Univerza v Ljubljani
Naravoslovnotehniška fakulteta
Oddelek za geologijo

Vodič za predmet

SKLEPNE TERENSKE VAJE

2008 - Severna Italija in Toskana

26.05.–04.06.2008

Avtorja in organizatorja vaj:

doc. dr. Andrej Šmuc
asist. dr. Timotej Verbovšek

Ljubljana, 21.05.2008

Trasa ekskurzije v Severno Italijo in v Toskano, 26.05.-04.06.2008

1. Dan: ponedeljek 26.05.2008

1. Vajont: plaz
2. Belluno bazen:mezozojska stratigrafija

Spanje: Trento

2 Dan: torek 27.05.2008 (vodstvo prof. Luca Martire)

1. Monte Baldo: Madona di Corona (Trento plato)
2. Cava Viannini: aktivni kamnolom ammonitica rossa

Spanje: Biassa (Cinque Terre)

3. Dan: sreda 28.05.2008

1. Cinque Terre – oligocenski fliši ter ligurijski ofioliti

Spanje: Marina di Massa

4. Dan: četrtek 29.05.2008

1. Carrara: marmor
2. Metamorfni kompleks Apuanskih alp

Spanje: Marina di Massa

5. Dan: petek 30.05.2008 (vodstvo dr. Alessia Arias)

1. Larderello (geotermija) in fumarole pri Monterotondu
2. Posledice rudarjenja pri Boccheggianu

Spanje: Castle of Selvole (pri Sieni)

6. Dan: sobota 31.05.2008 (vodstvo Anton Marn)

1. prost dan za obisk Siene in Chiantija

Spanje: Castle of Selvole (pri Sieni)

7. Dan: nedelja 01.06.2008, (vodstvo prof. Andrea Brogi)

1. Le Crete, Rapolano terme: izločanje travertina ob aktivnem prelomu
2. Bagno Vignoni ali Bagni San Filippo: termalne vode
3. Radicofani: vulkanizem

Spanje: Bolsena

8. Dan: ponedeljek 02.06.2008

1. Bolsena: bazalti s stebrastim krojenjem
2. Latera: lavin tok
3. Pitigliano: ignimbriti
4. Terme di Saturnia (Le Cascatelle): travertin

Spanje: Piombino

9. Dan: torek 03.06.2008 (vodstvo lokalnega geologa)

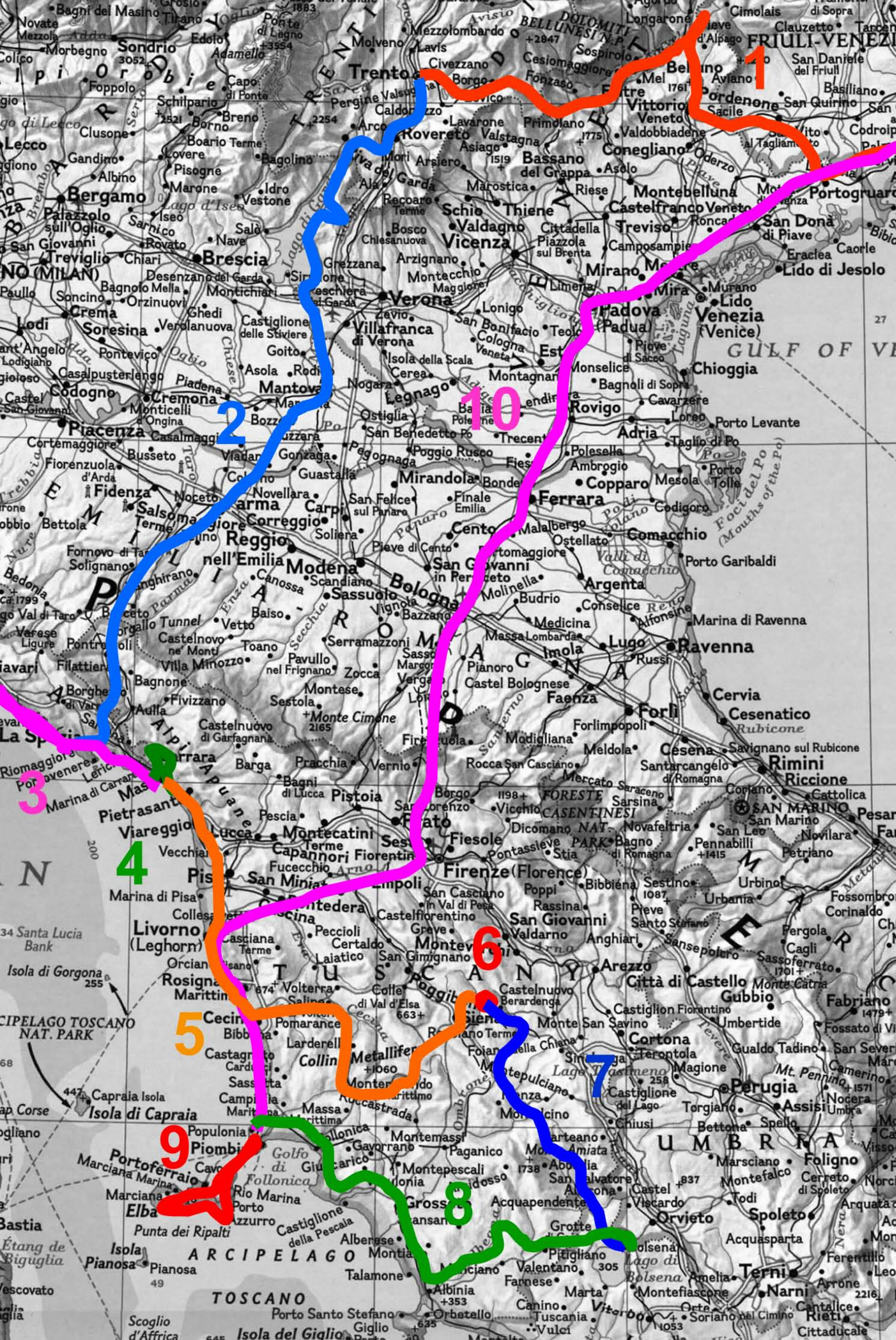
1. Elba: orudenja, skarni, graniti

Spanje: Piombino

10. Dan: sreda 04.06.2008 (vodstvo Nevio Pugliese)

1. Gradež - Oglej: lagune

Vrnitev v Ljubljano



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Trento
Belluno
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Portogruaro
Rovereto
Bergamo
Brescia
Verona
Vicenza
Treviso
Padova (Padua)
Venezia (Venice)
GULF OF VE
Chioggia
Mantova
Cremona
Piacenza
Parma
Reggio nell'Emilia
Modena
Bologna
Ferrara
Rovigo
Ravenna
Rimini
Forlì
Cesena
SAN MARINO
Firenze (Florence)
Prato
Pistoia
Lucca
Pisa
Livorno (Leghorn)
Viareggio
La Spezia
Massa
Carrara
Siena
Arezzo
Perugia
Assisi
Orvieto
Spoleto
Terni
Viterbo
Grosseto
Piombino
Portoferraio
Elba
ARCIPELAGO TOSCANO
Isola di Capraia
Isola del Giglio
Lago di Bolsena
TUSCANY
UMBRIA
EMILIA-ROMAGNA
FRIULI-VENEZI
DOLOMITI BELLUNESI N.P.
Golfo di Follonica
Lago Trasimeno

Seznam udeležencev

Št.	Priimek
01	BAVEC ŠPELA
02	BIZJAK NEJC
03	ČINČ PETRA
04	DOBNIKAR META
05	FUKS TADEJ
06	GAJSER ANDREJA
07	JERIČ MARJA
08	KLASINC MATJAŽ
09	KOMAR DARJA
10	KOVAČIČ ŽIGA
11	LAJMIŠ LEA
12	MARUŠIČ DAVOR
13	MAVC MIRO
14	MILAVEC JURIJ
15	MILETIĆ SNJEŽANA
16	MLINAR TJAŠA
17	PETERNEL TINA
18	PRKIČ NINA
19	ROJC KRISTJAN
20	ROZMAN DAVID
21	ROŽIČ BOŠTJAN
22	SENEKOVIČ MOJCA
23	SLAPNIK ANDREJA
24	ŠMUC ANDREJ
25	ŠRAM DEJAN
26	ŠVARC KATJA
27	UDOVČ MIRAN
28	VERBOVŠEK TIMOTEJ

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by William Cavazza[1] and Forese Carlo Wezel[2]

