either (i) behind active subduction zones during Permian-Mesozoic time, or (ii) possibly associated to slab roll-back during Neogene time, when the advanced stage of lithospheric coupling reduced the rate of active subduction.

This contribution is by no means intended as a thorough description of the geological structure of the Mediterranean region. As an introduction to this special issue of *Episodes*, this paper aims at (i) providing the reader unfamiliar with the geological structure of the Mediterranean with an updated, although opinionated, overview of such complex area, particularly in terms of description of the main geological elements and their paleogeographic-paleotectonic evolution, and (ii) setting the stage for the following articles dealing with various aspects of the geology of Italy. Given the space constraints, fulfilling these tasks clearly involved (over)simplification of a complex matter and in some cases rather drastic choices had to be made among different explanations and/or models proposed by various authors. Similarly, only the main references are cited and the interested reader should refer to the list of references therein for further details on the vast research dedicated to the area. Our sincere apologies to our Mediterranean colleagues for this simplistic synthesis of the magnificently complex geology of their countries.

Overview of present-day Mediterranean geological elements

The present-day geological configuration of the Mediterranean domain is dominated by a system of connected fold-and-thrust belts and associated foreland and back-arc basins (Figure 1). These different belts vary in terms of timing, tectonic setting and internal

Figure 1 Digital terrain model of the Mediterranean region with major, simplified geological structures. White thrust symbols indicate the outer deformation front along the Ionian and eastern Mediterranean subduction fronts. AB, Algerian basin; AS, Alboran Sea; AdS, Adriatic Sea; AeS, Aegean Sea; BS, Black Sea; C, Calabria-Peloritani terrane; CCR, Catalan Coast Range; Cr, Crimea; Ct, Crete; Cy, Cyprus; EEP, East European Platform; HP, High Plateaux; KM, Kirshehir Massif; IC, Iberian Chain; IL, Insubric line; IS, Ionian Sea; LS, Levant Sea; LiS, Libyan Sea; MA, Middle Atlas; MM, Moroccan Meseta; MP, Moesian Platform; PB, Provençal Basin; PaB, Pannonian Basin; PS, Pelagian Shelf; RM, Rhodope Massif; S, Sicilian Maghrebides; SP, Saharan Platform; TA, Tunisian Atlas; TS, Tyrrhenian Sea; VT, Valencia Trough.

architecture (see, for example, Dixon and Robertson, 1984; Ziegler and Roure, 1996) and cannot be interpreted as the end product of a single "Alpine" orogenic cycle (see following section). Instead, the major suture zones of this area have been interpreted as the result of the closure of different oceanic basins of variable size and age. In addition, some Mediterranean foldbelts developed by inversion of intracontinental rift zones (e.g. Atlas, Iberian Chain, Provence-Languedoc, Crimea). The Pyrenees—somehow transitional between these two end members—evolved out of an intercontinental transform rift zone.

The modern marine basins of the Mediterranean Sea (Figure 1) are variably floored by (i) remnants of the Tethyan oceanic domains (Ionian and Libyan seas, E Mediterranean), (ii) Neogene oceanic crust (Algero-Provençal basin and Tyrrhenian Sea), (iii) extended continental lithosphere (Alboran Sea, Valencia Trough, Aegean Sea), and (iv) thick continental lithosphere (Adriatic Sea). (i) In the **Ionian-Libyan Sea** and the **eastern Mediterranean** geophysical data (low heat-flow values and thick lithospheric mantle) and palinspastic reconstructions point to the presence of old (Permian?) oceanic crust underneath a thick pile of Mesozoic and Cenozoic sediments which hampers direct sampling and dating; these two oceanic domains are currently being subducted beneath the Calabria-Peloritani terrane of southernmost Italy (see Bonardi et al., 2001, for a review) and the Crete-Cyprus arcs, respectively. The more than 2,000 m deep **Black Sea** is partly floored by oceanic crust and probably represents the remnant of a complex Cretaceous-Eocene back-arc basin which developed on the upper plate of a north-dipping subduction zone (see following section). The western portion of the Black Sea opened in Cretaceous-Paleocene time whereas the East Black Sea basin has a Paleocene-Eocene age (see Robinson, 1997,

