Chapter 1

The Mediterranean Area and the Surrounding Regions: Active Processes, Remnants of Former Tethyan Oceans and Related Thrust Belts

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Abstract

The Mediterranean domain provides a present-day geodynamic analog for the final stages of a continent-continent collisional orogeny. Over this area, 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. A number 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 Myr, 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*) associated with slab roll-back during Neogene time, when during advanced stages of lithospheric coupling the rate of active subduction was reduced. The closure of these heterogenous oceanic domains produced a system of discrete orogenic belts which vary in terms of timing of deformation, tectonic setting and internal architecture, and cannot be interpreted as the end product of a single Alpine orogenic cycle.

Similarly, the traditional paleogeographic notion of a single – albeit complex – Tethyan ocean extending from the Caribbean to the Far East and whose closure produced the Alpine-Himalayan orogenic belt must be discarded altogether. Instead, the present-day geological configuration of the Mediterranean region is the result of the opening and subsequent consumption of two major oceanic basins – the Paleotethys (mostly Paleozoic) and the Neotethys (late Paleozoic-Mesozoic) – and of additional smaller oceanic basins, such as the Atlantic Alpine Tethys, within an overall regime of prolonged interaction between the Eurasian and the African-Arabian plates. Paradoxically, the Alps, that have long been considered as the classic example of Tethys-derived orogen, are instead the product of the consumption of an eastward extension of the central Atlantic ocean, the middle Jurassic Alpine Tethys, and of the North Atlantic ocean, the middle Cretaceous Valais ocean.

1.1 Introduction

From the pioneering studies of Marsili – who singlehandedly founded the field of oceanography with the publication in 1725 of the *Histoire physique de la mer*, a scientific best-seller of the time (Sartori 2003) – to the technologically most advanced cruises of the R/V *JOIDES Resolution*, the Mediterranean Sea has represented a crucible of scientific discoveries. Similarly, the peri-Mediterranean terranes on land have been accurately surveyed over the centuries by amateur, academic and industrial geologists alike, and some locations have represented geological training grounds for several generations of Earth scientists.

Although relatively small on a global scale – its area being roughly equivalent to half of that of the People's Republic of China – the Mediterranean region has an exceedingly complex geological structure. For example, tectonic activity here spans from the Panafrican orogeny (Precambrian) of the Gondwanan, northern Africa craton to the destructive present-day seismicity along the North Anatolian Fault. Many important ideas and influential geological models were 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 provided a spectacular example for the generation of widespread basinal evaporites. Other portions of this region are less well understood and are now receiving much international attention.

Apart from its historical and cultural importance as a crossroad among various religions, trade routes and civilizations, the Mediterranean constitues also a geological transition between the Middle East and the Atlantic, as well as between Europe and Africa. For example, the Mediterranean represents a proxy of the long-lasting interactions between Eurasia and Gondwana, with successive episodes of continental break-up and oceanic development, subduction, continental collision and orogeny. The Neotethyan oceanic domain initiated during the Permian in the eastern part of the Mediterranean, separating progressively the Gondwanan continental margin to the south from the Eurasian margin to the north. Opening in a scissor-like fashion, the Neotethys widened toward the east where it was connected with the world ocean, whereas it remained closed to the west until the onset of spreading in the central Atlantic during the Jurassic. As a result, oceanic crust in the former Tethyan domains and in currently allochthonous ophiolitic units still preserved in Alpine thrust belts display a wide range of ages, depending on whether they are derived from the Neotethys (in areas already oceanic before the Jurassic) or from younger Piedmont-Ligurian-Carpathian oceanic domains once connected with the Atlantic (see Stampfli and Borel, this volume). The present-day Mediterranean Sea is a composite array of oceanic and continental domains of various affinities, including remnants of the Neotethys in the easternmost Mediterranean and the Ionian Sea, neoformed oceanic basins in the western Mediterranean, and shallow epicontinental seas like the Adriatic.

This contribution is by no means intended as a thorough description of the geological structure of the Mediterranean region. This and the following two chapters by Spakman and Wortel and by Stampfli and Borel aim at (*i*) providing the reader unfamiliar with the Mediterranean domain with an updated, yet opinionated, overview of such complex area, particularly in terms of description of those geological-geophysical features which the authors deem representative, and (*ii*) setting the stage for the TRANSMED CD-ROM which contains detailed information on the majority of the most significant areas of the Mediterranean. 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 CD for further details on the vast research dedicated to the area. In the search for clarity and conciseness we plead guilty of deliberate simplifications and omissions.

1.2 Mediterranean Fold-and-thrust Belts

The Mediterranean domain is dominated geologically by a system of connected fold-and-thrust belts and associated foreland and back-arc basins (Fig. 1.1). These belts cannot be interpreted as the end product of a single "Alpine" orogenic cycle as they vary in terms of timing, tectonic setting and internal architecture (see, for example, Dixon and Robertson 1984; Ziegler and Roure 1996). Instead, the major suture zones of this area are the result of complex tectonic events which closed different oceanic basins of variable size and age (see Stampfli and Borel, this volume). 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 a continental transform rift zone.

A large wealth of data – including deep seismic soundings, seismic tomographies, paleomagnetic and gravity data, and palinspastic reconstructions – constrains the lithospheric structure of the various elements of the Mediterranean Alpine orogenic system and indicates that the late Mesozoic and Paleogene convergence between Africa-Arabia and Europe has totalled hundreds of kilometers. Such convergence was accomodated 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; Wortel and Spakman 2000; Spakman and Wortel, this publ.). 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 extensional basins. Some elements of the Mediterranean orogenic system, such as the Pyrenees and the Greater Caucasus, may comprise local ultramafic rock bodies but are devoid of true ophiolitic sutures. Following is a very concise description – largely abstracted from the text accompanying the TRANSMED transects – of the main fold-and-thrust belts of the Mediterranean orogenic system.

The **Pyrenees** run between the Bay of Biscay and the Gulf of Lion for a length of about 450 km and a width of 50–100 km. In spite of some 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. Overall, the Pyrenees are characterized by an asymmetric and bivergent, V-shaped upper crustal wedge which developed along the collision zone between the Iberian and the Eurasian plates, where limited continental subduction of the Iberian lower crust and lithospheric mantle underneath the European plate occurred. The northern wedge is formed by a northward directed stack of thrusts; the southern wedge is wider and shows greater displacement and cumulative shortening. The analysis of kinematic indicators point to a convergence nearly orthogonal to the Iberia-Eurasia plate boundary through most of the tectonic evolution of this area (see TRANSMED Transect II). There is no consensus on the exact nature and geometry of the deeper crustal levels beneath the Pyrenees: previous studies (e.g. ECORS Pyrenees Team 1988; Torné et al. 1989) pointed to a narrow and thinner lithospheric root than the one shown by Roca et al. (this publ.). At any rate, the Pyrenees are characterized by a limited

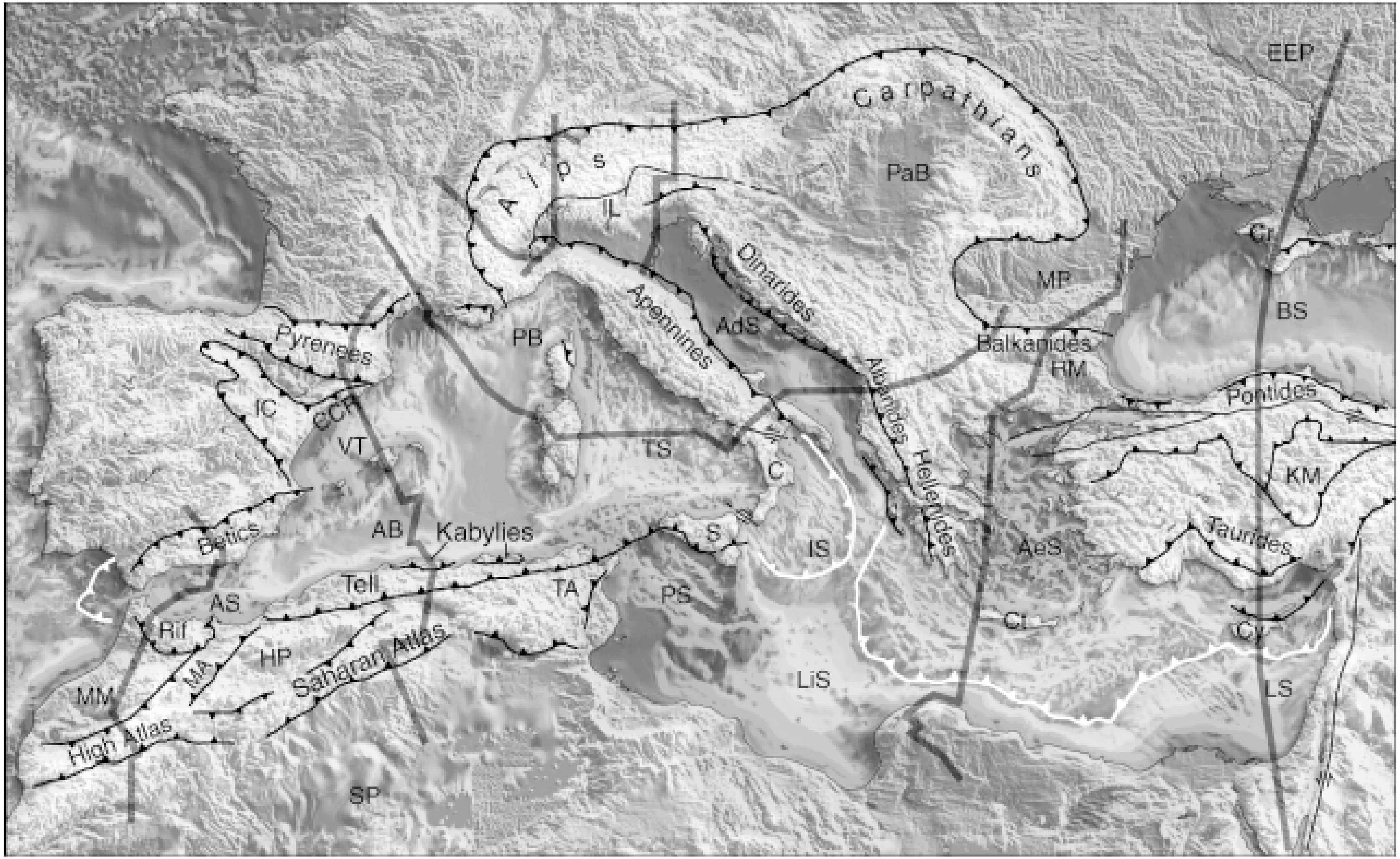

Fig. 1.1. Digital terrain model of the Mediterranean region with major, simplified geological structures. *White thrust symbols* indicate the submarine deformation front along the Ionian and eastern Mediterranean accretionary wedges. Shown are the traces of the eight TRANSMED transects included and discussed in the CD-ROM. *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*, Kirsehir 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 (from Cavazza and Wezel 2003, modified)

crustal root, in agreement with relatively small lithospheric contraction during the late Senonian-Paleogene Pyrenean orogeny. In contrast to the Iberian Moho, the European Moho shallows toward the Axial Zone of the Pyrenees, pointing to a pre-orogenic thinning of the European crust in the vicinity of former Albian rift basins. As pointed out by Ziegler and Roure (1996), some 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. However, unlike the Pyrenees where crustal thinning in the European foreland predates the orogeny, 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.