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

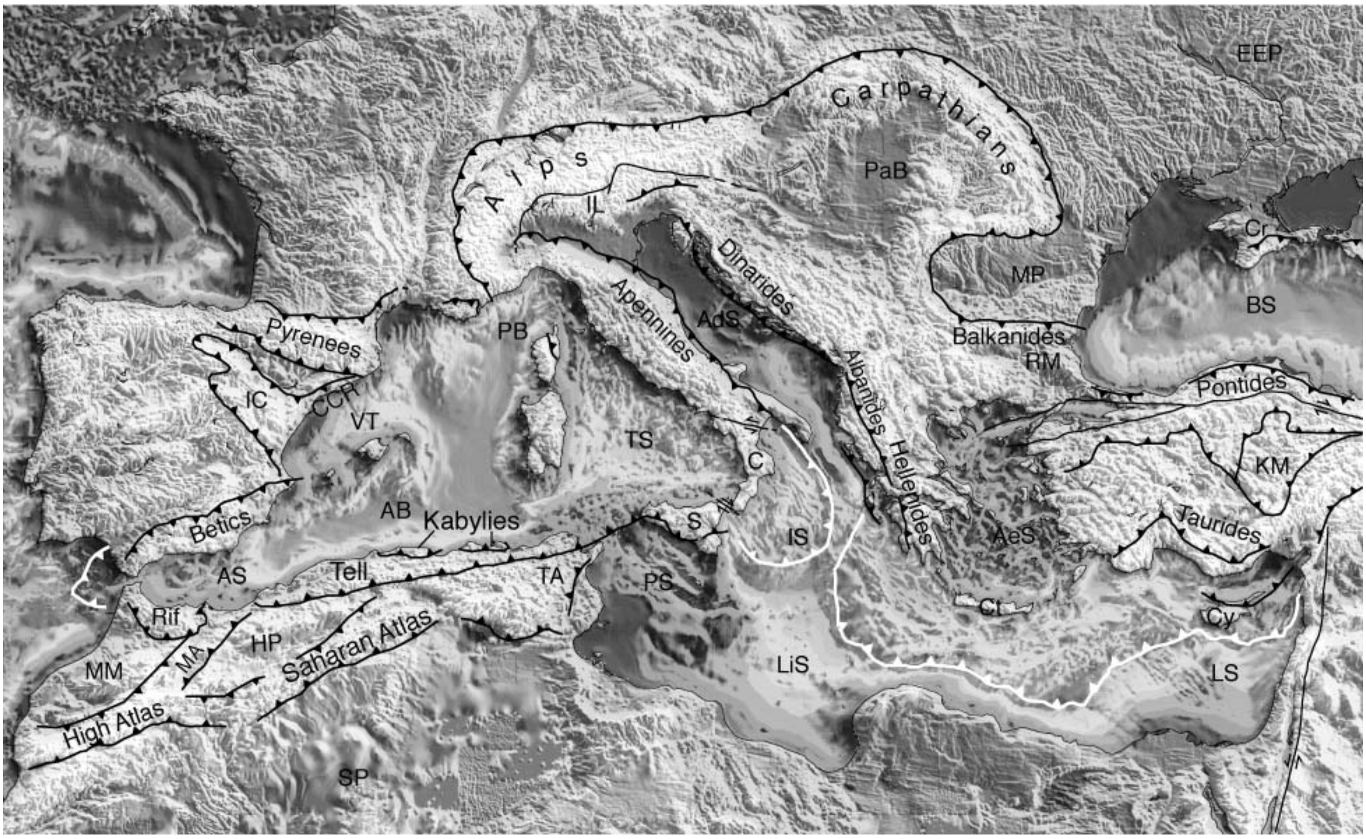

Figure 1 Digital terrain model of the Mediterranean region with major, simplified geological structures. White thrust symbols indicate the outer deformation front along the Ionian and eastern Mediterraean subduction fronts. AB, Algerian basin; AS, Alboran Sea; AdS, Adriatic Sea; AeS, Aegean Sea; BS, Black Sea; C, Calabria-Peloritani terrane; CCR, Catalan Coast Range; Cr, Crimea; Ct, Crete; Cy, Cyprus; EEP, East European Platform; HP, High Plateaux; KM, 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.

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 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-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., 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 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, 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 (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

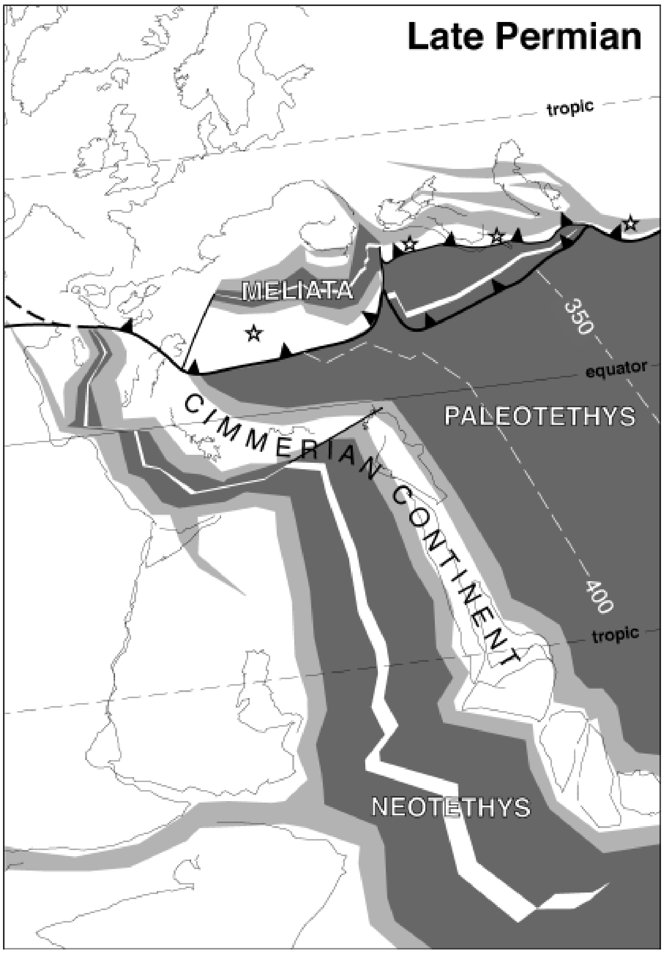

Figure 2 Paleogeographic reconstruction of the western Tethyan area during the late Permian (from Stampfli et al., 2001b, with minor modifications). Stars indicate magmatic activity.

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

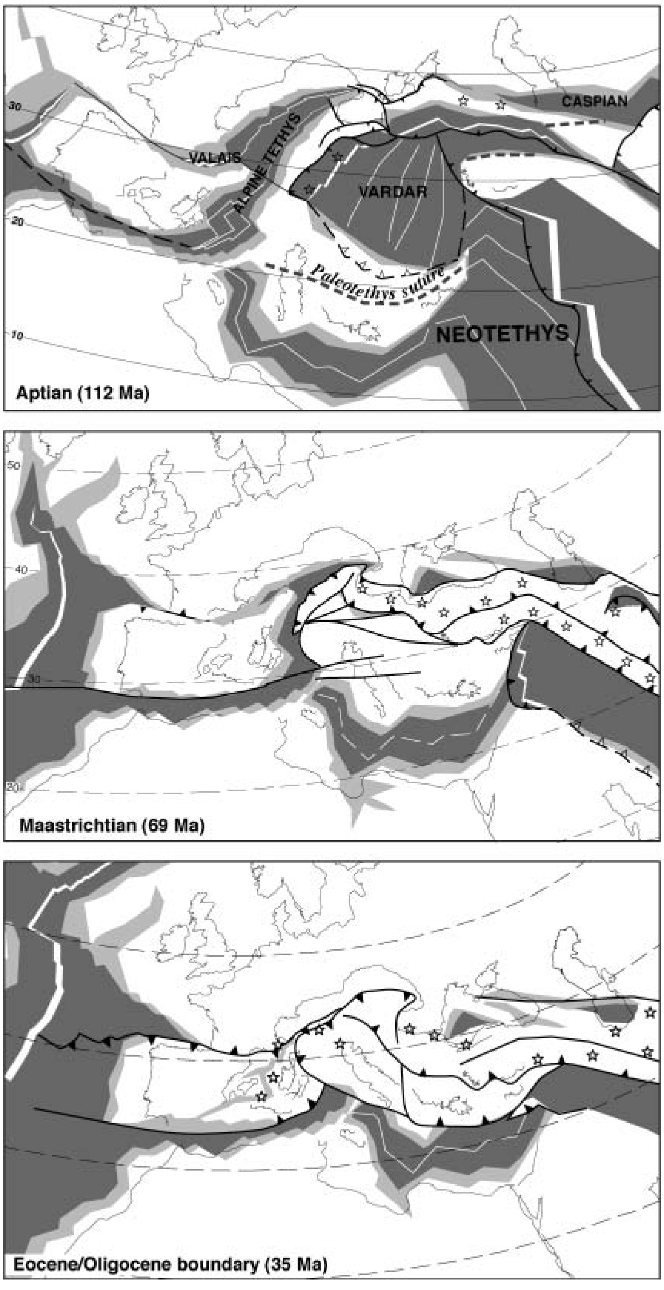

Figure 3 Paleogeographic reconstructions of the western Tethyan area during the Aptian, Maastrichtian and at the Eocene/ Oligocene boundary. Note the progressive narrowing and suturing of the oceanic domains comprised between the Eurasian and Iberia continental blocks to the north and the Africa/Arabia 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 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 periaegean 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-Apennines-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).

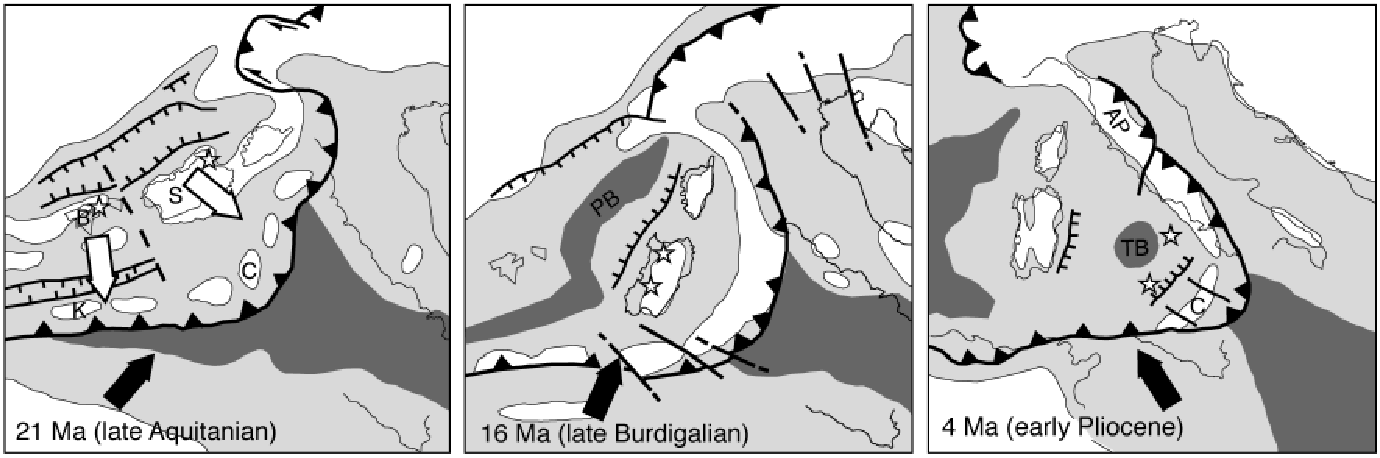

Figure 4 Schematic maps showing the paleotectonic evolution of the W Mediterranean during Neogene time (modified after Bonardi et al., 2001, and Roca, 2001). Only active tectonic elements are shown. White, exposed land; light gray, epicontinental sea; darker gray, oceanic crust. Black arrows indicate 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, Kabilies; PB, Provençal Basin, S, Sardinia; TB, Tyrrhenian Basin.