for a review). (ii) The oceanic **Algero-Provençal basin** opened in the Burdigalian, as indicated by paleomagnetic data and by the transition from syn-rift to post-rift subsidence of its margins (Vially and Trémolières, 1996). Rifting in this area occurred as early as the early Oligocene and induced the development of a series of grabens in southern France and Sardinia both on-land and offshore. The deepest portion of the **Tyrrhenian Sea** is floored by Plio-Quaternary oceanic crust; along its western and eastern margins rift-related grabens contain sedimentary deposits as old as ?Serravallian-Tortonian, thus marking the age of the onset of extension in this region (e.g. Kastens et al., 1990; Mattei et al., 2002). (iii) The **Alboran Sea** is floored by thinned continental crust (down to a minimum of 15 km) and it is bounded to the north, west and south by the Betic-Rif orocline. The basement of the Alboran Sea consists of metamorphic rocks similar to those of the Internal Zones of the Rif-Betics (see below). During the Miocene, considerable extension in the Alboran domain and in the adjacent internal zones of the Betic-Rif occurred coevally with thrusting in the more external zones of these mountain belts. Such late-orogenic extension can be interpreted as the result of subduction roll-back toward the west whereby thickened continental crust extends rapidly as the subduction zone retreats (Lonergan and White, 1997; Gutscher et al., 2002). The **Valencia Trough** is floored by thinned continental crust covered by Mesozoic sedimentary deposits; this assemblage underwent extension starting from the late Chattian. Structurally related to the oceanic Provençal basin to the northeast, the Valencia Trough displays younger syn-rift deposits thus indicating a progressive southwestward rift propagation from southern France (Camargue, Gulf of Lions) (Roca, 2001). The **Aegean Sea** is located in the upper plate of the Hellenic subduction zone. Crustal-scale extension in this region has been accommodated

by shallow dipping detachment faults. It has started at least in the early Miocene, and continues today in areas like the Corinth-Patras rift and the southern Rhodope Massif in western Turkey. Miocene extension was accompanied by exhumation of metamorphic rocks and by the intrusion of granitoid and monzonitic magmas at upper crustal levels. According to Jolivet (2001), the engine for Aegean extension is gravitational collapse of a thick crust, allowed by extensional boundary conditions provided by slab retreat; the rather recent tectonic "extrusion" of Anatolia added only a rigid component to the long lasting crustal collapse in the Aegean region. (iv) The **Adriatic Sea** is flooded by 30–35 km thick continental crust whose upper portion is mostly made of a thick succession of Permian-Paleogene platform and basinal carbonates. The Adriatic Sea is fringed to the west and east by the flexural foredeep basins of the Apennines and Dinarides-Albanides, respectively, where several kilometers of synorogenic sediments were deposited during the Oligocene-Quaternary. The Mesozoic Adriatic domain has been considered a continental promontory of the African plate (e.g., Channel et al., 1979; Muttoni et al., 2001); such domain—also known as *Adria*—includes not only what is now the Adriatic Sea but also portions of the Southern Alps, Istria, Gargano and Apulia.

A large wealth of data—including deep seismic profiles, seismic tomographies, paleomagnetic and gravity data, and palinspastic reconstructions—constrains the lithospheric structure of the various elements of the Mediterranean Alpine orogenic system (see Cavazza et al., in press, for a review) and indicates that the late Mesozoic and Paleogene convergence between Africa-Arabia and Europe has totalled hundreds of kilometers. Such convergence was accommodated by the subduction of oceanic and partly continental lithosphere (de Jong et al., 1993), as indicated also by the existence of lithospheric slabs beneath the major fossil and modern subduction zones (e.g. Spakman et al., 1993). Unlike the present-day western and eastern Mediterranean basins, which both still comprise relatively undeformed oceanic crust, the Mediterranean orogenic system features several belts of tectonized and obducted ophiolitic rocks which are located along often narrow suture zones within the allochthon and represent remnants of former ocean basins. Some elements of the Mediterranean-Alpine orogenic system, such as the Pyrenees and the Greater Caucasus, may comprise local ultramafic rock bodies but are devoid of true ophiolitic sutures despite the fact that they originated from the closure of oceanic basins.

The **Pyrenees** are characterized by a limited crustal root, in agreement with a small lithospheric contraction during the late Senonian-Paleogene Pyrenean orogeny. Other Alpine-age Mediterranean chains (western and eastern Carpathians, parts of the Apennines) are also characterized by relatively shallow crustal roots and by a Moho which shallows progressively toward their internal zones. Such geometry of the Moho probably results from the extensional collapse of the internal parts of these orogens, involving structural inversion of thrust faults and lower-crust exhumation on the footwalls of metamorphic core complexes. In spite of differences in terms of chronology and structural style, the Pyrenees are physically linked to the Languedoc-Provence orogen of southern France and—ultimately—to the western Alps.