The Pyrenees originated from the tectonic inversion of Triassic-Cretaceous rift systems that had developed during the fragmentation of southern Variscan Europe and western Tethys in conjunction with the break-up of Pangea, the opening of the central Atlantic and the Bay of Biscay, and the resulting rotation of Iberia (see Roca et al. and Stampfli and Borel, this publ., for a review). Convergence occurred from late Santonian to middle Miocene times as the African-Iberian plates moved generally northward against Europe. The Pyrenees are flanked by two main foreland basins, the Aquitaine basin to the north and the Ebro basin to the south (see TRANSMED Transect II).

From west to east, the **Alps** extend from the Gulf of Genoa to the Vienna basin, where their connection with the Carpathians can be traced only in the subsurface. Based on tectonic vergence, the Alps can be subdivided across tectonic strike into (*i*) a Europe-vergent belt of Cretaceous-Neogene age and (*ii*) the Southern Alps, a subordinate and shallower south-verging fold-and-thrust belt of Neogene age (see Dal Piaz et al. 2003, and contributions in Moores and Fairbridge 1997, for an introduction to the Alps). Overall, the Alpine orogen is highly asymmetric, being volumetrically dominated by north-vergent tectonic structures, whereas the Southalpine structures are surficial (see Pfiffner et al. 1996; Roure et al. 1996; TRANSALP Working Group 2002; Schmid et al., this publ.).

The Alps record the closure of different oceanic basins during the Late Cretaceous and Cenozoic convergence of the African (or Apulian) and European plates. Despite being the birthplace of cylindrism, the Alps display significant along-strike changes in their overall architecture, thus supporting the notion that the oceanic and continental paleogeographic domains from which the Alpine tectonic units derive were arranged in a rather non-cylindrical fashion (see Stampfli and Borel, this publ., and references therein). This is reflected by the areal arrangement of the Alpine terranes, in the deep structure of the Alps, and in the different ages of metamorphism (see Schmid et al., this publ., for a review). For example, the eastern Alps are largely made up of tectonic units derived from Apulia, the Austroalpine nappes, while the western Alps are nearly exclusively composed of more external, and tectonically lower units of the European margin, the Briançonnais terrane (a microcontinent rifted off Europe and separated from it by the narrow Valais ocean, which probably did not extend into the eastern Alps) and the intervening oceanic units. The main metamorphic events are also irregularly distributed in time and space (Frey et al. 1999): Tertiary in the western Alps, Cretaceous in the Austroalpine units of the eastern Alps. The western Alps include outcrops of blueschists and coesite-bearing, eclogite-facies rocks formed at pressures of up to 30 kbar at depths which may have reached 100 km (Compagnoni 2003 and references therein). 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. This supports the notion that collisional coupling was stronger to the west.

Within this context, the Alps can be considered the product of two discrete orogenies: a Cretaceous one related to the closure of an embayment of the Meliata ocean into Apulia, followed by a Tertiary one due to the closure of the Alpine Tethys between Apulia and Europe (Haas et al. 1995; Stampfli et al. 2001a, b). The former affected only what are now the Eastern Alps: this implies that the Austroalpine wedge contains two sutures, namely the Meliata and the Penninic sutures, both of which are associated with HP/LT rocks (Thöni 1999).

The Alps continue eastward into the **Carpathians**, a broad (ca. 1500 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 compressional events are recognized (e.g. Ellouz 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 rollback, involving delamination of the lower mantle-lithosphere from the continental foreland (Sperner et al. 2004).

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 the south, three domains can be recognized: the Fore-Balkan, Stara Planina (Balkans s.s.), and Srednogorie (Georgiev et al. 2001; see TRANSMED Transects III and VII). The Fore-Balkan is the transitional zone between the Balkan thrust belt front to the south and the Moesian platform, i.e. the foreland, to the north. The transitional character is indicated by the lower degree of deformation of the foredeep deposits affected only by the late stages of the orogeny and by the lack of Jurassic-Lower Cretaceous flysch. The Balkans (Stara Planina) are a complex system of north-verging nappes thrust over the South Carpathian units and the southern margin of the Moesian platform. Key characteristics of the Balkans are: (*i*) widespread Mesozoic to Early Tertiary flysch; (*ii*) general lack of products of Alpine magmatic activity; (*iii*) intense mid-Eocene compressional deformations in the central and eastern segments; (*iv*) relatively thick continental crust (38–34 km), gradually thinning toward the Moesian platform. The Balkans include thrust sheets of mid-Cretaceous, Late Cretaceous and Paleogene age. The Srednogorie zone is a segment of a Late Cretaceous magmatic belt that extends from former Yugoslavia, through Romania and Bulgaria into Turkey and the Lesser Caucasus. This belt was interpreted by Boccaletti et al. (1974) as the remnant of a volcanic island arc related to northward subduction of Tethyan oceanic crust beneath the Eurasian continent. The Srednogorie zone in the area crossed by TRANSMED Transects III and VII is characterized by thick Upper Cretaceous volcano-sedimentary successions and numerous intrusive bodies of island-arc signature, and by main Late Cretaceous compressional deformation followed by mid-Eocene north-verging thrusting over the Balkans s.s.

The stable Adriatic (Apulian) platform is flanked to the east by the **Dinarides** which continue to the southeast into the **Albanides** and **Hellenides**. 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 Paleogene Alpine collision (Pamic et al. 2002). The Dinarides-Albanides are bordered to the west by a foredeep lying in the eastern Adriatic basin, filled by Eocene-Quaternary turbiditic sediments (see Frasheri et al. 1996 for the Albanian sector). South of the Scutari-Pec transversal tectonic structure, the Albanides are characterized by thin-skinned thrust sheets which are detached from their basement at the level of Triassic evaporites. The Hellenides are bordered to the southeast by the Antalya convex zone, which separates them from the Taurides in the east. The Dinarides-Hellenides are the birthplace of the now abandoned concept of geosyncline, elaborated by Aubouin and co-workers in the 1960s.

The **Apennines** of Italy are one of the youngest mountain chains in the world as they formed during Neogene-Quaternary time. In addition, during the last 3 Myr the Apennines have experienced significant shortening (80–200 km), uplift (up to 2.5 km) at fast rates (up to about 1 mm a^{-1} in the Apuane Alps, Calabria and the Peloritani Mts. of NE Sicily), and considerable subsidence (up to 5 km) at a relatively fast rate in the related peri-Adriatic foredeeps (e.g. 2.5 mm a^{-1} in the Po plain) (Vai and Martini 2001).

Along their length the Apennines can be separated into two somewhat different arcuate segments: the northern segment is convex – and vergent – toward the northeast, the southern segment is convex – and vergent – toward the southeast. Structurally, the northern segment is a fairly regular orogenic wedge with in-sequence thrusts and significant piggy-back basins; conversely, the southern segment is more complex and has duplexes and widespread out-of-sequence thrusts. The southern segment was tectonically overridden by the Calabria-Peloritani terrane (Bonardi et al. 2001), a fragment of the Alpine Chain drifted off the Corsica-Sardinia block during the opening of the Tyrrhenian Sea (see following section). Only a few outcrops along the Tuscany coast of the Tyrrhenian Sea could suggest the existence of a similar terrane in the northen segment of the Apennines.

The Apennines feature a series of detached sedimentary nappes involving Triassic-Paleogene shallow water and pelagic, mostly carbonate series and late 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 the coexistence during Neogene-Quaternary time of compression in the frontal (northeastern) portion of the orogenic wedge and widespread extension in its rear portion. The Apennines were generated by limited subduction of the Adriatic sub-plate toward the west. See Vai and Martini (2001) and Elter et al. (2003) for further details.

The rock units of the **Betic Cordillera** of Spain and the **Rif** of northern Morocco form together an arc-shaped mountain belt encompassing the Alboran basin and referred to as the the Gibraltar arc (see Frizon de Lamotte et al., this publ.). This orogenic system developed during late Mesozoic to Cenozoic convergence and strike-slip movements between NW Africa and Iberia, and – notwithstanding a rather complex and locally cumbersome tectonostratigraphic nomenclature – it can be broadly subdivided into External Zones, Flysch nappes and Internal Zones. The Guadalquivir and Pre-Rif foreland basins fringe the orogenic system north and south of the Gibraltar arc.