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 (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 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).

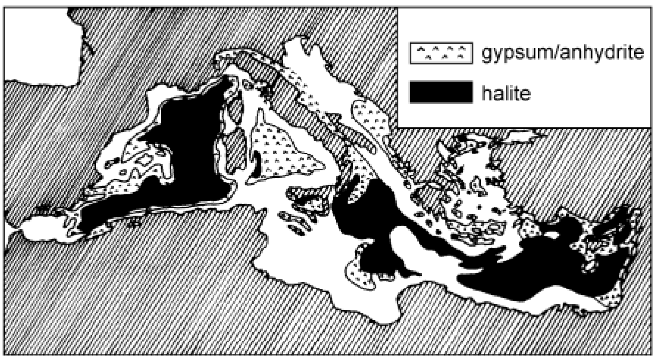

Figure 5 Areal extent of the Messinian evaporites in the Mediterranean region. Modified after Rouchy (1980).

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 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 sedimentaion 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".

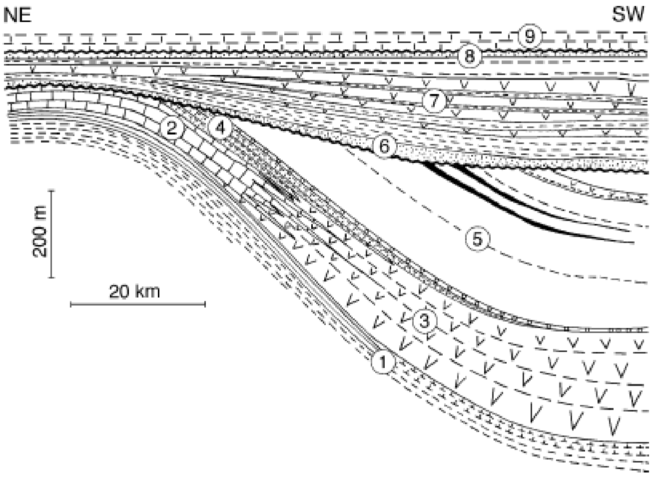

Figure 6 Schematic stratigraphic cross section of the Messinian of Sicily (modified after Decima and Wezel, 1973). 1) Pre-evaporitic clay, marl and diatomite (Tripoli Fm.); 2) evaporitic limestone (Calcare di Base); 3) lower gypsum beds (Gessi di Cattolica); 4) gypsum turbidites; 5) halite and potash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

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 compression 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 subaerial 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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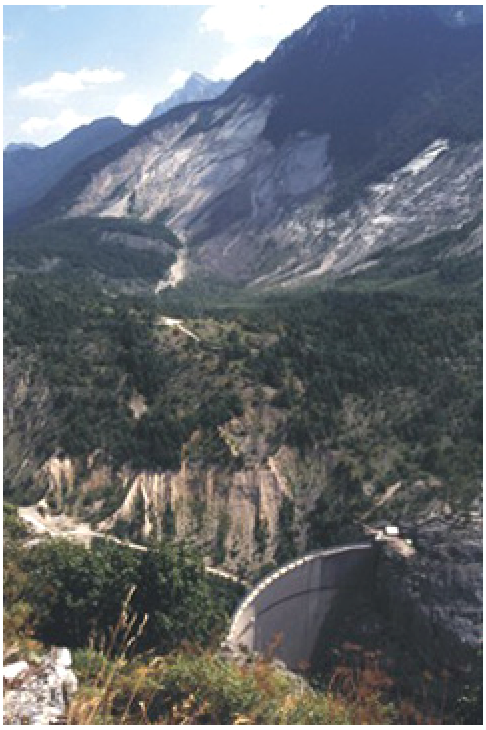
Udarni val v dolini Vajont (Vaiont)

Besedilo: B. Komac in M. Zorn, foto: M. Zorn
(http://www.zrc-sazu.si/giam/vajont.htm)



V dolini Vajont (Vaiont) v Italiji se je dne 9. 10. 1963 ob 22:38 zgodila katastrofa; ogromna kamnita gmota je zdrsela v akumulacijsko jezero in povzročila nastanek udarnega vala, ki je pod seboj pokopal skoraj dva tisoč ljudi.

Nesreča je nastala izključno zaradi človeškega dejavnika, v prvi vrsti želje po dobičku, pa tudi nezmožnosti odgovornih strokovnjakov, da bi pravilno razlagali dogajanje v naravi.

Vajontski jez s kamnitim zdrsom pod goro Toc v ozadju (foto: M. Zorn, 11. 8. 2003).

Zdrs je nastal na gori Toc (Monte Toc), ki ji domačini pravijo tudi *la montagna che camina*, t. j. 'gora, ki hodi'. V umetno jezero je s hitrostjo do 30 m/s zgrmelo kar 270 mio. m^3 gradiva. Nastal je val, ki bi ga lahko imenovali antropogeni tsunami in se je vzpel v pobočje. Dosegel je vas Casso, ki leži 250 m višje, vendar je ni huje prizadel. Voda, ki je nato zgrmela nazaj, je razgalila površino kamnite zdrsine in jo erodirala. Manjši vodni val je segel po dolini Vajonta navzgor, glavni udarni val pa po dolini navzdol. Preskočil je jez, ki praktično ni utrpel posledic, in z veliko močjo pometel z vasmi pod njim. Uničil oziroma poškodoval je naselja Longarone, Rivalta, Pirago, Faé, Villanova in prizadel občino Castellavazza. Terjal je več kot 1900 življenj. Longarone in Vajont sva podpisana obiskala 11. avgusta 2003 (Historical 2000; Paoloni in Vacis 2000; Petley 2000; The Vaiont 2000).

"*Zgodba o Vajontu (...) je zgodba nadzornikov, ki niso nadzorovali, modrih mož, ki niso vedeli, inženirjev brez talenta (...) in številnih mož, ki so zavrnili uporabo zdrave pameti za varovanje človeških življenj*" (I. Calvino).

Več o dogodku pred štiridesetimi leti ...

Jez v ozki dolini Vajont so zgradili septembra leta 1960 kot del velikega hidroenergetskega sistema v porečju Piave za potrebe hitro rastočih severnoitaljanskih mest, zlasti Milana, Torina in Modene. Z višino 261,6 m je bil takrat najvišji jez na svetu, zanj so porabili kar 353.000 m^3 betona. Prvi načrti za jez segajo že v leto 1900, prva dela so stekla šele leta 1956 (Petley 2000).

Dolina Vajont ima sinklinalno geološko zgradbo, tako da kamninske plasti na severnih in južnih pobočjih vpadajo proti dolinskemu dnu, kar je zelo ugodno za drsenje. Pobočja nad dolino sestavljajo apnenci jurske in triasne starosti, med temi plastmi pa so tudi tanjše glinene plasti. Dolina je bila preoblikovana še v ledenih dobah, ko so jo poglobili ledeniki. V začetku holocena so se pobočja ponovno stabilizirala s pobočnimi procesi (Nelson 2000; Petley 2000).

Že med gradnjo jezu so na desnem pobočju našli sledi starega zdrsa. Kljub temu so domnevali, da so zdrsi večjih razsežnosti zelo malo verjetni, čeprav manjših niso izključili.

Februarja 1960, so še pred dokončanjem jezu, začeli s polnjenjem jezera. Marca 1961, ko je na pobočjih nad jezerom prišlo do prvih nestabilnosti, je umetno jezero že bilo 130 m globoko. Oktobra istega leta so dosegli globino 170 m. Takrat so južna pobočja polzela s hitrostjo približno 3,5 cm/dan, nastala je tudi 2 km dolga razpoka. 4. 11. 1960, ko so gladino jezera dvignili na 180 m, je nastal kamniti zdrs. V približno desetih minutah v jezero zdrselo približno 700.000 m^3 gradiva. Pristojni so ugotovili, da lahko obvladajo drsenje z nadzorovanim spreminjanjem gladine jezera.

Po tem dogodku so jezero spustili na globino 135 m. S tem se je polzenje zmanjšalo na 1 mm/dan. Zato so oktobra 1961 spet začeli dvigati gladino in februarja 1962 dosegli relativno višino 185 m, novembra 1962 pa 235 m. Ob dviganju gladine se polzenje pobočij sprva ni bistveno spremenilo, sčasoma pa se je hitrost povečala na 1,2 cm/dan. Novembra 1962 so spet začeli spuščati gladino in do aprila so dosegli raven 185 m, premikanje pobočij pa se je skoraj ustavilo.

Sledil je še tretji dvig ravni jezera. Maja 1963 so dosegli raven 231 m, pri čemer so pobočja polzela s hitrostjo 0,3 cm/dan. Junija so raven dvignili na 237 m, hitrost polzenja pobočij se je dvignila na 0,4 cm/dan. Ko so julija dosegli raven 240 m je bila hitrost 0,5 cm/dan, avgusta pa se je povečala na 0,8 cm/dan. Septembra so dosegli raven 245 m, polzenje pobočij pa se je povečalo na 3,5 cm/dan. Sledilo je počasno spuščanje gladine jezera. Devetega oktobra 1963 je bila globina jezera 235 m, hitrost polzenja pa je tega dne znašala do 20 cm/dan.