The **Alps** are the product of continental collision along the former south-dipping subduction zone between the Adriatic continental domain of the African plate to the south and the southern continental margin of the European-Iberian plate to the north. The lithosphere is thicker (ca. 200 km) in the western Alps, while it is in the order of 140 km along the central and eastern Alps (see Dal Piaz et al., this issue, and contributions in Pfiffner et al., 1996, and Moores and Fairbridge, 1997, for an introduction to the Alps). This supports the notion that collisional coupling was stronger to the west. In fact, the eastern Alps are largely made up of tectonic units derived from Apulia, the Austroalpine nappes, while the western Alps are exclusively made up by more external, and tectonically lower units of the European margin, the Briançonnais terrane and the intervening oceanic units (see Piccardo, this issue). The western Alps include outcrops of blueschists and coesite-bearing, eclogite-facies rocks formed at pressures of up to 30 kbars at depths which may have reached 100 km

(see Compagnoni, this issue). Such rocks have yielded radiometric ages as old as 130 Ma, although widespread Eocene metamorphic ages constrain—along with other structural and stratigraphic data—the timing of the collision.

The Alps continue eastward into the **Carpathians**, a broad (ca. 1,500 km long) arcuate orogen which extends from Slovakia to Romania through Poland and Ukraine. To the south, the Carpathians merge with the east-west-trending, north-verging Balkanides through a complex north-trending wrench system. Three major tectonic assemblages are recognized (see, for example, Royden and Horvath, 1988): the Inner Carpathians, made of Hercynian basement and Permian-lower Cretaceous rocks; the tectonic *mélange* of the Pieniny Klippen Belt; and the Outer Carpathians, a stack of rootless nappes made of early Cretaceous to early Miocene turbidites. All these units are thrust towards the foreland and partly override shallow-marine/continental deposits of the foredeep. Two distinct major compressive events are recognized (e.g., Ellouzi and Roca, 1994): thrusting of the Inner Carpathians took place at the end of the Early Cretaceous, while the Outer Carpathians underwent thrusting in the late Oligocene-Miocene. The present-day arcuate shape of this complex mountain belt is mostly the product of Neogene eastward slab retreat (e.g. Linzer, 1996) and displacements along shear zones. The recent seismic activity in the Romanian sector of the Carpathians—the most severe seismic hazard in Europe today—is inferred to be the final expression of such slab roll-back.

The **Balkanides** are an east-west-trending, north-verging thrust belt located between the Moesian Platform to the north and the Rhodope Massif to the south. Underneath the Black Sea, the Balkanides continue with a NW-SE trend. From north to south, three domains can be recognized: the ForeBalkan, i.e., foredeep deposits deformed during late stages of the orogeny, Stara Planina (Balkans s.s.), and Srednogorie. According to Doglioni et al. (1996), the Balkanides can be viewed as the back-thrust belt of the Dinaric-Hellenic subduction and they formed through transpressional inversion of a Jurassic-Cretaceous basin during Paleogene time. Nevertheless, the Balkanides have incorporated much older structures dating back at least to the Early Cretaceous (see Georgiev et al., 2001).

The stable Adriatic (Apulian) platform is flanked to the east by the **Dinarides-Albanides** which continue to the south into the **Hellenides**. Here orogenic activity began during the late Jurassic and persisted until the Neogene. The Dinarides-Albanides-Hellenides are a fairly continuous orogenic belt connected with the southern Alps to the north. It derives from the collision in the Tertiary between the Adriatic promontory and the Serbo-Macedonian-Rhodope block(s). Ophiolites are widespread and crop out along two parallel belts; these ophiolites were obducted in the late Jurassic and then involved in the Alpine collision from the Paleogene. The west-verging Albanides are characterized by thin-skinned thrust sheets which are detached from their basement at the level of Triassic evaporites. This area is the birthplace of the now abandoned concept of geosyncline, elaborated by Aubouin and co-workers in the 1960s.