The *External Zones* represent the deformed southern Iberian and northwestern African paleomargins of the Alpine Tethys (see Stampfli and Borel, this publ.) and consist of autochthonous, parautochthonous and/or allochthonous non-metamorphic Mesozoic and Tertiary sedimentary covers which were detached during Miocene time from a Variscan basement, the Iberian and Moroccan Meseta, respectively. The *Flysch nappes* are derived from an oceanic or semi-oceanic basin overlain by Jurassic to Miocene deep-water sediments. These units, now deformed in an accretionary prism presently located in the western Betics and along the northern part of Africa from the Strait of Gibraltar to Tunisia, demonstrate the existence of a long oceanic basin fringing the North African paleomargin. The *Internal Zones* consist mainly of three nappe complexes of variable metamorphic grade, two of which can be traced all along the Gibraltar arc. The Internal Zones record an Alpine-age tectonometamorphic evolution, which includes an early high-pressure low-temperature event. These relics of a former orogenic wedge are now *megaboudins* of the Alpine metamorphic belt, stretched and separated during later extensional episodes (see below); they are the remnants of older subduction processes and are present not only in the Betic-Rif orocline, but also in the Kabylies of Algeria (Caby et al. 2001) and in the Calabria-Peloritani terrane of Italy (Bonardi et al. 2001). Starting from the early Miocene, the Internal Zones were thrust onto the Flysch nappes, followed by the development of a thin-skinned fold-and-thrust belt in the External Zones. Contemporaneous crustal extension affected the center of the Gibraltar orocline, leading to the development of the Alboran Sea (see following section), which is floored by metamorphic rocks of the Internal Zones (Comas et al. 1999).

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. 2003, 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 sedimentary rocks. In the Maghrebian domain, despite the presence of contractional deformation at upper crustal levels, the lithosphere and crust do not show evidence of significant thickening. Gravity and heat-flow models (see TRANSMED Transect II) point to fairly constant crustal and lithospheric thicknesses of 35 km and 175–185 km that decrease rapidly near the coastline to reach low values offshore in Algerian basin. Such configuration could be explained by the presence of a detachment at the base of the upper crust and by the subduction of the underlying levels beneath the Iberian plate. However, mantle tomographic (Spakman et al. 1993), seismological (Buforn et al. 1995), gravity and heat-flow (Roca et al., this publ.) data argue against the presence of a subducting slab along the northern border of Africa. This does not imply that such subduction zone did not exist in the past. In fact, seismic tomography provides evidence for a deep detached slab below the Tell (Carminati et al. 1998b; Piromallo and Morelli 2003), in agreement with the petrological features of Neogene magmatism (e.g. Coulon et al. 2002). From this viewpoint, such detached slab is the remnant of the subduction responsible for the formation of the entire Maghrebides Chain.

Two major mountain belts characterize the geological structure of Turkey: the Pontides and the Taurides (for a review, see Stephenson et al., this publ.). 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 (the so-called Istanbul Zone) and the Sakarya Zone, which in turn is separated from the Taurides to the south by the Izmir-Ankara suture. 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 Paleogene (Okay et al. 1994; Göncüoglu et al. 2000).

The **Taurides** are made of both allochthonous and, subordinately, autochtonous 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. The stratigraphy of the Taurides consists of rocks ranging in age from Cambrian to Miocene, with a characteristic abundance of thick carbonate successions. The Izmir-Ankara suture represents the former plate boundary between the Pontides to the north and the Anatolide-Tauride block to the south, and it is marked by a zone of southward imbrication of Upper Cretaceous and Triassic accretionary complexes (see TRANSMED Transect VIII). The collision of these two continental domains started in the early Tertiary, while the continuing convergence and contractional tectonism persisted until the late Miocene (Okay et al. 2001). The appearance of the North Anatolian transform fault in the late Miocene marks the end of the north-south collision between Eurasia and Africa-Arabia.

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, Stampfli et al. 2001b, and Ziegler et al. 2001b). 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 (Stampfli and Borel, this publ.). 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 (Fig. 1.1) 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. This main inversion phase has been interpreted as resulting either from far-field stress transferred across the oceanic crust of the Maghrebian Tethys in response to its accelerated northward subduction beneath Iberia (Frizon de Lamotte et al. 2000) or from the arrival of an unspecified obstacle in the trench (Ziegler et al. 2002). After detachment of the Kabylian terrane from Iberia and its Langhian collision with North Africa, increased coupling between the Kabylian orogenic wedge and its foreland gave rise to further inversion of the Atlas troughs during Serravallian-Tortonian times and, after slab detachment, during the Pliocene-Quaternary.

1.3 Mediterranean Marine Basins

The modern marine basins of the Mediterranean domain are heterogeneous both in terms of age and geological structure. They are floored by

1. thick continental lithosphere (Adriatic Sea),
2. continental lithosphere thinned to a variable extent (Alboran Sea, Valencia trough, Aegean Sea) up to denuded mantle (central Tyrrhenian Sea),
3. relics of the Permo-Triassic Neotethyan oceanic domains (Ionian and Libyan Seas, E Mediterranean), and
4. oceanic crust of back-arc basins of Late Cretaceous-Paleogene age (Black Sea) or Neogene age (Algero-Provençal basin).

In detail, several of these basins have a more complex structure: for example, only the central, areally subordinate portion of the Black Sea is made of oceanic crust – which, in turn, can be subdivided in two smaller oceanic domains of different ages – whereas all the rest of it is made of stretched continental crust.

1. The **Adriatic Sea** is floored 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, 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); this domain – also referred to as Adria – includes not only what is now the Adriatic Sea but also portions of the Southern Alps, Istria, Gargano and Apulia. The southern Adriatic Sea (crossed by TRANSMED Transect III) is characterized by the updoming of the Adriatic lithosphere both on land (in the Puglia peninsula) and offshore Italy (Carminati et al., this publ.). A broad anticline of 100–150 km wavelength exposes part of the foreland in the Puglia peninsula; such updoming probably constitutes a forebulge related to the subduction of thick Adriatic lithosphere under the southern Apennines since the Pliocene (Doglioni et al. 1994).
2. The **Alboran Sea** is floored by thinned continental crust (down to a minimum of 15 km) and is bounded to the north, west and south by the Betic-Rif orocline. Where sampled by drilling or dredging, the basement of the Alboran Sea consists of metamorphic rocks similar to those of the internal domain of the Rif-Betics. The acoustic basement along several deep seismic profiles is locally formed by volcanic rocks (ba-

saltic andesites to rhyolites) from calc-alkaline series 10 Ma old. (Frizon de Lamotte et al., this publ.). During the Miocene, considerable extension in the Alboran domain and in the adjacent internal domain of the Betic-Rif occurred coevally with thrusting in the more external zones of these mountain belts. Syn-rift sediments are early Burdigalian to late Serravallian in age (Comas et al. 1999). Such late-orogenic extension can be interpreted as the result of westward roll-back of the subducted African lithospheric slab whereby thickened continental crust extends rapidly as the subduction zone retreats (Lonergan and White 1997; Gutscher et al. 2002).

The **Valencia trough** is located between the Iberian mainland and the Balearic Islands. Together with the Liguro-Provençal basin to the NE, it constitutes the oldest western Mediterranean basin, although the Valencia trough displays younger syn-rift deposits, thus indicating a progressive southwestward rift propagation from southern France (Camargue, Gulf of Lion) (Roca 2001). Overall, both basins are characterized by water depths of up to 2 200 m and by an Oligocene-to-Recent sedimentary fill ranging in thickness between 2 and 6 km. The Valencia trough is floored by continental crust which was consolidated during the Variscan orogeny and was extended during the Mesozoic rifting phases preceding the middle Jurassic opening of the oceanic Atlantic-Tethys basin and the mid-Cretaceous opening of the Bay of Biscay-Pyrenean basin (see Stampfli and Borel, this publ.). The Mesozoic rift basins in the area of the Valencia trough were inverted and uplifted during latest Cretaceous-Paleogene time, thus inducing the development of a major unconformity. Finally, the Variscan basement and its Mesozoic sedimentary cover underwent extension starting from the late Chattian (Roca et al., this publ.). Main extensional deformation in the Valencia trough took place during late Oligocene-Aquitanian times although most faults were also active during the entire duration of the early Miocene, as indicated by large lateral thickness variations; a few coastal faults were active throughout the Miocene.

The **Aegean Sea** is located in the upper plate of the Hellenic subduction zone. The arcuate structure of the southern Aegean Sea features the geological and geophysical characteristics typical of an island arc. A Benioff plane defined by seismicity dips from the Hellenic trench towards the NNE as deep as 180 km, and a calcalkaline volcanic arc outlines its curvature. South of the volcanic arc the southern portion of the Aegean Sea is a fore-arc basin (see TRANSMED Transect VII). Crustal-scale extension in this region has been accommodated by shallow-dipping detachment faults, 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 in core complexes and by the intrusion of granitoid and monzonitic magmas at upper crustal levels. The northern Aegean Sea is characterized by a complex fault pattern resulting from east-west-trending strike-slip movements related to the westward propagation of the North Anatolian fault and from north-south-trending extension. 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. From this viewpoint, the rather recent tectonic "extrusion" of Anatolia added only a rigid component to the long-lasting crustal collapse in the Aegean region.

The **Tyrrhenian Sea** is the youngest Mediterranean basin: the oldest sedimentary deposits filling the rift-related grabens along its western and eastern margins are ?Serravallian-Tortonian, thus marking the age of the onset of extension in this region (e.g. Kastens et al. 1990; Mattei et al. 2002). The development of the Tyrrhenian basin has been interpreted as resulting from back-arc extension above the NW-dipping Ionian subduction zone, possibly with a significant component of passive subduction (see Carminati et al., this publ., for further details). The presence of young basaltic bodies in the deepest portions of the Tyrrhenian basin has been somehow overemphasized in the past, leading to a vision of the central Tyrrhenian basin as underlain by true oceanic crust. The results of the reprocessing of preexisting geophysical data integrated with more recent acquisitions indicate that basalts in the central Tyrrhenian are volumetrically limited and that the deepest portions of the basin are mostly made of denuded serpentinized mantle overlain by a veneer of sediments (TRANSMED Transect III). At the scale of the entire Tyrrhenian Sea, the vast majority of the basin is floored by stretched continental crust forming rotated blocks bounded by listric, crustal-scale faults flattening close to the Moho. This structural configuration is particularly clear in the western part of the basin whereas along the Italian peninsula normal faults tend to have a higher dip angle and the stretching factor is lower. North of the 41st parallel the Tyrrhenian Sea shows only a limited degree of crustal stretching.