Ob 22:38 se je sprožil 1,8 km dolg in 1,6 km širok zdrs, ki je trajal 45 sekund. Jezero je imelo približno 115 mio. m^3 vode. Ko je ogromna gmota zdrsela vanj, je nastal velik val (50 mio. m^3), ki je potoval proti zahodu in vzhodu. Na nasprotni strani doline je segel do 260 m nad raven gladine jezera in ob tem poplavil vas Casso. Nazadnje se je proti dolini Piave izlilo 25 - 30 mio. m^3 vode. Nastal je 70 m visok udarni val, ki je 500 m niže uničil vasi Longarone, Pirago, Villanova, Rivalta in Fae ter segal še 2 do 2,5 km po dolini Piave navzgor. Reka Piava je bila še 60 km po toku navzdol visoka 12 m. Betonski jez je ob tem dogodku ostal skoraj popolnoma nepoškodovan. Tudi vzhodni val je poplavil nekaj vasi in terjal več 100 smrtnih žrtev. Cel dogodek je trajal le sedem minut (Abele 1971; Historical 2000; Paolini in Vacis 2000; Petley 2000; The Vaiont 2000).

Zdrs je nastal v 5 - 15 cm debelih glinenih plasteh v apnencu. Po mnenju nekaterih naj bi šlo za reaktivacijo starejšega zdrsa. Povod za zdrs leta 1963 pa so bile obilne desetdnevne avgustovske in septembrske padavine, ki so dodatno obremenile pobočja nad jezerom. Zelo pomembna vzroka sta tudi antropogeno spreminjanje ravni talne vode v pobočjih prek sprememb gladine jezera in vpad z glino prepojenih plasti proti jezeru (Nelson 2000; Petley 2000; Semenza 2001).

Preglednica: število stavb pred nesrečo in število uničenih stavb (vir: http://www.vajont.net).

	Število stavb pred nesrečo	Število uničenih stavb
Longarone	372	361
Pirago in Rivalta	159	159
Villanova in Faé	59	32
Ostala naselja	635	0
Skupaj	1225	552

Kronologija (1961-1963)

3. 2. 1961 Leopold Müller pravi v geološkem poročilu: "*... Ni dvoma, da lahko v veliki globini obstaja drsna ploskev. Prostornino kamnitega zdrsa ocenjujem na 200.000.000 m^3. Premikov tako velikega kamnitega zdrsa zagotovo ne bi mogli ustaviti. Popolnoma zaustavi se lahko le sam...*" Poudaril je tudi, da je zelo verjeten hipen zdrs. Podjetje za proizvodnjo električne energije SADE (Societa Adriatica Di Elettricita) ni sprejelo tega mnenja, temveč je verjelo manj alarmantnim zagotovilom iz poročila, ki so ga prej napisali Müller, Semenza and Giudici in pravi, da naj bi kamniti zdrs iz dveh med seboj ločenih delov meril od 20 do 40 mio m^3.

Poleti 1961 je SADE odredila izdelavo računalniške simulacije morebitnega drsenja z gore Toc v merilu 1 : 200, ki so jo naredili v Padovi. Istočasno so začeli z obsežnimi deli, ki naj bi preprečila drsenje oziroma omilila njegove posledice. Raven jezera so spustili na 600 m in skozi goro Salta do hidroelektrarne izkopali *by-pass* dovodno cev.

10. 5. 1961 so gladino ponovno dvignili na 660 m in sproti spremljali drsenje.

Avgusta in septembra 1961 so geološko zgradbo podrobneje raziskali z vrtinami, v katere so tudi vstavili 4 piezometre. Pokazalo se je, da je drsna ploskev zelo globoko. V bližini betonskega jezu so postavili seizmografski laboratorij za spremljanje potresov ter opazovali premike s pomočjo reflektorjev, ki so bili postavljeni na geodetsko natančno izmerjene kontrolne točke. Odgovorni v hidroelektrarni so v tem času začeli ponarejati poročila, ki so jih pošiljali vladi. Poneverjali so poročila o seizmični dejavnosti, do katere je prihajalo ob zviševanju gladine jezera.

Oktobra 1961 je umrl "oče" vajontskega jezu, inženir Carlo Semenza. Njegov sin, geolog Edoardo Semenza, je skupaj z Leopoldom Müllerjem napovedal, kaj se bo v Vajontu zgodilo, vendar ga odgovorni niso poslušali. Potem, ko je 20. aprila 1962 umrl še geolog Giorgio Dal Piaz, ni SADE uradno naročila nobenega uradnega geološkega poročila več, ki bi moglo spremeniti zastavljeno pot.

8. junija 1962 je vlada dovolila dvig gladine na 700 m. Slab mesec dni kasneje pravi Augusto Ghetti v poročilu: "*Raven jezerske gladine na 700 m je popolnoma varna glede drsenja*." Kamniti zdrs bi lahko po njegovem mnenju obsegal največ 40 mio. m^3. Kljub oceni o majhni ogroženosti območja je predlagal nadaljevanje raziskav o vplivu morebitnega preliva vode čez vajontski jez. Tudi njegovega predloga niso sprejeli. Jezero je doseglo gladino 700 m 17. novembra 1962, ko so tudi drugi začeli dvomiti o upravičenosti posegov. Vladni svetovalec Bertolissi je premike opisal kot "*bližajoče se kritični točki*". Zato so 2. decembra začeli zniževati raven gladine.

14. marca 1963 je italijanska vlada z dekretom prenesla pristojnosti SADE na novo ustanovljeno ENEL ("Ente Nazionale per l'energia Elettrica", Nacionalna družba za električno energijo). Ob predaji poslov so vajonsko elektrarno v uradnih dokumentih opisali kot delujočo, kar pa bi lahko formalno potrdili šele po preizkusu delovanja na višini 715 m. SADE bi morala s testiranji končati preden bi bil končan birokratski postopek. V nasprotnem primeru bi morali elektrarno prodati drugemu lastniku z oznako *v fazi testiranja* oziroma *nedelujoče*, kar pa bi ji močno zmanjšalo ceno.

10. aprila 1963 so jezero napolnili do višine 647,5 m. Gradivo je bilo še vedno nestabilno, vrstili so se šibki potresi. Hišam so odpadali ometi, pojavljale so se razpoke, zaradi premikov niso mogli zapirati oken in vrat, izsušili so se vodnjaki. Lokalne oblasti in prebivalci so zaman protestirali. SADE se je želela prebivalcem odkupiti z gradnjo šole za družine, ki so živeli pod goro Toc, in so morali sicer otroke voziti v Casso. Ta čas so piezometri pokazali, da je gladina talne vode v gradivu enaka kot v jezeru, kar je pomenilo, da je kamniti zdrs že povsem samostojna geološka enota, tako rekoč ločena od podlage.

20. 3. 1963 je ENEL-SADE odredila dvig gladine na 715 m, kar je močno povečalo nasprotja med prebivalci naselij Erto in Casso ter omenjeno družbo. Potem ko je ENEL-SADE dala denar za novo šolo, je župan prebivalcem dovolil dostop do obale jezera, kar je bilo prej prepovedano. Družba ENEL-SADE ga je zaradi tega javno grajala, saj naj bi na območju veljala *stalna znana nevarnost*. Cesta, ki jo je okrog jezera zgradila SADE, je v tem času utrpela očitne spremembe in je postala valovita. Šola v Pinedi pa že mesec dni po odprtju ni bila več primerna za pouk. Povečala se je seizmična aktivnost območja, tla so se tresla in grmela.

22. julija je župan Erta in Cassa poslal telegram na prefekturo (pokrajinsko policijsko upravo) v Videm in upravi ENEL-a v Benetke. Jezernica je postajala kalna in nemirna.

1. septembra so dosegli gladino 709,40 m, kar je bilo že 10 m nad varno ravnijo in je destabiliziralo kammniti zdrs. Zato so se odločili, da bodo za nekaj časa prenehali polniti jezero in njegovo gladino ustalili na 710 m, kar se je zgodilo 4. septembra. V tem času naj bi se drsenje umirilo v novem ravnotežnem stanju.

2. septembra je ob 10:30 nastal močan potres (magnitude 4 po Richterjevi lestvici), ki ga je čutilo vseh 2000 prebivalcev petih naselij nad jezom. Na gori Salta se je ena od hiš premaknila v središče naselja. Zaradi potresa se je celoten kamniti zdrs pomaknil navzdol za 22 mm, čez dan pa še za nadaljnjih 6,5 mm. Ljudi je postalo strah. Župan Erta je poslal pismo ENEL-SADE: "… *Drugi lahko zmanjšujejo pomen teh stvari, vendar zadevajo življenje, varnost in lastnino prebivalcev Erta*." Nanj ni dobil odgovora. Odgovorni so nekako želeli zaustaviti drsenje, vendar niso vedeli, kako. Hitro praznjenje jezera bi povzročilo hiter premik gmote. Po nasvetu strokovnjakov naj gladina ne bi nikoli presegla 700 m, takrat pa je bila deset metrov višja. Rob jezu je bil le 11,6 m nad njo, nekatere hiše v Ertu pa le 20 m. Po ustalitvi vodne gladine na 710 m so ugotovili tendenco umirjanja drsenja, ki bi jo lahko še izboljšali s počasnim zniževanjem gladine. Prvotno zamisel o višini gladine jezera na 715 m so opustili, vendar si še vedno niso upali spustiti gladine na varnih 700 m.

12. septembra je Biadene odgovoril na pismo župana Erta, rekoč: "*vse je popolnoma pod nadzorom*." 15. septembra se je kamniti zdrs pomaknil za nadaljnjih 12 mm, 26. septembra pa za 22 mm. Odločitev za znižanje gladine jezera je bila sprejeta brez podrobnejšega razmisleka o tem, ali to narediti s hitrim ali počasnim praznjenjem in brez vladnega dovoljenja. Gladino jezera so začeli 27. septembra zniževati za 0,7 m/dan. Dogovor o predaji pristojnosti s SADE na ENEL je še vedno ostal tajen in tudi o zadnjih aktivnostih niso obvestili javnosti. Stanje je bilo uradno še vedno pod nadzorom, zato niso izdali nobenih opozoril.