The **Apennines** of Italy feature a series of detached sedimentary nappes involving Triassic-Paleogene shallow water and pelagic, mostly carbonate series and Oligocene-Miocene turbidites, deposited in an eastward migrating foreland basin. A nappe made of ophiolitic *mélange* (Liguride unit) is locally preserved along the Tyrrhenian coast. The Apennines have low structural and morphological relief, involve crustally shallow (mainly sedimentary Mesozoic-Tertiary) rocks, and have been characterized by widespread extension in their rear portion. The Apennines were generated by limited subduction of the Adriatic sub-plate toward the west. [See Elter et al. (this issue) and Vai and Martini (2001), for further details].

The rock units of both the **Betic Cordillera** of Spain and the **Rif** of northern Morocco have been traditionally subdivided into External Zones, Internal Zones and Flysch nappes (e.g., Lonergan and White, 1997). In the Betic Cordillera, the Internal Zone is made of Mesozoic-Tertiary sedimentary rocks deposited on the Iberian margin of the Alpine Tethys (see following section) and deformed by NW-directed, thin-skinned thrusting during the early-middle

Miocene. The Internal Zone to the south consists of Paleozoic-Mesozoic rocks affected by Paleogene-early Miocene regional metamorphism. The Internal Zone of the Rif belt contains metamorphic rocks broadly similar to those of its counterpart in the Betics. The intermediate Flysch nappes to the south consist of Early Cretaceous to early Miocene deep-marine clastics, whereas the External Zone further south consists of Mesozoic-Tertiary sedimentary rocks deposited on the African margin. Starting from the early Miocene, the Internal Zone was thrust onto the Flysch nappes, followed by the development of a thin-skinned fold-and-thrust belt in the External Zone.

The **Tell** of Algeria and the Rif are parts of the Maghrebides, a coherent mountain belt longer than 2,500 km running along the coasts of NW Africa and the northern coast of the island of Sicily, which belongs geologically to the African continent (see Elter et al., this issue, for an outline of the Sicilian Maghrebides). The Tell is mostly composed of rootless south-verging thrust sheets mainly emplaced in Miocene time. The internal (northern) portion of the Tell is characterized by the Kabylies, small blocks of European lithosphere composed of a Paleozoic basement complex nonconformably overlain by Triassic-Eocene, mostly carbonate rocks.

Two major mountain belts characterize the geological structure of Turkey: the Pontides and the Taurides. The **Pontides** are a west-east-trending mountain belt traceable for more than 1,200 km from the Strandja zone at the Turkey-Bulgaria border to the Lesser Caucasus; they are separated from the Kirsehir Massif to the south by the Izmir-Ankara-Erzincan ophiolite belt. The Pontides display important lithologic and structural variations along strike. The bulk of the Pontides is made of a complex continental fragment (Sakarya Zone) characterized by widespread outcrops of deformed and partly metamorphosed Triassic subduction-accretion complexes overlain by early Jurassic-Eocene sedimentary rocks. The structure of the Pontides is complicated by the presence of a smaller intra-Pontide ophiolite belt marking the suture between an exotic terrane of Laurasian affinity (the so-called Istanbul Zone) and the remainder of the Pontides. The Istanbul zone has been interpreted as a portion of the Moesian Platform which, prior to the Late Cretaceous opening of the west Black Sea, was situated south of the Odessa shelf and collided with the Anatolian margin in the early Eocene (Okay et al., 1994). The **Taurides** are made of both allochthonous and, subordinately, autochthonous rocks. The widespread allochthonous rocks form both metamorphic and non-metamorphic nappes, mostly south-vergent, emplaced through multiphase thrusting between the Campanian and the ?Serravallian (Sengor, 1997). The stratigraphy of the Taurides consists of rocks ranging in age from Cambrian to Miocene, with a characteristic abundance of thick carbonate successions.

Most syntheses of the geology of the Mediterranean region have focused on the orogenic belts and have largely disregarded the large marginal intraplate rift/wrench basins located along the adjacent cratons of Africa-Arabia and Europe, ranging in age from Paleozoic to Cenozoic. Peritethyan extensional basins are instead key elements for understanding the complex evolution of this area, as their sedimentary and structural records document in detail the transfer of extensional and compressional stress from plate boundaries into intraplate domains (see contributions in Roure, 1994, and Ziegler et al., 2001). The development of the peritethyan rift/wrench basins and passive margins can be variably related to the opening of the Tethyan system of oceanic basins and the Atlantic and Indian oceans (see following section). Some of these basins are still preserved whereas others were structurally inverted during the development of the Alpine-Mediterranean system of orogenic belts or were ultimately incorporated into it. Examples of inversion include the **Iberian Chain** and **Catalonian Coast Range** (Figure 1) which formed during the Paleogene phases of the Pyrenean orogeny through inversion of a long-lived Mesozoic rift system which developed in discrete pulses during the break-up of Pangea, the opening of the Alpine Tethys and the north Atlantic Ocean (Salas et al., 2001). The Mesozoic rift basins of the **High Atlas** of Morocco and Algeria underwent a first mild phase of inversion during the Senonian followed by more intense deformation during the late Eocene. Frizon de