3. The exact nature of the lithosphere underlying the **Ionian-Libyan Sea** and the **eastern Mediterranean** has been the topic of much debate, being interpreted either as a relic of oceanic crust (Biju-Duval et al. 1977; Vai 1994) or as thinned continental crust (Giese et al. 1982). The timing of the opening of these connected basins has also been a matter of discussion, with age attributions ranging from the late Paleozoic (Vai 1994) to the Cretaceous (Dercourt et al. 1985, 1993, 2000).

For the eastern Mediterranean most authors indicate a Late Triassic or Early Jurassic opening (Garfunkel and Derin 1984; Sengor et al. 1984; Robertson et al. 1996). The presence of a continental crust in the Ionian basin was postulated mainly on the evidence of the overall thickness of the crust (ca. 20 km; deduced from the dispersion of seismic surface waves) and the low heat flow. The issue of the nature of the crust in the deep portions of the Eastern Mediterranean basins was solved by De Voogd et al. (1992) with a two-ship refraction and oblique deep seismic survey showing a relatively thin crust (8–11 km) overlain by a thick pile of sediments (up to 10 km) (see also Finetti in press). Recent palinspastic reconstructions based on all available geological-geophysical evidence (e.g. Stampfli and Borel, this volume) point to the presence of old (Permian?) oceanic crust underneath a thick pile of Mesozoic and Cenozoic sediments which hampers direct sampling and dating. The Ionian-Libyan Sea and the eastern Mediterranean 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 (Fig. 1.1).

4. The more than 2 000 m deep **Black Sea** is partly floored by oceanic crust and probably represents the remnant of a composite Cretaceous-Eocene back-arc basin which developed on the upper plate during north-dipping subduction of the Neotethys (see Stampfli and Borel, this volume). Seismic studies and field evidence in the regions surrounding the Black Sea indicate that its geological make-up is the result of the post-rift coalescence of two different extensional basins (Zonenshain and LePichon 1986; Finetti et al. 1988; Robinson 1997). Rifting of the western Black Sea began in the middle Early Cretaceous (Okay et al. 1994) with the separation of a lithospheric fragment (the Istanbul Zone of Okay and Tüysüz 1999) from the Odessa shelf, i.e. the offshore continuation of the Moesian platform of Romania and Bulgaria. Opening of the western Black Sea came to an end during the early Eocene when the southward drifting of the Istanbul Zone led to collision with Anatolia to form the western Pontides. The age of rifting of the eastern Black Sea is not as well constrained because the relevant stratigraphy is poorly exposed. Nonetheless, several lines of evidence (see Spadini et al. 1996, for a review) support the hypothesis that the eastern Black Sea developed between the Paleocene and the middle Eocene. The crustal structure of the Black Sea is well constrained by a number of geophysical studies (e.g. Belousov et al. 1988). Beneath the central western Black Sea the Moho rises to a depth of about 20 km, including as much as 15 km of post-rift sedimentary fill; the Moho depth increases to 40–45 km both to the north (Russian platform) and to the south (Pontides). Beneath the eastern Black Sea the Moho rises to about 25 km and the thickness of the post-rift succession is about 13 km. Clear-cut magnetic anomalies are absent in both sub-basins, possibly the effect of the huge thickness of post-rift sediments.

Rifting in the **Provençal basin** area occurred at least from the Oligocene (34 Ma) to the middle Aquitanian (21 Ma), according to the age of the sediments drilled in the Gulf of Lion, just beneath the break-up unconformity (Gorini et al. 1993). Drifting and the creation of the central, oceanic portion of the basin took place in the Burdigalian, as indicated by paleomagnetic data (Vigliotti and Langenheim 1995) and by the transition from syn-rift to post-rift subsidence of its margins (Vially and Trémolières 1996; Roca 2001). It is a commonly held notion that the Provençal basin is a partly oceanised back-arc basin produced by the southeastward roll-back of the Apennines-Maghrebides subduction (see Carminati et al., this publ., for further details). Paleomagnetic studies (Alvarez et al. 1974; Vigliotti and Langenheim 1995; Speranza et al. 2002) indicate a counterclockwise rotation of the Corsica-Sardinia block between 19 Ma and 16 Ma (Burdigalian), synchronous with the formation of oceanic crust in the Provençal basin. In a pre-rotation fit, Corsica is to be considered adjacent to Provence and the original position of Sardinia was in the centre of the present-day Provençal basin, far from the continental slope of the Gulf of Lion.

The **Algerian basin** is located between the Balearic block and the North African coast and morphologically represents the continuation of the Provençal basin toward the southwest. In the absence of deep drilling in the Algerian basin, little is known on the age and characteristics of its sedimentary fill whose stratigraphy has been inferred by correlation with better known areas nearby (Mauffret et al. 1973; Sans and Sàbat 1993). However, the prominent difference in the thickness of Miocene sediments between the Provençal basin (4 km) and the Algerian basin (1.8 km) suggests that the latter is younger (Roca et al., this publ.). Despite the absence of direct evidence, the thin crust (4–6 km) of the Algerian basin is probably oceanic (Hinz 1972) and can be compared to the oceanic crust (about 5 km thick) of the Provençal basin to the northeast (Pascal et al. 1993). The age of the Algerian basin must be pre-Messinian, as the oldest dated deposits filling the basin are Messinian. The unloaded depth of the basement gives an apparent age of 20 Ma whereas the 125 mW m^{-2} heat flow of the East Alboran basin corresponds to an apparent age of 16 Ma. Tomographic studies (Carminati et al. 1998a,b) and the magmatic history (Maury et al. 2000) suggest an age comprised between 15 Ma and 10 Ma.

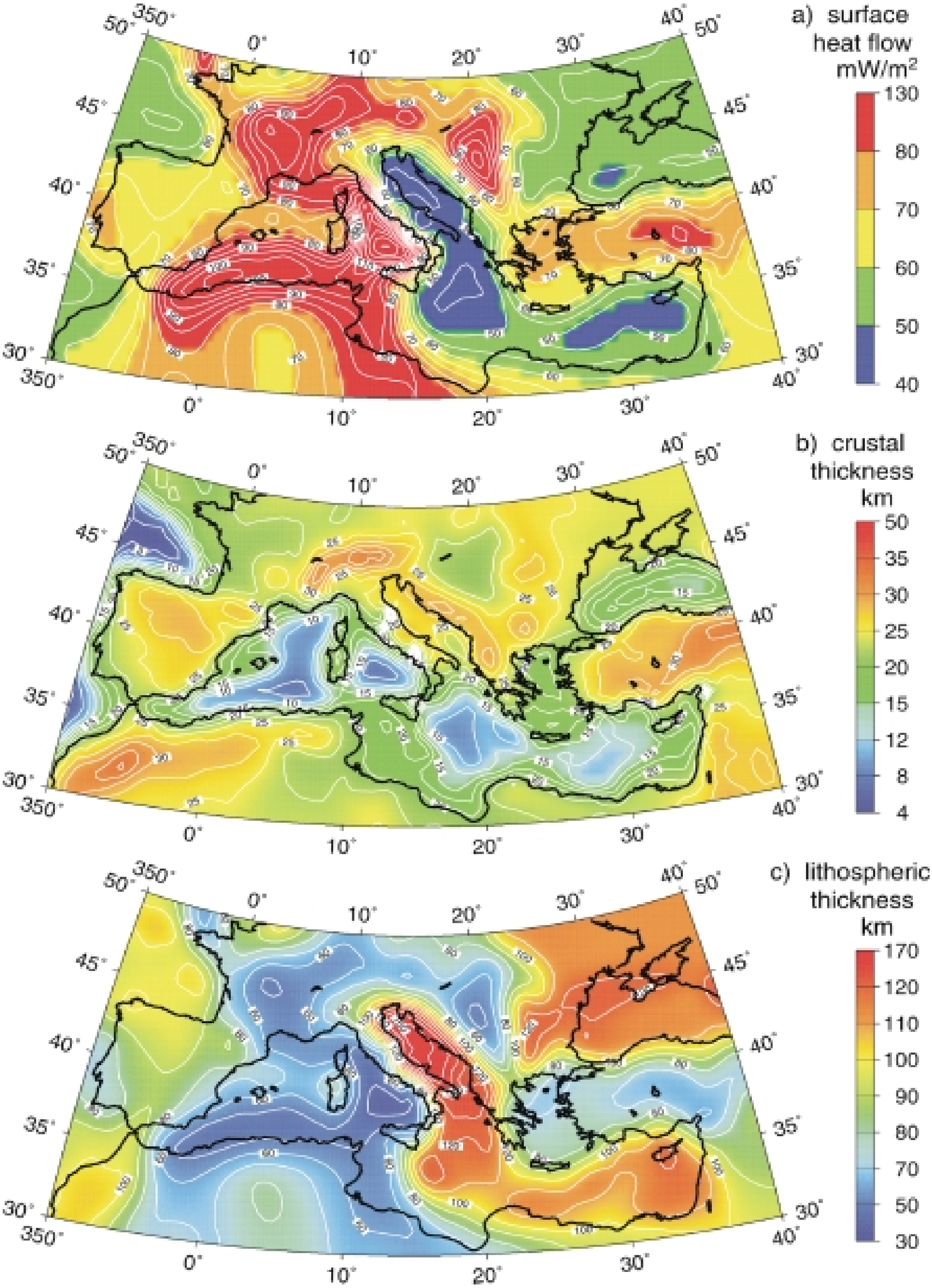

Fig. 1.2. a Surface heat-flow data (mW m^{-2}; contour interval 5 mW m^{-2}); **b** calculated crustal thickness (contour interval 2.5 km); **c** calculated lithospheric thickness, contour interval 10 km. After Jiménez-Munt et al. (2003)

1.4 Geological-geophysical Baseline

1.4.1 Heat Flow

Figure 1.2a is a schematic surface heat-flow map of the Mediterranean area and the surrounding regions. There is a good correlation between lithospheric thickness (Fig. 1.2c) and heat-flow values. The entire southern portion of western and central Europe is subject to normal or high heat flow, with higher values in south-central France (Massif Central), northern Switzerland and the Pannonian basin. The western Mediterranean basins show the highest heat-flow values: 60–100 mW m^{-2} in the Gulf of Lion and Provençal basin, around 100 mW m^{-2} in the southern part of the Algerian basin, and values in excess of 120 mW m^{-2} in the deepest portion of the Tyrrhenian Sea, thus suggesting that heat flow is inversely proportional to the age of rifting/drifting of the basins (Carminati et al., this publ.). Other regions of relatively high heat flow are Anatolia and the Aegean Sea.