Sodelavca ENEL-SADE Pancinija, ki je 1. oktobra zapustil Vajont in odšel na počitnice v ZDA, je nadomestil ing. Beniamino Caruso, nadzornik SADE za hidrotehnične naprave na območju "Srednje Piave". 2. oktobra se je kamniti zdrs spet pomaknil, tokrat za 40 mm, prav tako dan kasneje.
V nedeljo, 6. oktobra, je nekaj delavcev na gori Toc ranil podor. Krožna cesta okrog jezera ni bila več prevozna, na gori so nastajale razpoke, drevesa so padala. Zvečer so prebivalcem na območju gore Toc ukazali evakuacijo. Samo tri vasi (Pineda, Liron and Prada) so še ostale dostopne.

Osmega oktobra ob 10:30 zjutraj je Biadene, kot predstavnik ENEL-SADE, poslal telegram županu Erta in Cassa: "*Zaradi povečanega drsenja na levi obali jezera pod goro Toc je nujna evakuacija prebivalcev. Opozarjamo na nujnost in prepovedujemo prisotnost prebivalcev na tem območju. Prepovedan je tudi dostop do celotnega jezera pod koto 730 m in do ceste na levi obali jezera, ki vodi do vasi Costa, Gervasio in Pineda ter do jezera*." V obeh naseljih so objavili naslednje opozorilo: "*Prebivalce obveščamo, da so uslužbenci ENEL-SADE zaznali nestabilnost na pobočju gore Toc, zato se priporoča, da zapustite območje med Costa Gervasio in Pinedo. Prebivalcem Cassa svetujemo, naj uporabijo pomoč ENEL-SADE in zapustijo območje (…) in ponovno opozarjamo vse prebivalce, da je ekstremno nevaren vsak dostop do obale jezera, saj lahko ob drsenju nastanejo nekajmetrski valovi, ki so nevarni tudi za dobre plavalce…*" Ob približno 15:00 je vladni svetovalec Bertolissi obiskal jez, ugotovil nevarnost in Ministrstvo vprašal za navodila. Poročilo so naslednje jutro izročili krajevnemu vodji in je bilo popoldan z navadno pošto odposlano v Rim.

7. oktobra so na gori Toc na mestu starega kamnitega zdrsa (iz leta 1960) opazovali skale, ki so se kotalile v jezero. Slišali so tudi zamolkle udarce, »*kot bi se v podzemlju nekaj trgalo*«. Tudi na gozdnatem pobočju so opazili številne, nekaj deset metrov dolge in meter široke razpoke, ki so potekale vzporedno z jezersko obalo. Številni premiki gradiva na južnem pobočju so bili vidni celo iz Cassa. Na večer so iz »*previdnosti*« evakuirali stanovalce Pinede, Prade in Lirona.

V sredo, 9. oktobra 1963, je bil sončen dan. Biadene, odgovorni na vajontski hidroelektrarni, se je odločil, da bi bilo dobro Pancinija poklicati z dopusta in mu je napisal pismo: *"…v zadnjih nekaj dnevih je hitrost drsenja očitno narasla (…). Nastale so razpoke na pobočjih in cestah, drevesa na območju La Pozza so nagnjena; zaledna razpoka se je povečala, zaradi premikov merilnih točk v bližini Pinede, ki so bile do danes stabilne, mislim na najslabše. (…) Gladino jezera želimo spustiti do 695 m, da bi tako povečali varnosti pas okrog jezera. Do današnjega večera bi morali doseči 700 m. (…) Žal mi je, da ti pošiljam toliko slabih novice in prekinjam tvoj dopust. (…) Bog nam pomagaj."*

Približno ob dvanajstih so delavci ENEL, ki so malicali vrh jezu, videli večji premik.

Uro kasneje se je na levem bregu jezera pojavila 5 m dolga in 50 cm široka razpoka, ki se je vidno širila. Ob štirih popoldan je bila že meter široka. Na cesti okrog jezera so nastale nove razpoke. Do večera se je kamniti zdrs premaknil za 20 cm, hitrost premikov pa je naraščala.

Delavec, ki je med 15. in 16. uro prečkal goro Toc, je videl, kako se drevesa nagibajo in padajo.

ENEL-SADE je približno ob 17. uri ukazal Carusu, naj pokliče policijo. Na ogroženem območju v dolini Vajont so prepovedali ves promet.

Ob 17:50 je Biadene v telefonskem pogovoru z geologom Pento iz Rima potrdil, da jezerska gladina pada in da je potrebno doseči raven 700 m. Penta je predlagal, naj se izogibajo paniki, "*preden bi za to imeli dobre razloge*."

Ob 18. uri je Biadene zapustil jez in se vrnil v Benetke. Ritmeyer z ostalo posadko pa je ostal na jezu in nadaljeval z opazovanjem razmer. Zvečer so prižgali reflektorje, v stavbi pa odstranili okno, da bi zaposlenim olajšali zasilni izhod. V primeru zdrsa naj bi skočili iz stavbe in se po potki zatekli v zatočišče, ki je bilo izvrtano v goro.

Tovornjaki, ki so popravljali dovozno cesto do elektrarne, ob 20. uri niso več mogli voziti po njej, zato se je ENEL-SADE odločila, da jo zapre. Medtem se je Caruso v Bellunu srečal s poveljnikom policije in ga zaprosil za popolno zaporo prometa po cesti med naseljema Ponte nelle Alpi in Castellavazzo. Policist je privolil, vendar prebivalstvo o tem ni bilo obveščeno.

Ob 22. uri Rittmeyer je poklical Biadeneja in ga opozoril, da se je gora začela podirati ter vprašal za navodila. Biadene ga je poskušal pomiriti. Dogovorila sta se o evakuaciji naselja Spesse, ki je ostalo edino poseljeno v nevarnem območju (729 m).

Ob 22:38 uri se sproži kamniti zdrs. 270 mio. m^3 pretrte, a kompaktne kamnine glasno zgrmi v jezero. To povzroči izpad električnega toka in kratke stike v celi dolini, ki jo osvetlijo bliski. Bliske so videli tudi v Longaroneju.

Don Onorini, ki je bil duhovnik v Cassu, je ob tem času delal v svoji pisarni. Slišal je grom in stekel k oknu. Videl je, kako je 100 m visoko in 2000 m široko pobočje gore Toc s hitrostjo približno 100 km/h drvelo proti jezeru. Udar gmote je izpraznil jezero in odrinil jezernico. Nastala sta dva tokova. Prvi je stekel po dolini navzgor proti Ertu, ki je ostal praktično nepoškodovan razen nekaj najnižjih hiš.

V tem času je vodni val segel tudi približno 250 m visoko na desno pobočje in dosegel vas Casso. Tam vodni val ni povzročil večje škode, nevarne pa so bile številne skale, ki so letele po zraku. Ena od njih, ki je predrla cerkveno streho, je danes podlaga ambonu. Vodna gmota, ki je obsegala približno 50 mio. m^3 vode, je nato padla nazaj na gmoto in močno erodirala njegov severni del, kar lahko vidimo še danes.

Večji vodni val se je usmeril po dolini navzdol. Na ustju doline Vajont je bil vodni val visok 70 m, potoval pa je s hitrostjo 100 km/h. Cesta vrh jezu je bila uničena, sicer pa je jez ostal nepoškodovan. Deloma je bila poškodovana tudi cesta Longarone-Vajont. Voda je pljusknila tudi skozi predore, ki so ostali nepoškodovani. V njih je voda, ko se je njen tok umiril, odložila normalne rečne sedimente.

V Longaroneju niso vedeli, kaj se dogaja v dolini Vajont. Približno minuto po sprožitvi kamnitega zdrsa je v dolini Piave zavel lahen veter. Ta pojav v angleščini imenujejo "*underground-train effect*" in je nastal, ko je pritekajoča vodna gmota odrivala zrak pred seboj. Hitrost vetra je zato naraščala, povečala se je tudi vlažnost zraka. Veliko ljudi je sedelo pred televizorji v gostilnah, saj so prek televizijske postaje RAI prenašali ponovitev nogometne tekme med Realom Madrid in Rangersi iz Glasgowa. V naselju je nastala panika, ko je veter postal orkanski. Vrata in okna so začela vibrirati in se odpirati. Zatresla se je tudi zemlja. Nekateri so poskušali pobegniti, se vračali v hiše, da bi rešili družine ali priti do avtomobilov, drugi pa so se vzpenjali v pobočje. Ko je vodni val dosegel sredino doline Piave, ki je na tem mestu široka nekaj manj kot 2 km, so naselje že dosegle številne skale. Zaslišal se je grom, strešniki so leteli s streh. Tik pred vodnim valom je bil zrak tako stisnjen, da je imel veliko udarno moč, ki je lahko predmete dvigovala v velike višine. Vodni val je porušil, zradiral s površja zemlje, tako rekoč vse stavbe v Longaroneju. Tiste, ki so preživeli vodni val, so ubili padajoče kamenje ali rušenje zidov. Voda je dosegla tudi večino tistih, ki so se vzpenjali v pobočje, toda ker je že izgubila večino svoje moči, jih je le dvignila na višjo lego. Začel se je povratni tok, ki je dosegel še naselja Castellavazzo, Pirago, Rivalta, Villanova in Fae. Reka Piave je narasla za več kot 10 m.