Lamotte et al. (2000) have interpreted the latter, main inversion phase as the result of far-field stress transfer from the north during initiation of northward subduction along the southern margin of Iberia and contemporaneous development of the Rif-Tell accretionary prism. Increased coupling between the prism and the African continental margin induced a third phase of inversion in the Quaternary.

A paleogeographic-paleotectonic scenario for the evolution of the Mediterranean domain

Plate-motion vectors are essential elements to understand the geological evolution of the Mediterranean region and to constrain paleogeographic-paleotectonic reconstructions. In short, during late Jurassic-early Cretaceous time, relative motion between Africa-Arabia and Europe was dominated by sinistral strike-slip related to the progressive opening of the central Atlantic Ocean. Since Senonian times Africa-Arabia converged toward Eurasia in a N-S-directed counterclockwise rotational mode. Such overall sinistral motion decreased through time and ceased at the Paleocene-Eocene transition in conjunction with the opening of the Norwegian-Greenland Sea (Ziegler, 1988, 1990). During the Oligo-Miocene, a dextral component is evident in the convergence; such pattern has probably continued until the present. According to Mazzoli and Helman (1994), the relative motion path of the African plate with respect to the European plate from the Oligocene to the Recent can be divided into three phases: (1) NNE-directed during Oligocene to Burdigalian time (up to anomaly 5C: 16.2 Ma), (2) NNW-directed from Langhian to early Tortonian time (16.2–8.9 Ma, anomalies 5C to 5), (3) NW-ward from the late Tortonian (8.9–0 Ma, anomaly 5 to present).

Development of paleogeographic-paleotectonic maps has considerably advanced our understanding of the evolution of the Mediterranean orogenic system and the sedimentary basins associated with it. Yet, uncertainties persist among the various reconstructions proposed (cf. Ziegler, 1988; Dercourt et al., 1993, 2000; Yilmaz et al., 1996). A discussion of the various hypotheses proposed for the evolution of the western Tethyan domain goes beyond the purpose of this contribution. We provide here a brief summary of the post-Variscan evolution of the Mediterranean domain following the paleogeographic reconstructions presented in Stampfli et al. (2001a, b) and refer the interested reader to the abundant literature available on the subject.

Following the late Carboniferous-early Permian assemblage of Pangea along the Variscan-Appalachian-Mauritanian-Ouachita-Marathon and Uralian sutures, a wedge-shaped ocean basin widening to the east—the Paleotethys—was comprised between Eurasia and Africa-Arabia. At this time, global plate rearrangement induced the collapse of the Variscan orogen and continued northward subduction of Paleotethys beneath the Eurasian continent (e.g. Vai, 2003). A new oceanic basin—the Neotethys—began to form along the Gondwanian margin due to the rifting and NNE-ward drifting of an elongate block of continental lithosphere, the Cimmerian composite terrane (Sengor, 1979, 1984). The Cimmerian continent progressively drifted to the northeast, leaving in its wake a new ocean—the Neotethys (Figure 2). The Permo-Triassic history of this part of the world is hence characterized by progressive widening of Neotethys and contemporaneous narrowing of Paleotethys, culminating with final docking of the Cimmerian terrane along the Eurasian continental margin in the late Triassic (although portions of the Paleotethys closed as early as the late Permian). The Cimmerian collisional deformation affected a long yet relatively narrow belt extending from the Far East to SE Europe (see Sengor, 1984, for a discussion). Cimmerian tectonic elements are clearly distinguishable from the Far East to Iran, whereas they are more difficult to recognize across Turkey and SE Europe, where they were overprinted by later tectonism. The picture is complicated by back-arc oceanic

Figure 2 Paleogeographic reconstruction of the western Tethyan area during the late Permian (from Stampfli et al., 2001b, with minor modifications). Stars indicate magmatic activity.

basins (Halstatt-Meliata, Maliac, Pindos, Crimea-Svanetia and Karakaya-Küre) which formed along the southern margin of Eurasia during subduction of Paleotethys and were mostly destroyed when the docking of the Cimmerian continent occurred.