Low heat flows (< 50 mW m^{-2}) are present in the Adriatic and Ionian Seas and in the Eastern Mediterranean, all areas characterized by old lithosphere (either continental – Adriatic – or oceanic – Ionian and eastern Mediterranean) and by thick sedimentary successions. Another area of low heat flow is the Black Sea, where up to 15 km of sediments have blanketed the floor of this composite Cretaceous-Eocene back-arc basin (see Stephenson et al., this publ.).

1.4.2 Crustal and Lithospheric Structure

Figure 1.2b and c depict the overall structure of the crust and lithosphere, respectively, over the Mediterranean region. The depth of the Moho varies approximately between 5 and 50 km. Minimum crustal thickness of about 5–15 km is found in oceanic domains such as – from west to east – the Algero-Provençal basin, the Tyrrhenian Sea, the Ionian Sea, and the E Mediterranean basins (Jiménez-Munt et al. 2003). The crust is thick beneath the orogenic belts, like the Atlas, the Alps, and the Dinarides; a broad region of crustal thickening is present in eastern Anatolia. Significant crustal thinning is present in the Pannonian basin.

More detail on the crustal structure of the western Mediterranean (and NW European) region is shown in Fig. 1.3, a compilation of the most recent sources (Dèzes and Ziegler 2002). At this scale it is possible to appreciate details otherwise lost. These include (*i*) the significant crustal root of the Pyrenees, (*ii*) the extremely shallow Moho in areas of the Tyrrhenian Sea, where the mantle has been denuded (see Transect III, this publ.), and (*iii*) crustal thickening along the frontal part of the northern Apennines produced by thrust stacking and syntectonic sedimentation.

Lithospheric thickness (Fig. 1.2c) reaches minimum values in the Algero-Provençal and Tyrrhenian basins, both sites of Neogene extension. Conversely, the much older (Permian-Triassic?) Ionian and Levant oceanic basins have a thicker lithosphere.

1.4.3 Gravity

Figure 1.4 shows the gravity field over most of the Mediterranean region, with a few exceptions in eastern Europe, Asia Minor and the Middle East (Wybraniec et al. 2004). The figure shows Bouguer anomalies on land and free-air anomalies offshore. The general patterns of the Mediterranean gravity field can be explained to a large extent by the variation in Moho depth as well as by crustal-level igneous intrusions and the presence of large sedimentary basins. For instance, all major mountain ranges, including the Alps, Apennines, Carpathians, Pyrenees, and Betic Cordillera, are associated with strong negative Bouguer anomalies (shown in blue-violet color), a natural consequence of crustal thickening beneath the mountains. Sedimentary accumulations, such as the Po basin south of the Alps and north of the northern Apennines, are clearly associated with large negative anomalies. The prominent lows in most of the Iberian peninsula may indicate low-density mantle, as significant crustal thickening beneath the elevated plateau has not been documented. The active subduction zone in the region from the Hellenic arc to Cyprus is also associated with large negative anomalies.

A belt of strong positive Bouguer gravity anomalies (shown in red) along the Tyrrhenian coast of Italy may be the result of crustal thinning in the inner portion of the Apenninic orogenic wedge (see Elter et al. 2003 for a review of the Apennines). The marked high centered in the northern Aegean Sea can be attributed to the extensional tectonic regime characterizing the area during the Neogene (see TRANSMED Transect VII).

Figure 1.5 shows the Bouguer gravity anomalies for the Mediterranean Sea and most of the adjacent regions on land. Overall, the gravity fabric of the Mediterranean Sea is characterized in almost all deep physiographic basins by strong to very strong positive Bouguer anomalies. In the deep basins west of the Corsica-Sardinia block $\Delta g''$ values range between 100 and 180 mGal. The Algero-Provençal basin and most of the Ligurian Sea have $\Delta g''$ values in excess of 140 mGal, typical of the oceanic crust. Along the Iberian coast the gravity values follow the bathymetry, also outlining with a certain detail the Balearic continental block. In the Alboran Sea Bouguer anomaly values range from about –20 mGal at the Gibraltar arc to +40 mGal where it merges with the Algerian basin; these values agree with the stretched continental crust flooring the basin (Frizon de Lamotte et al., this publ.). The western Mediterranean basins are

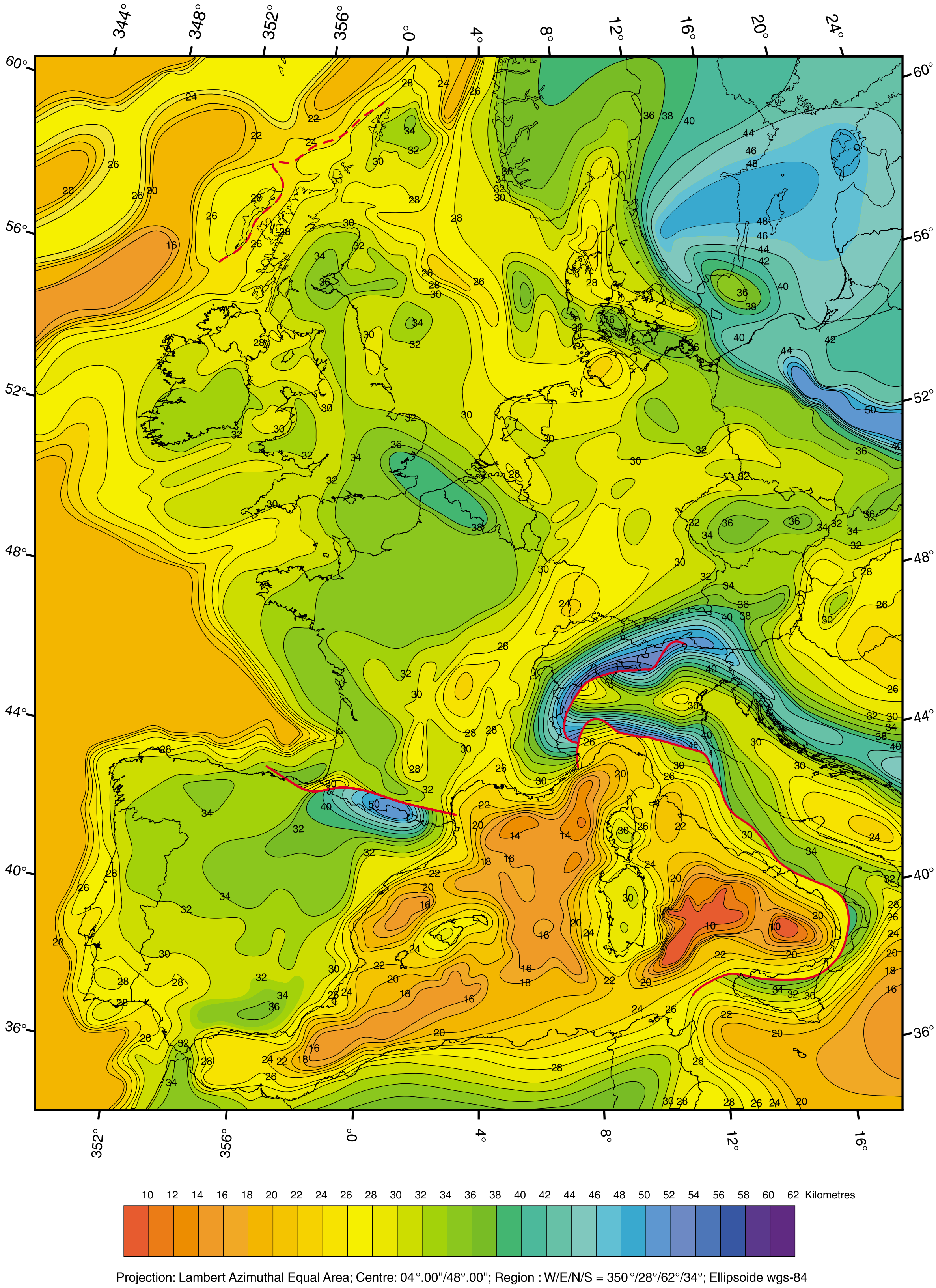

Fig. 1.3. Map of the depth of the Mohorovicic discontinuity in western Europe. From Dèzes and Ziegler (2002)