VIR: http://www.geocities.com/Athens/Acropolis/2907/vajont.html

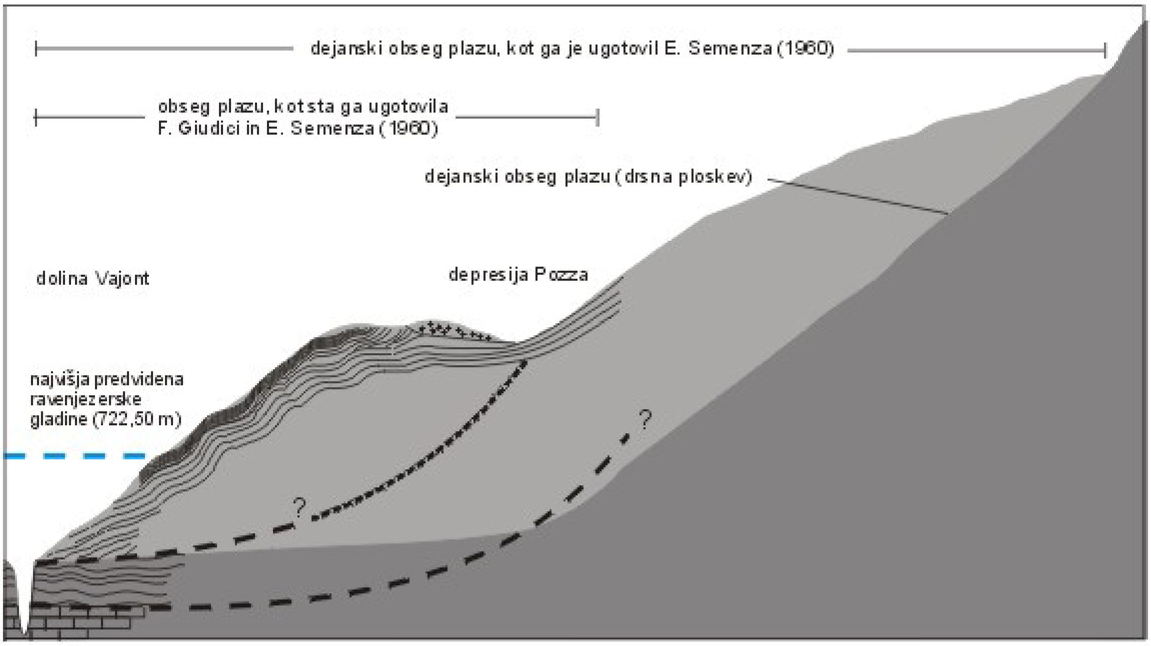
Moj oče, pesem osnovnošolca...

Oče, daleč proč
je delal,
dežural je za božič,
in jaz sem čakal.

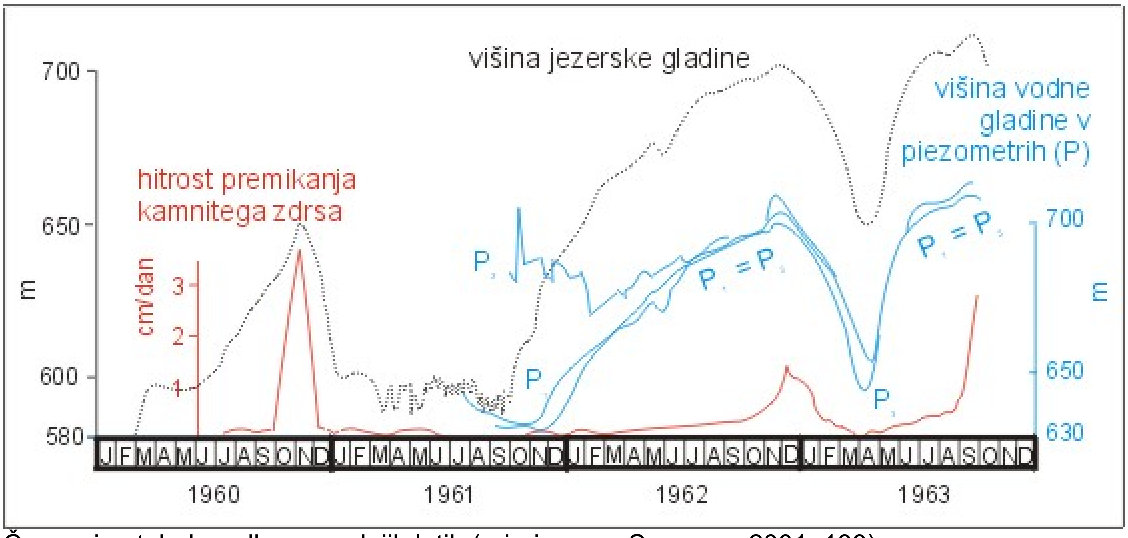
Vrnil se je pred božičem,
vendar ga nisem mogel počakati;
odšli smo stran -
pred božičem -
babica, mami in jaz sam.
Nismo krivi, oče, kličem.
Večno te bomo imeli radi.

(Vajont - not to forget, 2002, 67; prevod pesmi iz angleškega jezika: B. Komac)

Slikovno gradivo:



Reaktivacija starejšega zdrsa (prirejeno po Semenza 2001, 76).



Časovni potek dogodkov v zadnjih letih (prirejeno po Semenza 2001, 133).

Potem...

Lokalne oblasti na nesrečo niso bile pripravljene, vendar je kmalu stekla obsežna reševalna akcija. Prihajala je pomoč z vsega sveta. Naselje so sprva nameravali na novo postaviti drugje, vendar so preživeli zahtevali obnovo na istem mestu. Že čez nekaj mesecev so postavili začasne stavbe (pošto, zdravstveni dom in cerkev), kar je omogočilo vrnitev prebivalcev. Obnova je bila dolgotrajna tudi zaradi polemik o načinu gradnje. Do leta 1973 so izdelali kar 24 različnih urbanističnih planov, zaradi česar je Longarone danes povsem drugačen od sosednjih mest. Veliko stavb je zgrajenih iz betona, posebno znane so vrstne hiše v zahodnem delu naselja, šola in cerkev. Cerkev so postavili leta 1983. S spiralno zasnovo spominja na turbino, jez oziroma vodno ujmo in je poglavitni spomenik nesrečnikom. Spominska obeležja so še v Fortogni (pokopališče) in Piragu, kjer kot spomenik stoji cerkveni zvonik, ki je v celem naselju edini prestal katastrofo. Naselja so sčasoma ponovno zaživela, v dolini Piave so usposobili veliko industrijsko cono. Longarone skupaj z Bellunom danes slovi po izdelovanju očal.
Katastrofe se spominjajo vsako leto 11. oktobra. Priprave na spominsko slovesnost ob letošnji štirideseti obletnici so se začele že poleti.

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by Piero Elter[1], Mario Grasso[2], Maurizio Parotto[3], and Livio Vezzani[4]

Structural setting of the Apennine-Maghrebian thrust belt

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The Apennine-Maghrebian fold-and-thrust belt developed from the latest Cretaceous to Early Pleistocene at the subduction-collisional boundary between the European and the westward-subducted Ionian and Adria plates. Large parts of the Mesozoic oceanic lithosphere were subducted during an Alpine phase from the Late Cretaceous to Middle Eocene. The chain developed through the deformation of major paleogeographic internal domains (tectono-sedimentary sequences of the Ligurian-Piedmont Ocean) and external domains (sedimentary sequences derived from the deformation of the continental Adria-African passive margin). The continuity of the Apennine chain is abruptly interrupted in the Calabrian Arc by the extensive klippe of Kabylo-Calabrian crystalline exotic terranes, derived from deformation of the European passive margin.

Major complexities (sharp deflections in the arcuate configuration of the thrust belt, out-of-sequence propagation of the thrusts) are referred to contrasting rheology and differential buoyancy of the subducted lithosphere (transitional from continental to oceanic) and consequent differential roll-back of the Adria plate margin, and to competence contrasts in the Mesozoic stratigraphic sequences, where multiple décollement horizons at different stratigraphic levels may have favored significant differential shortening.

From the Late Miocene, the geometry of the thrust belt was strongly modified by extensional faulting, volcanic activity, crustal thinning and formation of oceanic crust correlated with the development of the Tyrrhenian Basin.

Introduction

The large-scale geometry of the Apennine-Maghrebian chain is that of an arcuate thrust belt with convexity towards the Adria-Africa foreland. Nested arcs of different size and curvature show a progressive change from the WNW-ESE trends of the Torino and Monferrato hills, to the Ferrara fold-and-thrust belt beneath the Po Plain, the NNW-SSE trends of the Marche and Abruzzi segment, the NW-SE trends in Molise-Puglia-Lucania, and the N-S trends in Calabria, which gradually deflect E-W in Sicily (Figure 1). The Apennine-Maghrebian fold-and-thrust belt developed from the latest Cretaceous to the Early Pleistocene at the subduction-collisional boundary between the European and the westward-subducted Ionian and Adria plates. Large parts of the Mesozoic oceanic crust were subducted during an Alpine phase, from the Late Cretaceous to Middle Eocene; starting in the Oligocene, continental collision of the European margin occurred against the Adria-Apulia-African margin. From the Late Miocene, the geometry of the thrust belt was strongly modified by extensional faulting, volcanic activity, crustal thinning and formation of oceanic crust in the southern Tyrrhenian Sea.