The multi-phased Cimmerian collisional orogeny marked the maximum width of the neotethyan ocean, which during Jurassic-Paleogene time was then progressively consumed by northward subduction along the southern margin of the Eurasian plate (Figure 3). Whereas the Paleotethys was completely subducted or incorporated in very minor quantities in the paleotethyan suture, remnants of the Neotethys are possibly still present in the Ionian Sea and the Eastern Mediterranean. Throughout the Mesozoic new back-arc marginal basins developed along the active Eurasian margin. Some of these back-arc basins are still preserved today (Black Sea and Caspian Sea) but most (e.g. Vardar, Izmir-Ankara) were closed, and the resulting sutures mask the older suture zones of the two main paleotethyan and neotethyan oceanic domains.

The picture is further complicated by the Valais-Pyrenean rift zone which started to develop in the early Jurassic as an eastward extension of the central Atlantic, detaching Iberia from Europe (Figure 3, Aptian), and closed by late Eocene time to form the Alps-Carpathians orogenic system (Figure 3, Eocene-Oligocene boundary) (Stampfli et al., 2002). Mid-Jurassic opening of the Ligurian-Piedmont-south Penninic ocean resulted in the development of a new set of passive margins which were traditionally considered for a long time as segments of the northern margin of a single "Tethyan Ocean" stretching from the Caribbean to the Far East. It is somehow a paradox that the Alps—which for almost a century served as an orogenic model for the entire Tethyan region—are actually related to neither

paleotethyan nor neotethyan evolution and instead have their origin in the Atlantic Ocean to the west.

Paleogene collision along the Alpine front *sensu stricto* induced progressive collisional coupling of the evolving orogenic wedge with its forelands, as well as lateral block-escape and oblique motions. For example, eastward directed orogenic transport from the Alpine into the Carpathian domain during the Oligo-Miocene was interpreted as a direct consequence of the deep indentation of Adria into Europe (Ratschbacher et al., 1991) although this process may have been driven by roll-back and detachment of the westward-dipping subducting slab (Wortel and Spakman, 2000). From a wider perspective, strain partitioning clearly played a major role in the development of most of the Mediterranean orogenic wedges as

Figure 3 Paleogeographic reconstructions of the western Tethyan area during the Aptian, Maastrichtian and at the Eocene/Oligocene boundary. Note the progressive narrowing and suturing of the oceanic domains comprised between the Eurasian and Iberia continental blocks to the north and the Africa/Arabia continent to the south (from Stampfli et al., 2001b, with minor modifications).

major external thrust belts parallel to the former active plate boundaries coexist with sub-vertical, intra-wedge strike-slip faults which seem to have accommodated oblique convergence components (e.g. Insubric line of the Alps, intra-Dinarides peri-Adriatic line).

In spite of prolonged indentation along the Alpine front, the Neogene of the Mediterranean region is characteristically dominated by widespread extensional tectonism. A number of small continental microterranes (Kabylies, Balearic Islands, Sardinia-Corsica, Calabria) rifted off the European-Iberian continental margin and drifted

(late Miocene) time. Such evaporites and—to a lesser extent—the associated post-evaporitic siliciclastics have been the focus of much attention and debate; this section summarizes some salient geological data collected at sea and on land in order to interpret the boundary conditions leading to their deposition. The literature available on this subject is abundant; only a few references are reported here.

During Messinian time, convergence between the African and Eurasian plates, associated with glacioeustatic sealevel falls, isolated the Mediterranean Sea from the world ocean, the basin episodically

Figure 4 Schematic maps showing the paleotectonic evolution of the W Mediterranean during Neogene time (modified after Bonardi et al., 2001, and Roca, 2001). Only active tectonic elements are shown. White, exposed land; light gray, epicontinental sea; darker gray, oceanic crust. Black arrows indicate the direction of Africa's motion with respect to Europe (from Mazzoli and Helman, 1994). White arrows indicate upper-plate direction of extension. Stars indicate subduction-related magmatism. AP, Apennines; B, Balearic block; C, Calabria-Peloritani terrane; K, Kabylies; PB, Provençal Basin, S, Sardinia; TB, Tyrrhenian Basin.