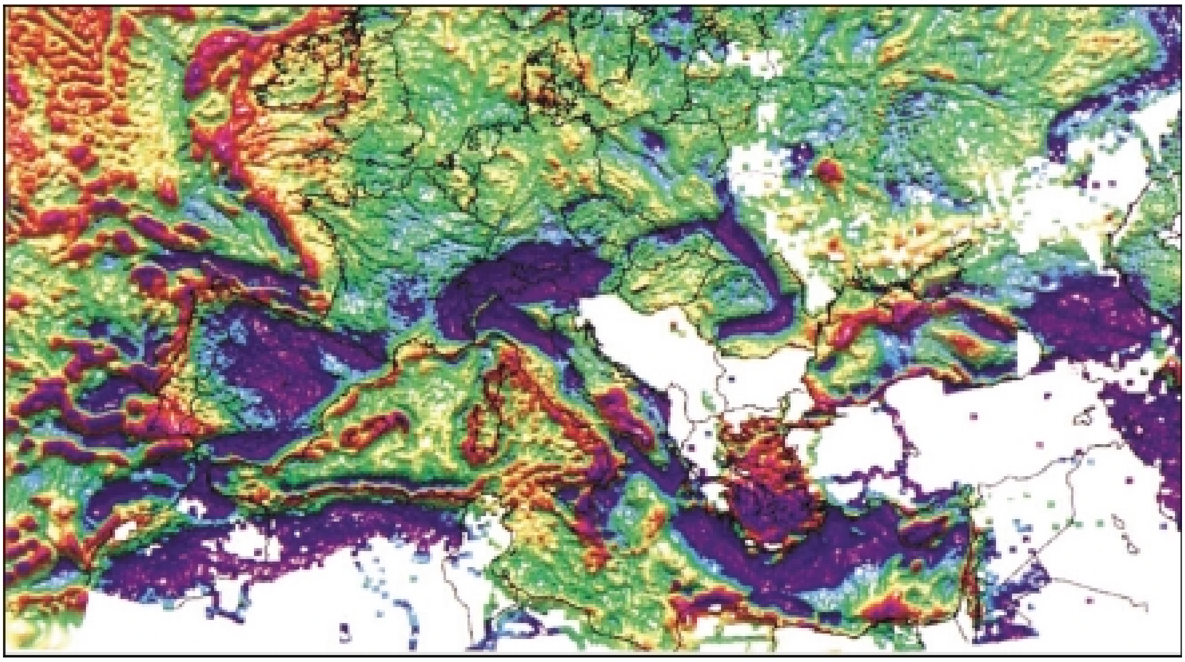

Fig. 1.4. Gravity map showing Bouguer anomalies on land and free-air anomalies offshore (after Wybraniec et al. 2004)

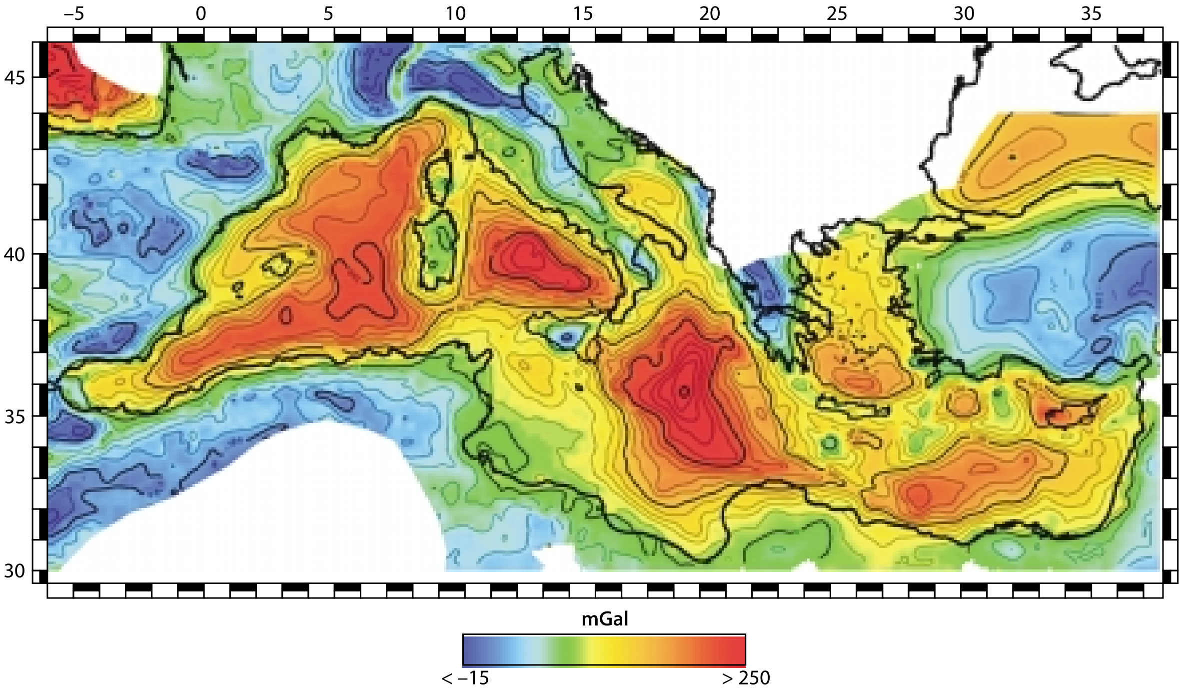

Fig. 1.5. Bouguer gravity anomalies (after Makris et al. 1998, with minor modifications)

bounded to the south by the northern margin of the African plate and by the strong negative anomalies along the Atlas Mountains. To the north, the thickened continental crust of the Betic Cordillera, the Pyrenees and the Alps mark areas of strong negative Bouguer anomalies (down to –180 mGal in the central Alps).

Bouguer gravity values in the Tyrrhenian Sea mimic closely the bathymetry and range between 180 mGal in the central part to about 20 mGal on the coasts of Italy and along the eastern margin of the Corsica-Sardinian block. As discussed in the text accompanying TRANSMED Transect III, these gravity values are associated with stretched continental crust and denuded mantle, with volumetrically limited outpourings of basalts.

The highest values (ca. 300 mGal) of the Bouguer anomaly in the Mediterranean coincides with the deepest portion of the Ionian Sea, where the crust is thin and overlain only by a 4–5 km-thick sedimentary column (Makris et al. 1998).

The Aegean Sea is an area underlain by thinned continental crust and by relatively high values of $\Delta g''$, particularly in the Cretan Sea. The Aegean Sea is bounded by two areas of low Bouguer anomalies: the Hellenides to the west and Anatolia to the east. The Cretan arc has $\Delta g''$ values ranging between –10 and +30 mGal and separates the stretched Aegean Sea from the Ionian and Libyan Seas.

The eastern Mediterranean is dominated by a prominent ENE-WSW-trending positive anomaly culminating gravimetrically at about 220 mGal in the Herodotus abyssal plain. The anomalies decrease toward the east, mainly as the result of the increase in sediment thickness, reaching nearly 16 km in the Nile delta and offshore Israel. This broad anomaly is bounded to the north by a series of gravity highs and lows aligned east-west parallel the Anatolian margin; these highs and lows follow the bathymetric features. As to Cyprus, it shows a strong positive anomaly.

The Black Sea offshore Turkey is characterized by $\Delta g''$ values between 20 and 100 mGal, in agreement with the presence of stretched continental crust and small portions of oceanic crust, both overlain by a thick sedimentary sequence (see Stephenson et al., this publ.).

1.4.4 Magnetic Field

Figure 1.6 shows the magnetic anomalies in the Mediterranean basin at 50 nT intervals. This region is characterized both by very localized anomalies as well as by anomalies distributed throughout the area; seafloor magnetic lineations, as evident in the major oceans, are absent. Most local anomalies are the strong ones related to the fault systems along the continental slopes, particularly in the western Mediterranean, which is characterized by young, deep basins bounded by high-angle normal faults indicating very rapid subsidence. In the deep basinal areas of the western Mediterranean (i.e. west of Corsica and Sardinia) the depth of the top of the layer responsible for the magnetic anomalies is at 8–9 km (Le Borgne et al. 1971), which is the depth of the top of oceanic layer 2 as indicated by deep seismic soundings (Zanolla et al. 1998). This confirms that the crustal structure in the deep basinal area is typical of an oceanic crust. The nature of the crust in rifted areas such as the Alboran Sea (thinned continental crust, widespread yet areally discrete volcanism) and the Valencia trough (large volcanic body) makes the magnetic signatures of these areas much different from those of the deep basinal areas. In the eastern portion of the Algerian basin the magnetic anomalies follow a clear-cut N-S trend which is the southward continuation of the magnetic anomalies following the western Corsica-Sardinia margin. As elsewhere in the Mediterranean, they can be interpreted as the result of magmatic intrusions

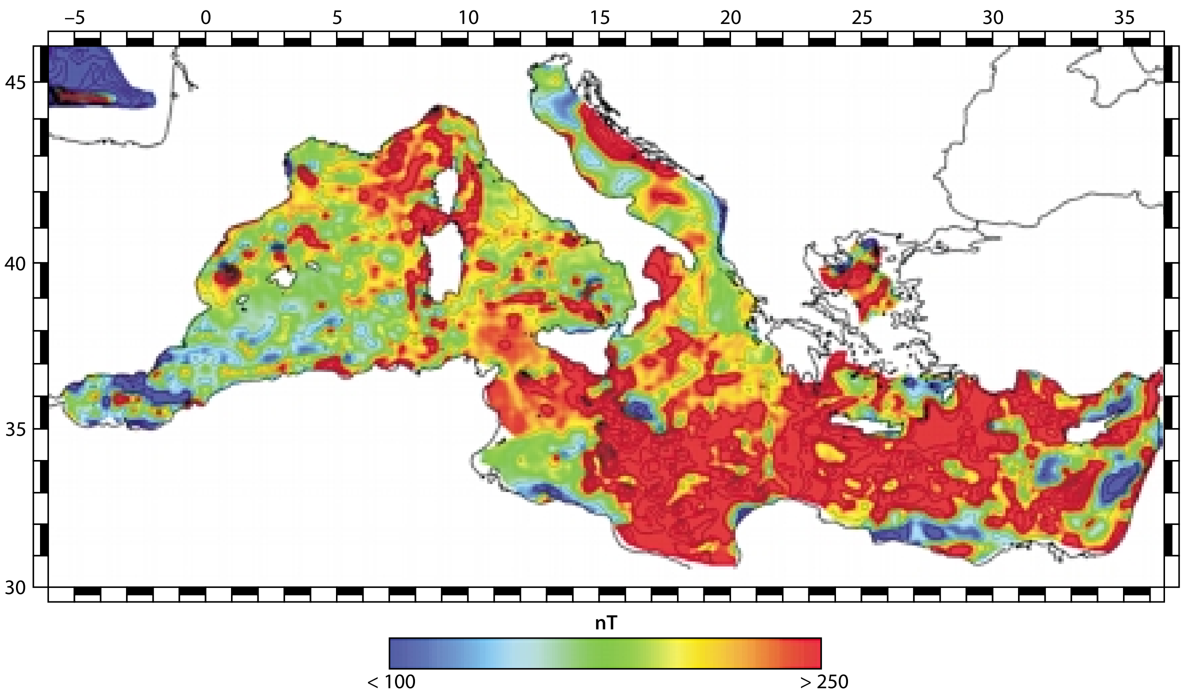

Fig. 1.6. Magnetic anomalies (after Zanolla et al. 1998, with minor modifications)

along extensional faults located at the transition between the bathyal plain and the continental margins. The magnetic anomalies along the western Sardinian margin thus confirm the results already obtained from seismic data: the continental crust of the Corsica-Sardinia block extends far offshore and oceanic crust is present only in the central part of the Algerian basin (see TRANSMED Transect II).