The Apennines comprise a stack of Adria-verging thrust sheets bounded by a complex system of frontal arcs, which overlie with a festoon-like pattern the upper Pliocene-lower Pleistocene terrige-

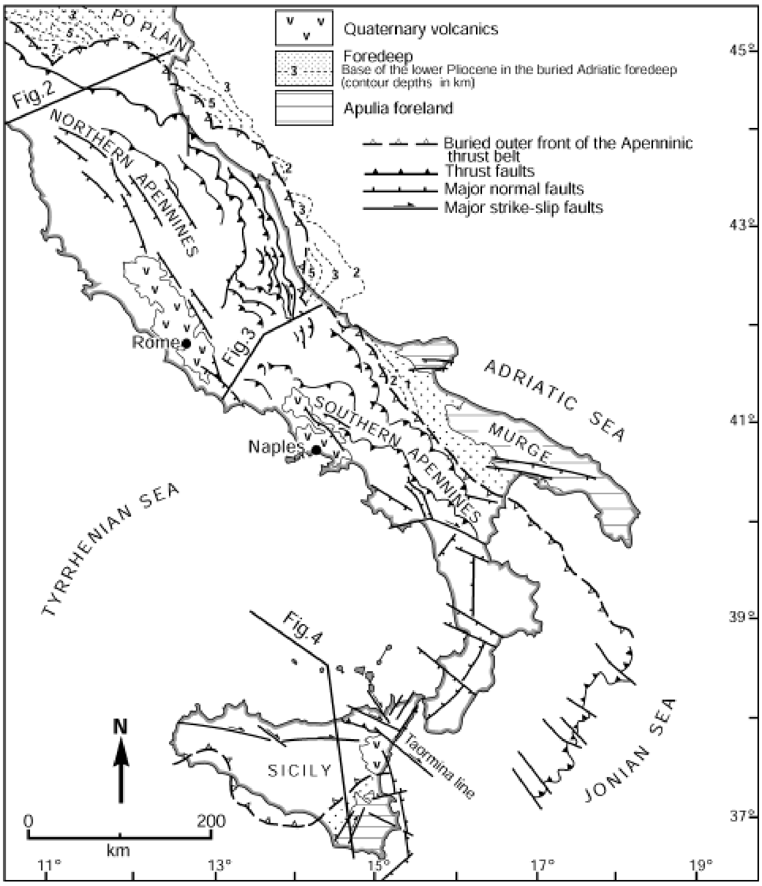

Figure 1   General structural map of the Apennine-Maghrebian chain. Fig.2, Fig.3, Fig.4: locations of the cross-sections of Figures 2–4. (After Ghisetti and Vezzani, 1999, modified)

nous sequence of the Adriatic foredeep and the slightly deformed margin of the Adria foreland. The Maghrebian chain in Sicily shows a stack of thrust sheets verging toward south, where part of the Hyblean foreland crops out.

Within the Apennine chain, tectonic segmentation and changes in structural trends are controlled by partitioning of thrusting and strike-slip transfer along transverse discontinuities connected with thin-skinned differential rotations. The degree of shortening varies irregularly according to the inherited paleogeography, contrasting rheology and differential sinking and roll-back of the subducting plate.

The chain developed through the deformation of two major paleogeographic domains: the *internal domain*, i.e. Late Jurassic to Oligocene tectono-sedimentary sequences of the Ligurian-Piedmont Ocean, which originally was linked to the Tethyan Sea, and the *external domain*, i.e. Triassic to Early Miocene sedimentary sequences derived from the deformation of the continental Adria-Africa passive margin.

The continuity of the Apennine-Maghrebian chain is abruptly interrupted in the Kabylo-Calabrian Arc by huge volumes of crystalline basement rocks and related Mesozoic-Paleogene carbonate covers thrust over Cretaceous to Miocene basinal sequences, belonging to the Liguride Units in northern Calabria and to Sicilide Units in Sicily.

This paper attempts to synthesize the content of a large volume of published papers; due to the breadth of the discussed topic, quoted references are not comprehensive but were selected to guide readers through literature.

The Kabylo-Calabride terranes

The orogenic hinterland mostly consists of metamorphic Calabride basement units, largely submerged offshore northern Sicily but cropping out in northeast Sicily (Peloritani Mts.) and in Calabria, and linked westwards to the Kabylies of North Africa.

These exotic terranes, referred to as Calabride units, are located at the intersection between the NW-SE-trending southern Apennines and the E-W-trending Sicilian Maghrebides. They are characterized by a pre-Mesozoic crystalline basement, and show evidence of pre-Alpine tectonism and a wide range of metamorphic processes (Bonardi et al., 2001). In the Peloritani Mountains (Sicily) and Calabria, several tectonic units are believed to derive from a former "internal massif" consisting of crystalline terrains (with metamorphic grade increasing from outer to inner zones) transgressively covered by different Mesozoic to Tertiary sedimentary sequences characterized by thinning and later subsidence toward the interior. In Calabria, the crystalline nappes and their related non-metamorphic Mesozoic-Paleogene carbonate covers were thrust northward onto the Liguride ophiolitic unit. In Sicily, the front of the Calabride units, which were thrust onto the Cretaceous-Miocene basinal sequences of the Sicilide Complex, has been traced across the Nebrodi-Peloritani chain from the Tyrrhenian Sea to the Ionian Sea along the Taormina Line (Figure 1).

Internal domain

This domain includes the Liguride units and Sub-Liguride units that crop out extensively in the northern Apennines, western Alps, and in the southern Apennines and Sicily, where the latter are described as Sicilide units.

The Liguride and Sicilide units experienced "Alpine" tectonics before being thrust onto the domains of the Adria-Africa continental margin. This tectonic phase leads to the Late Cretaceous-Middle Eocene closure of the Liguride-Piedmont oceanic basin, probably in relation to east-dipping subduction. The subsequent thrusting of the Liguride, Sub-Liguride and Sicilide units onto the outer domains was due to "Apennine" tectonics, which developed during Oligo-Miocene west-dipping subduction, and to continent collision connected with the migration of the Sardinia-Corsica continental block and opening of the Balearic Basin.

Liguride units

The northern Apennine Liguride units are ascribed to two different paleogeographic areas, one Internal (IL) and the other External (EL). The IL units are characterized by a basement mainly consisting of serpentinized peridotites, regarded as exhumed lithosphere, intruded by gabbros in the Permian, i.e. before the opening of the ocean. This basement (peridotites + gabbros) was exhumed in the Late Jurassic up to the sea floor. The overlying volcano-sedimentary sequence includes basalts and ophiolitic breccias topped by Late Jurassic to Late Cretaceous radiolarites, Calpionella-bearing Limestones and Palombini Shales. The latter formation is overlain by Campanian-Early Paleocene siliciclastic turbidites (Val Lavagna Shales and Gottero Sandstones) representative of a deep-sea fan fed by the European continental margin. Early Paleocene ophiolite-bearing debris flow deposits, fed by an Alpine accretionary wedge, represent the last sedimentary deposits preserved in the IL units.

The EL units are characterized by thick successions, mainly Late Cretaceous carbonate turbidites (Helminthoid Flysch), in which the ophiolites only occur as slide blocks or as fragments in coarse-grained deposits. These turbidites are overlain, mainly in the easternmost areas, by carbonate turbidites of Paleocene-Early Eocene age. Helminthoid Flysch is characterized by basal complexes consisting of coarse-grained clastic deposits of Albian-Campanian age; these deposits are ophiolite bearing in the westernmost areas, whereas they are fed by a continental margin in the easternmost ones. Although all EL unit successions are detached from their basement, the basal complex in the westernmost areas shows evidence of a basement: an ocean-continent transition characterized by the association of sub-continental mantle, granulites and continental granitoids s.l. By contrast, the analysis of basal complexes in the easternmost areas reveals that they were fed by the Adria continental margin.

The IL and EL units are characterized by a different structural history (Figure 2). The IL units display a west-verging evolution in the Alpine accretionary wedge. This evolution predates the eastward thrusting over the EL units, which are characterized by mainly east-vergent deformation. The Middle Eocene-Miocene deposits of the Epi-Ligurian Basin, a thrust-top basin above the Liguride units, seal the contacts among IL and EL units.

In the southern Apennines the Liguride units (also referred to as "Liguride complex") consist of a Mesozoic-Paleogene deep-water sequence interpreted as a detached sedimentary cover of the Liguride-Piedmont oceanic crust. The sequence has been subdivided into the Frido and Cilento tectonic units. The lowermost Frido Unit underwent a HP/LT subduction-related event followed by a greenschist-facies re-equilibration; this unit, cropping out in southern Lucania and northern Calabria, was thrust above the limestones of the Apennine platform and lies beneath the Cilento Unit. The Frido metasedimentary sequence consists of a highly variable alternation of shales, quartzarenites, and silty and arenaceous limestones of Neocomian-Albian age; it includes slices of Late Jurassic-Early Cre-

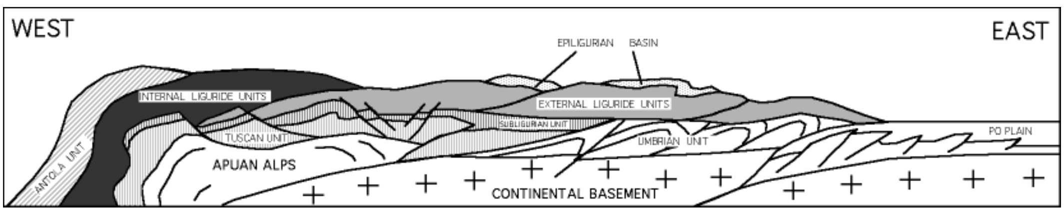

***Figure 2** Schematic cross-section showing the geometric relations among the major structural units of the northern Apennine. For location, see Figure 1. (P. Elter)*

taceous pillow lavas, diabase breccias, radiolarian cherts, jaspers and cherty limestones.

The uppermost Cilento (or Flysch Calabro-Lucano) sequence crops out from the Tyrrhenian coast to the Ionian slopes of Mt. Pollino. It includes a basal Crete Nere Fm, which consists of prevailing black shales alternating with siliceous calcilutites, marls and graded quartzarenites of Aptian-Albian age. The Pollica-Saraceno Fm lies above, i.e. turbiditic calcarenites and lithic-arkosic sandstones alternating with calcilutites and local conglomerates of Cenomanian to Paleocene age. The Cilento sequence is unconformably covered by the Albidona-S. Mauro Fm, which consists of 2,000 m of alternating silty-argillaceous marls in beds of up to 10 m, well-bedded sandstones with megabeds of calcilutites, and conglomerates with crystalline, calcareous clasts and predominant matrix; the age of this formation is still debated (Early-Middle Eocene, Vezzani, 1966; Baruffini et al., 2000; Miocene, Bonardi et al., 1985).

Sub-Liguride units and Sicilide units

Sub-Liguride units occur between the Liguride and Tuscan-Umbrian units (external domain, described later). The Sub-Liguride units display successions characterized by Late Cretaceous-Early Eocene shales and carbonates, showing Ligurian affinity, and arenites and conglomerates of Late Eocene-Early Oligocene age; the latter are characterized by andesitic clasts probably connected with Alpine subduction.