toward the south or southeast, leaving in their wake areas of thinned continental crust (e.g. Valencia Trough) or small oceanic basins (Algerian, Provençal and Tyrrhenian basins) (Figure 4). The E Mediterranean is similarly characterized by widespread Neogene extensional tectonism, as indicated by thinning of continental crust along low-angle detachment faults in the Aegean Sea and the peri-aegean regions (see Durand et al., 1999, and references therein). Overall, Neogene extension in the Mediterranean can be explained as the result of roll-back of the subducting slabs of the Ionian-Apenines-E Mediterranean subduction zone (e.g. Malinverno and Ryan, 1986). As pointed out by Royden (1993), rapid extension of thickened crust in a convergent setting is a consequence of subduction roll-back. During the late stages of orogenesis, Neogene mountain belts throughout the Mediterranean region are characterized by contemporaneous shortening in the frontal portion of the orogenic wedge and extension in its rear portions (e.g. Patacca et al., 1993).

Seismic tomographic models of the upper mantle velocity structure of the Mediterranean-Carpathian region (e.g. Wortel and Spakman, 2000; Panza et al., this issue) point to the important role played by slab detachment, in particular by lateral migration of this process along the plate boundary, in the lithosphere dynamics of the region over the last 20–30 Ma. If the viewpoint provided by this method is accepted, it provides a comprehensive explanation not only of arc-trench migration but also of along-strike variations in vertical motions, stress fields and magmatism. From this viewpoint, slab detachment represents the terminal phase in the gravitational settling of subducted lithosphere.

The Messinian salinity crisis

The complex Neogene geologic context of the Mediterranean region, characterized by the advanced stage of collisional coupling between the Eurasian and the African plates, is further complicated by an important episode of evaporitic deposition during Messinian

desiccated, and large volumes of evaporites precipitated on the floor of what had been a deep marine basin, as well as on its marginal, shallower portions (see Ryan et al, 1973; Kastens et al., 1990; and references therein for a thorough review) (Figure 5). Messinian evaporitic deposition did not occur in a single large depression, but in a series of discrete basins delimited by local barriers and different in form and dimensions from the large pre-Messinian basins, in which hemipelagic facies were associated with open marine conditions. Somewhat overshadowed by the spectacular sea-level event is the fact that the Messinian was also a period of widespread albeit short-lived tectonic activity—the so-called *intra-Messinian tectonic phase*—along the contractional fronts active at the time, at least from Sicily and the Italian peninsula to Corfù, Crete and Cyprus, with thrusting, deposition of syntectonic coarse-grained sediments (including reworked evaporites), and development of widespread angular unconformity and disconformities (e.g. Decima and Wezel, 1973; Montadert et al., 1977; Vai and Ricci Lucchi, 1977; DeCelles and Cavazza, 1995; Cavazza and DeCelles, 1998; Butler et al., 1995).

Figure 5 Areal extent of the Messinian evaporites in the Mediterranean region. Modified after Rouchy (1980).

Astronomically calibrated high-resolution stratigraphy (Krijgsman et al., 1999) shows that the onset of the Messinian salinity crisis is synchronous over the entire Mediterranean basin, dated at 5.96 ± 0.02 Ma. This is in contrast with the magnetostratigraphic results of Butler et al. (1999), indicating that on a much smaller area (within the foreland basin to the south of the Sicilian Maghrebides) the beginning of evaporite precipitation is diachronous over a period of at least 800 ka.

The well-exposed Messinian outcrops of central Sicily provide one of the thickest and most complete occurrences of this stage and have been instrumental in the development of current thinking on the Mediterranean evaporites (Figure 6). Hereafter we provide a short description of the stratigraphy of this area as an example of the complexities of the Messinian stratigraphy. At the periphery of the basin the Lower Evaporites—i.e. the Messinian succession below the

Figure 6 Schematic stratigraphic cross section of the Messinian of Sicily (modified after Decima and Wezel, 1973). 1) Pre-evaporitic clay, marl and diatomite (Tripoli Fm.); 2) evaporitic limestone (Calcare di Base); 3) lower gypsum beds (Gessi di Cattolica); 4) gypsum turbidites; 5) halite and potash (in black) beds; 6) gypsum arenite; 7) upper gypsum beds (Gessi di Pasquasia); 8) Arenazzolo Fm.; 9) Trubi Fm. (lower Pliocene).