The eastern Ligurian Sea can be envisioned as the stretched southward continuation of the western Alps. Medium-high frequency magnetic anomalies, with approximately 100 nT maximum intensity and associated with 2–3 km deep, highly magnetized bodies, show an overall NNW-SSE trend connecting northeastern Corsica with the Alpine orogenic belt of western Liguria.

A complex array of local and regional magnetic anomalies characterizes the Tyrrhenian Sea, the result of the very young age and complexity of this peculiar basin born from the SE-ward drift of the Calabria-Peloritani terrane in latest Miocene-Pliocene times (see Bonardi et al. 2001 for a review). Rifting started along the Sardinian margin during the middle-late Miocene and the magnetic anomalies thus generated are evident along a N-S-trending strip east of Sardinia. The deep portion of the Tyrrhenian basin displays a magnetic fabric characterized by anomalies of medium wavelength (30 km) and high amplitude (> 30 nT) associated with a basic magnetic substratum (magnetic susceptibility $k = 0.020–0.045$). South of the 41st parallel magnetic anomalies trend approximately E-W. The calcalkaline and tholeiitic volcanic edifices of the Tyrrhenian bathyal plain are associated with very high-amplitude anomalies which mask the trend of the basic substratum as well as the weaker anomalies generated by small seamounts made of continental basement. Two clear-cut E-W-oriented alignments of local magnetic anomalies are present in the Tyrrhenian Sea at 41° and 39° latitude N: the northern one marks the boundary of the deepest part of the basin.

The foreland areas located on the Adriatic plate and genetically associated with the Apenninic and Dinaric fold-and-thrust belts are characterized by regional magnetic anomalies of long wavelength (60–100 km) and amplitudes ranging between 40 and 200 nT which are associated with structural highs in the basement. The very strong regional anomaly (> 320 nT) off the Dalmatian coast is probably caused by structural uplift of the basement or changes in its magnetic susceptibility (Zanolla et al. 1998).

Gravity anomalies (Fig. 1.5), seismic results (see TRANSMED Transects VII and VIII), the reduced magnetic anomalies in the deep basins (Fig. 1.6), and heat-flow data (Fig. 1.2) all suggest that the deep abyssal plains of the Eastern Mediterranean are underlain by a thick sedimentary cover lying on a thin crust of oceanic or intermediate nature which thickens to the south towards the African margin and becomes progressively more disrupted and tilted from east to west.

The entire easternmost Mediterranean, from the Libyan Sea to the Levant Sea, is characterized by incipient continental collision. This tectonically active area shows local magnetic anomalies: *positive* ones are related to igneous bodies (as for the continental margin of Israel and offshore Egypt between about 27° and 29° longitude); *negative* ones correspond to small trenches or sedimentary basins. The dipolar anomalies of the Cyprus arc can be attributed to the Cretaceous ophiolites.

As to regional magnetic anomalies, a large positive anomaly is centered south of Cyprus and corresponds to the Eratosthenes seamount (TRANSMED Transect VIII). Two large negative anomalies – interpreted by Zanolla et al. (1998) as the result of an oceanic crust of inverse polarity flooring the deepest parts of the E Mediterranean – are located off the coast of Egypt (between 25° and 27° longitude) and between the Eratosthenes seamount and Israel. The high amplitude anomalies of the Ionian Sea are associated with a semi-oceanic basement located at depths of 5–7 km b.s.l. and are probably linked to the rifting episode which formed the basin (Finetti 1982). In the rest of the eastern Mediterranean the magnetic field is mostly regular: this can be attributed to the thick pile of sediments covering the area and which becomes thicker in the Nile Cone (16 km) and the Eratosthenes trench (15 km).

1.4.5 Seismicity

Seismicity in the Mediterranean area is high both in terms of frequency and magnitude. Figure 1.7 depicts the geographic distribution of the epicenters of 1 112 seismic events with surface magnitudes $2.8 < M_s < 8$ for the period 1903–1999. Although rather diffuse, seismicity in the region is particularly concentrated in the Aegean and peri-Aegean area. Other areas of high seismicity include the Maghrebian orogen of northern Africa and Sicily, the Apennines, the Carnic Alps and the southern Dinarides. All these regions have been repeatedly affected in historical times by catastrophic earthquakes causing a high death toll (Giardini 1999).

Mediterranean seismicity distribution is related to the major known tectonic systems and clearly follows the boundary between the Eurasian and African plates. A few discrete seismogenic zones can be recognized, such as the subduction belt of the Hellenic arc and the geologically complex zone that includes the southern Dinarides, the northern Aegean Sea along the trace of the North Anatolian Fault system, and western Anatolia. Seismicity in this region is dominated by extensional focal mechanisms. Another significant seismogenic zone starts at the Azores, continues eastward and, after crossing Gibraltar, follows the Maghrebian orogen of northen

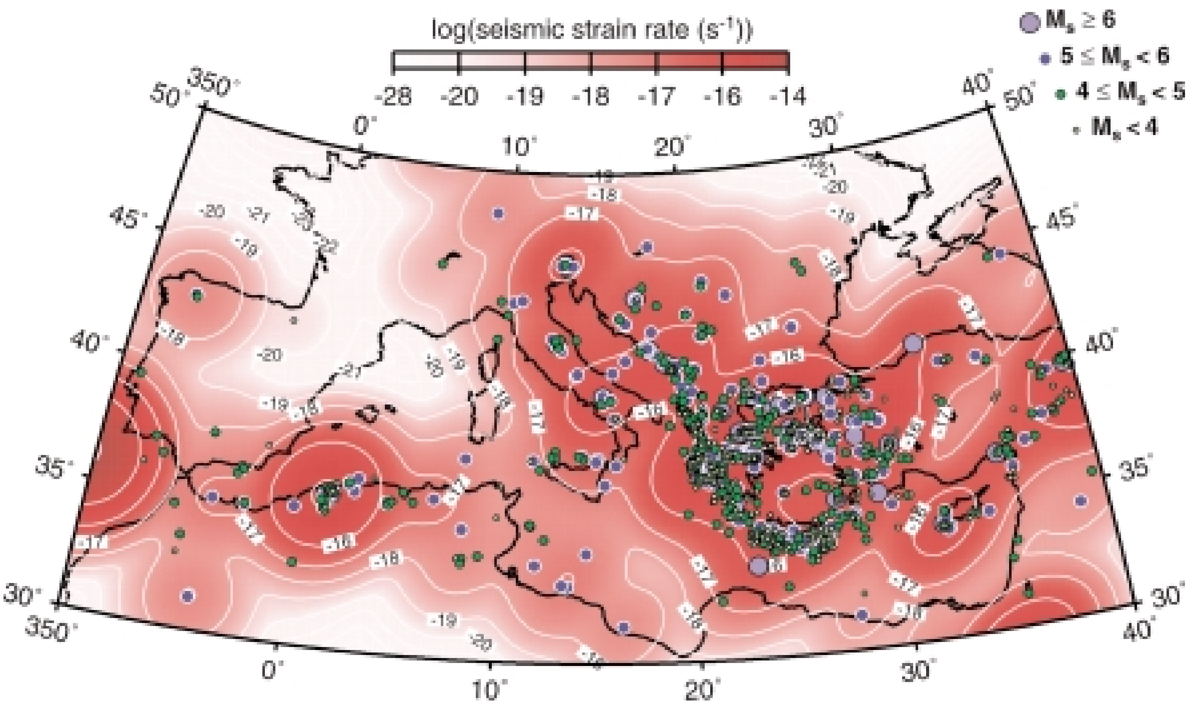

Fig. 1.7. Seismicity with surface magnitude (M_S) between 2.8 and 8 in the Mediterranean region, and calculated seismic strain. From Jiménez-Munt et al. (2003)

Africa (Rif and Tell Chains) and Sicily (Sicilian Maghrebides) as well as the Calabria-Peloritani terrane. This belt is considered representative for the collisional margin between the African and Eurasian plates, since along it there is a uniform stress field with a direction of relative movement in agreement with that expected from global plate-motion models. Conversely, in the remaining Mediterranean area, the situation is more complex.

In Europe there are other seismogenic areas where low-magnitude events ($M_S < 3$) are common and higher magnitude earthquakes occur sometimes. Among the more active zones there is the Rhine Graben, where, besides a frequent yet weak seismicity, there are also events of $M_S > 5$. The same pattern is observed in the Jura Mountains, in the Swiss-French Alps, in the Italian-French Alps and also in the lower Rhone and Durance valleys. Although less important, some seismic activity occurs also in the Central Massif of southern France.

Earthquake focal depths are generally restricted to crustal levels, except in areas where oceanic or continental lithosphere is subducted, either actively or passively. From west to east, this is the case in the Gibraltar region, beneath the Calabria-Peloritani terrane, along the Hellenic arc, in Vrancea along the SE Carpathians, and in the Antalya Gulf of southern Turkey.

Figure 1.7 also shows the associated seismic strain rates calculated by Jiménez-Munt et al. (2003). The results of such calculations indicate that the largest seismic strain rate release is of the order of 10^{-16}–10^{-15} s^{-1} occurring – from west to east – along the plate boundary in north Africa, in southern and northeastern Italy, in the eastern Alps, the southern Dinarides, and in the Aegean and peri-Aegean region, comprising the entire western Anatolia. Other areas of high seismic strain rates include easternmost Anatolia along the trace of the Assyrian-Zagros suture and Cyprus.