intra-Messinian unconformity—consist only of two relatively thin units (Figure 6): the Tripoli Formation (laminated diatomites) and the Calcare di Base (evaporitic limestone). In the deepest portions of the basin, the Lower Evaporites are much thicker and comprise, from bottom to top, the Tripoli Fm, the Lower Gypsum Fm (LGF), and the Halite Fm (HF). The LGF is composed of up to 300 m of selenite gypsum with random orientation, indicating that gypsum from the periphery was reworked, deposited in deeper water, and recrystallized; its upper parts consists of gypsum turbidites. The HF is made of up to 800 m of halite with intercalations of potash/magnesium salt beds; this unit was deposited in deep depressions, fed also by clastic resedimentation and slumping. Related to intra-Messinian tectonics, slumping began when the gypsum turbidites of the LGF were deposited and reached its acme at the end of the sedimentation of the HF. Subaerial erosion occurred in the marginal zones of the basins at the same time as the strata of salts filled up the deep, subsiding depressions. As the potash beds were covered by halite and anhydrite, there are indications of freshening of the brine during the late stages of salt deposition. It appears that these cannot be easily explained by Hsü's (1972) hypothesis of a "deep, dry basin".

In Sicily the Lower Evaporites close with the HF, whereas at other Italian sites they terminate with a flysch-like, marly-arenaceous deposit (for example, in the Marche Region), which indicates rapid filling of subsiding troughs. Terrigenous sedimentation was accompanied by cinerite deposition. Taken together, these events suggest that the salts are relatively deep marine syn-diastrophic deposits which correspond to a significant phase of marine regression. In Sicily the salts have been affected by intense tectonic com-

pression with diapiric folds (Decima and Wezel, 1973). The Lower Evaporites were thus deposited during widespread regression which created barriers and subdivided the Tortonian depositional area, with the emersion of vast tracts of land, such as the Central Alboran Sea and the northern Tyrrhenian Sea. At the peak of the lowstand a sub-aerial erosional surface developed and resulted in the widespread *intra-Messinian inter-regional discontinuity*, which corresponds to a sequence boundary separating the Lower and Upper Evaporite deposits.

The late Messinian Upper Gypsum Formation (UGF) of Sicily onlaps the underlying intra-Messinian erosional surface. This unit is vertically organized in transgressive-regressive cycles, each characterized by a reduction in depth and an increase in the degree of salinity. The presence of *Ammonia tepida* indicates that the water was hypo-haline and no deeper than about 50 m. The regionally transgressive UGF contains the so-called "*Congerie* fauna", a paleontological assemblage interpreted as indicative of low-salinity conditions and of an eastern European affinity, leading some scientists to infer that the Mediterranean had been a brackish lake or "lago-mare", fed by the influx of vast quantity of freshwater from the Paratethys of eastern Europe (e.g. Hsü et al., 1978). However, in this concept it is unclear whether we are dealing with a giant lake or a series of isolated brackish lakes. The upper evaporites include thick clastic successions that are possibly reflecting an increased continental run-off.

Throughout much of the Mediterranean basin, siliciclastics deposits are invariably concentrated in the uppermost portion of the Messinian succession. In the type area of the Messinian in Sicily, this interval is referred to as the Arenazzolo Formation (Figure 6) (Decima and Wezel, 1973; Cita and Colombo, 1979) but a variety of local names still coexist. Published descriptions depict widely variable lacustrine and fluvial/alluvial facies that formed as the Mediterranean basin was partially inundated towards the end of the Messinian (Decima and Wezel, 1973). However, relatively little detailed information is available concerning this important transitional facies, and little effort has been made to incorporate it into a sequence-stratigraphic framework for the terminal Miocene transgression in the Mediterranean (e.g. Gelati et al., 1987; Roveri et al., 1992; Butler et al., 1995).

The coccolith-foraminiferal marls of the Pliocene Trubi Formation mark the end of the Messinian period of desiccation and the return to normal, open-marine sedimentation in the Mediterranean basin (e.g. Decima and Wezel, 1973; Cita and McKenzie, 1986). Because this lithologic change defines the Miocene-Pliocene boundary stratotype, the Trubi marls have been intensively studied (e.g. Cita and Gartner, 1973; Hilgen, 1987; Channell et al., 1988; Rio et al., 1991). A few occurrences of pre-Trubi marine faunas have been reported in the past (see Benson and Rakic-El Bied, 1995, for a review), and were discarded possibly because they challenged the widely accepted notion of the "Zanclean deluge," which is conceived as a virtually synchronous flooding of the Mediterranean basin. This "deluge" is thought to be marked by the base of the Trubi Formation, providing a convenient datum for the formal establishment of the base of the Pliocene (Van Couvering et al., 2000).

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