1.4.6 Geodetic Data

Figure 1.8 depicts a geodetic set of 190 vector velocities for the Mediterranean region with respect to a fixed Eurasia. Vector velocities were obtained and/or compiled by Jiménez-Munt et al. (2003) combining GPS, satellite laser ranging (SLR) and very long baseline interferometry (VLBI) data. The solution shown represents the residual velocity with respect to the Eurasian block obtained by subtracting the rigid motion of Eurasia expressed in the NUVEL-1A reference frame (see Devoti et al. 2002, for a discussion of the procedure).

In spite of areas such as the Iberian peninsula and the northern Adriatic block where large error ellipses result from the paucity of data, three coherent domains are present: (*i*) a generally SSE-ward direction in the Iberian peninsula and the Ligurian-Provençal coast; (*ii*) a generally northward direction for Italy, with a progressive clockwise rotation from Lampedusa Island (LAMP), through Calabria (COSE), to Matera (MATE); and (*iii*) a counterclockwise rotation from NW-ward to SSW-ward directions from eastern Anatolia to the south-

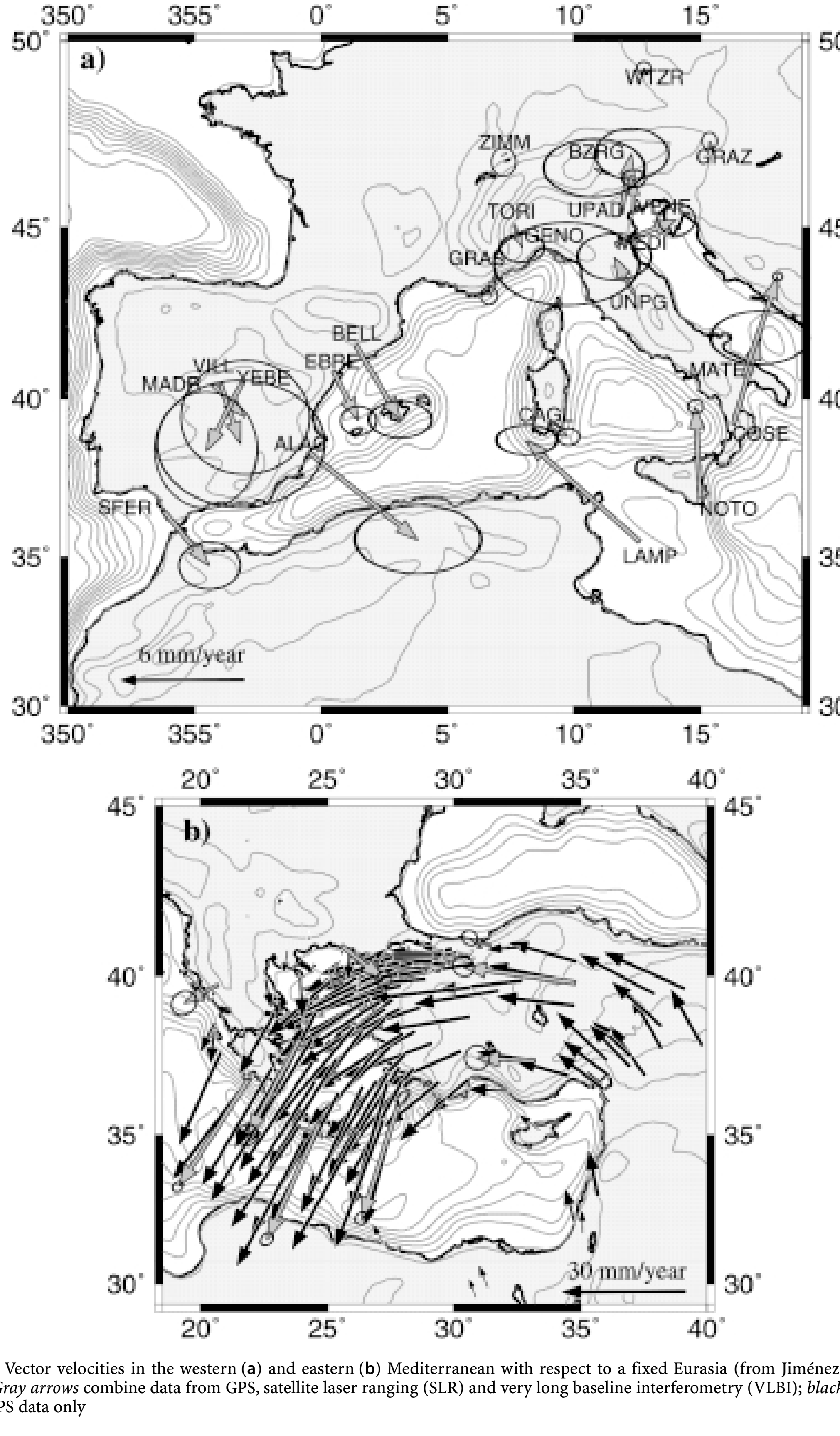

Fig. 1.8. Vector velocities in the western (**a**) and eastern (**b**) Mediterranean with respect to a fixed Eurasia (from Jiménez-Munt et al. 2003). *Gray arrows* combine data from GPS, satellite laser ranging (SLR) and very long baseline interferometry (VLBI); *black arrows* are from GPS data only

ern Aegean region. The geodetic pattern is also characterized by a substantial increase in vector magnitude both from north to south and from west to east.

In the eastern Mediterranean region (Fig. 1.8b) the velocity field shows the northward motion of the Arabian plate and the counterclockwide rotation of central and western Anatolia and the southern Aegean, which are bounded to the north by the North Anatolian Fault system and its extension in the northern Aegean Sea. It is noteworthy that in this region velocities progressively increase from eastern Anatolia to the southern Aegean Sea, where they reach values in excess of 30 mm yr^{-1}. This velocity gradient seems to contradicts the commonly held notion that the overall westward motion of Anatolia is driven by escape tectonics from the Bitlis-Zagros collision zone.

1.4.7 Stress Field

Figure 1.9 shows the tectonic stress orientation within the Mediterranean region, as compiled in the World Stress Map (Reinecker et al. 2003). Stress indicators used to this end include (*i*) earthquake focal mechanisms (69%), (*ii*) borehole breakouts and drilling-induced fractures (19%), in-situ stress measurements (8%), and young geologic data from fault-slip analysis and volcanic vent alignments (4%). [Readers are referred to Zoback and Zoback (1980, 1991) and Zoback et al. (1989) for a detailed description on the different methodologies to derive stress information from those types of indicators.] All data are quality ranked according to a scheme developed by Zoback and Zoback (1989) and based mainly on the number, accuracy and depth of the measurements. The following is a brief description of the overall stress field in the Mediterranean abstracted from Jiménez-Munt et al. (2003).

In the western Mediterranean, maximum horizontal compressional stress axis is roughly oriented NNW, i.e. parallel to the convergence vector between the African and the European plates. Exceptions are the arcuate structures of the western Alps and Gibraltar, where relatively small deviations are present. Conversely, in the central and eastern Mediterranean the stress field is rather variable yet regionally coherent, being associated with collisional orogens and with the Calabrian and Hellenic subduction zones.

In spite of the paucity of data in some regions, stress indicators along the northern African margin (and Sicily) indicate NW compression, thus reflecting again the overall direction of convergence between the Eurasian and the African plates. The Calabria-Peloritani terrane of southern Italy displays a complex stress regime, characterized by a combination of normal and strike-slip faulting which Rebai et al. (1992) ascribe to radial extension (Bonardi et al. 2001). The northern Apennines are dominated by extension in the internal (southwestern) part of the orogenic wedge and by compression-transpression in its external (Adriatic) portion. In the southern Apennines normal and strike-slip faulting prevail, with extension perpendicular to the axis of the orogenic belt (e.g. Frepoli and Amato 2000). The whole Aegean and peri-Aegean region is characterized by an extensional regime: radial extension parallel to the hinge line of subduction is localized within the southern part of the Aegean Sea, whereas the northern Aegean Sea and the surrounding landmasses are affected by N-S oriented extension. In Anatolia stress directions rotate progressively counterclockwise from NE-trending compression in eastern Anatolia to NE extension in western Anatolia. This stress pattern mirrors the westward movement of Anatolia, that is driven by the combination of syn-collisional escape tectonics (e.g. Kahle et al. 2000) and rollback of the subducting Ionian Sea slab, entailing southwestward advance of the Hellenic arc-trench system (e.g. Jolivet 2001).

1.5 Global Dynamics and Active Processes Exemplified in the Mediterranean

Since the early beginning of urban civilization, Mediterranean people had to face the devastating effects of earthquakes and volcanic eruptions associated with active margins, either due to subduction of remnants of the oceanic Neotethys or to continental collision between the European and African plates, and the intervening Adria and Iberia microplates (Tapponnier 1977; Channell et al. 1979; Letouzey and Trémolières 1980; Platt et al. 1989; Mazzoli and Helman 1994; Ziegler and Roure 1996; Muttoni et al. 2001; Nikishin et al. 2002; Ziegler et al. 2002; Cavazza and Wezel 2003). Superimposed on this overall convergent, compressional regime, local back-arc extension accounts also for recent rifting and seafloor spreading in parts of the Western Mediterranean and Tyrrhenian basins (Le Pichon et al. 1971; Durand et al. 1999; Ziegler et al. 2001a). However, current Mediterranean tectonic activity is not only controlled by compressional and extensional forces, as density contrasts, vertical compaction and gravity account also for active mud and salt diapirism in areas of rapid sedimentation such as the Nile delta, the Alboran Sea and the Mediterranean Ridge.

Lastly, successive episodes of uplift and subsidence have controlled in the past, and still control, the connections between the world ocean and the Mediterranean and adjacent basins (i.e. the former Paratethys and the modern Black Sea), accounting for either slow or very fast sea-level rises or drops during the Messinian and the Pleistocene. As such, the geological record of the Mediterranean is likely to provide key analogues for the study of global warming, greenhouse effects, related sea-level rise and its potential impact on densely populated coastal areas.

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