In the central-southern Apennines and Sicilian thrust belt, the Sicilide units (also known as "Sicilide Complex") consist of a non-ophiolite-bearing, varicolored pelitic sequence of intensely deformed, Late Cretaceous-Early Miocene deep-marine sediments. The sequence includes a red and green basal pelitic member with intercalations of cherty limestones and quartzarenites (Mt. Soro Fm), which gives way above to alternating calcarenites, calcirudites and marly limestones (Pomiere facies, in Sicily, and Mt. Sant'Arcangelo facies, in Lucania), and to alternating andesitic tuffites and tuffitic sandstones, marly shales and marly limestones of Oligocene-Early Miocene age (Tusa facies, in Sicily and Lucania).

A large part of this varicolored sequence (the so-called "Argille scagliose") prevailingly crops out at the boundary between the Apennine thrust front and the Po Valley-Adriatic-Ionian and Catania-Gela foreland basins. Note that the attribution of this varicolored sequence to the Lagonegro succession (see External domain) rather than to the Sicilide units is in many cases matter of debate. This strongly deformed pelitic sequence constitutes the matrix of a fragmented formation, which derived from polyphase deformation of original pelitic, calcareous and arenitic multi-layered sequences along the Apennine accretionary frontal prism. This tectonic mélange includes slices of different size of resedimented calcarenites and calcilutites, cherty limestones, and quartzites pertaining to the Late Cretaceous section of the internal units, as well as fragments of Early Miocene Numidian quartzarenites and Tusa tuffites.

External domain

The large-scale structure of the entire Apennine Maghrebian chain is characterized by the thrusting of the Liguride, Sub-Liguride and Sicilide units onto the outermost domains, i.e. Tuscan and Umbria-Marches units in the northern Apennines, Latium-Abruzzo-Molise units in the central Apennines (Figure 3), Daunia-Lucania units in the southern Apennines and Mt. Iudica-Sicani Mts. in Sicily; as a whole, these units occupy the lowermost position in the thrust belt.

The Meso-Cenozoic stratigraphic successions outcropping in the external domain mainly accumulated along the Adria-Africa passive continental margin. The successions developed through a combination of geological processes. Of these, the most important were crustal extension, the cyclic production of marine carbonates and sea-level variations. The most ancient deposits, representing a long Late Triassic depositional phase in evaporitic to restricted-marine (dolomites with anhydrite levels) environments, directly onlap Permian continental deposits. A vast carbonate platform of regional extent began to develop at the start of the Jurassic. Subsequently, still in the Early Jurassic, the entire area experienced a rifting phase, which gave rise to a complex marine topography with various (especially carbonate) depositional environments.

Carbonate platform successions

The continental shelf deposits were characterized by the development of isolated peritidal carbonate platforms, pelagic basins and, locally, of pelagic carbonate platforms (portions of flooded peritidal platforms covered by condensed, discontinuous pelagic carbonate successions, such as the Sabine Plateau, in Latium). The strong topographic control of sedimentation ceased in the Early Cretaceous and was substituted by general natural processes (sea-level variations, currents, changing subsidence velocities, synsedimentary tectonics, etc.), which produced large lateral variations in carbonate sedimentation.

Remnants of a vast Apennine carbonate platform (or perhaps of several platforms separated by seaways) have been divided into several tectonic units that constitute the bulk of the central-southern Apennine thrust belt. The remnants of the Apulian carbonate platform, which acted as a foreland and was only partially involved in orogenic deformation, crop out east of the Apennine chain. Minor remnants of carbonate platform also outcrop in the Palermo and Madonie Mts. (northwestern Sicily). This succession consists of Late Triassic-Jurassic-Cretaceous reefal carbonates overlain by Late Cretaceous-Eocene wackestones and red marls exhibiting a typical Scaglia facies, Oligocene fine-grained marls, quartzarenites and calcarenites. Along the southern border of the Madonie Mts., this carbonate platform is characterized by swarms of platform carbonate blocks and megabreccias embedded within the brown shales of the Numidian Flysch. A platform carbonate sequence resting on volcanic seamounts also crops out in the Hyblean Plateau (southeastern Sicily), where it acted as the foreland of the Maghrebian thrust belt.

Environmental changes have continuously influenced the evolution of platforms in the Apennine-Maghrebian chain: in the Cenomanian, the breakup and flooding of the former Bahamian-type platforms gave rise to highly productive margins controlled by faults. As a result, the inner platform areas diminished, with the development of vast carbonate ramp systems which linked amply emerged portions of ancient platforms to the surrounding pelagic basins (Parotto

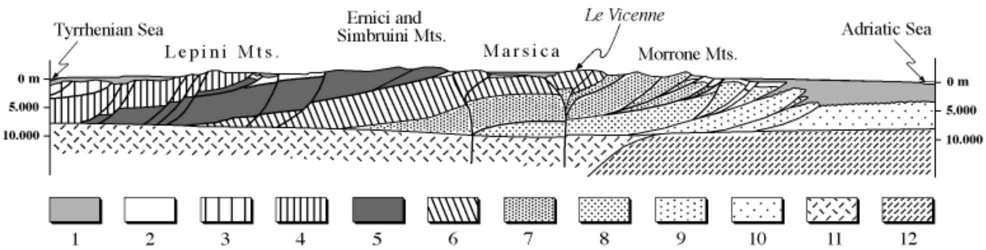

***Figure 3** Schematic cross-section of the central Apennine thrust belt. For location, see Figure 1. 1. marine and continental post-orogenic sedimentary cover (Tyrrhenian side) and intermountain basins (Late Messinian-Quaternary); marine syn- and post-orogenic deposits (Adriatic side: early Pliocene-Quaternary); 2. syn-orogenic deposits (late Tortonian to early Pliocene); 3–9. tectonic units mainly derived from the external domain (carbonate platforms and basins); 10. Adriatic foreland; 11. magnetic basement of the thrust belt; 12. magnetic basement of the Adriatic foreland. (From Cipollari et al., 1999)*

and Praturlon, 1975). Starting in the Middle Miocene, shallow water calcarenites unconformably or paraconformably overlay the Cretaceous limestones (the so-called “Paleogene hiatus”, well known in the central-southern Apennines) of the ancient, Mesozoic carbonate platforms; siliciclastic turbidites deposited in later Miocene-early Pliocene above the middle Miocene calcarenites.

Pelagic Basin successions

Pelagic basins developed around and between the platforms. The Sabine Basin opened to the west of the Apennine platform and was linked to the Tuscan Basin (the Sabine Plateau lay within these basins). The Umbria-Marche Basin lay to the north and was also linked to the Sabine and Tuscan basins. The Lagonegro-Molise Basin opened between the southern Apennine and Apulian platforms, while an outer basin (Ionian Basin) opened east of the Apulian platform. In Sicily an inner basin (Imerese Basin), which may be correlated with the Lagonegro Basin, widely crops out in the northern part of the island, while an outer basin (Sicani Basin) opened in the southwestern part of the island.

In the northern Apennines, the basin successions from Tuscany and Umbria-Marche started with a transgressive event (Triassic Verrucano-facies conglomerates, evaporites and dolostones covered by platform carbonates of Liassic age), followed by progressive sinking marked by the Rosso Ammonitico-facies deposits, older in the Tuscan zone (Sinemurian) than in the Umbria zone (Aalenian). The deepest pelagic deposits are represented by Upper Jurassic-Lower Cretaceous radiolarites and pelagic limestones (Maiolica facies), and are coeval with the older sedimentary deposits found in the Liguride-Piedmont oceanic basin.

The Umbria-Marche succession merges southward into the Sabina succession, in which limestones, marly limestones, marls and cherty levels alternate with frequent intercalation of resedimented rocks derived from the carbonate platform margin.

The paleogeographic features of the platform-basin system in the northern-central Apennines remained the same through to the Oligocene, when the structuring of the Apennine orogen had already begun. The inception of flexuring of the Adria continental margin in the Middle-Late Oligocene led to the development of the foredeep basin system, which was filled by thick siliciclastic turbiditic bodies. Infilling progressively shifted from internal to external zones (from the Oligocene for the Tuscan zone to the Early Pliocene for the outermost peri-Adriatic zones) due to the progressive migration of the orogenic belt-foredeep couple.

The evolution of the Lagonegro-Molise Basin (southern Apennines) and of the Imerese-Sicani Basin (central-north Sicily) was rather different. The basinal sequences of the Lagonegro and Imerese basins show a transition from terrigenous-carbonatic facies of coastal to shallow-water environments (Early Triassic-Anisian), to pelagic cherty-radiolaritic facies (Ladinian) followed by cherty limestones of Late Triassic age, dolomites and by a Jurassic-Cretaceous radiolarites succession with mafic volcanics and more or less pronounced hiatuses. The overlying Middle-Late Eocene to Early Oligocene sequence is composed of interbedded red marls and graded calcarenites with macroforaminifera. On it rests the alternation of quartzarenites and clays of the Numidian Flysch, which represents the earliest Late Oligocene to Middle Miocene filling of the precursor foredeep basins established after the collision between the African and European continental plates. In the outermost zones of the Apennine chain (e.g. the Molise Basin, Daunia, and Lucania “external zones”) and in Sicily, the Numidian Flysch is overlain by alternating marly limestones, calcarenites and calcirudites with reworked upper Miocene macroforams (e.g. Tufillo Fm, Masseria Palazzo Fm) grading to Tortonian (in part)-Messinian marls (*Orbulina* Marl Fm). These are followed by the siliciclastic turbidites of the Agnone Flysch in Molise and the Masseria Luci Flysch in Lucania, representing the Messinian stage of the eastward migration of the foredeep basin, which shifted up to the Bradanic-Gela-Catania foredeep in the Early Pliocene.

The reconstructed setting suggests that Numidian Flysch deposited over a large basin, the external side of which was represented by a still undeformed African plate margin. The internal margin of the Numidian Flysch basin is more difficult to reconstruct because of subsequent intense deformation and crustal shortening during the formation of the Apennine-Maghrebian chain. Although most of the original stratigraphic contacts between the Numidian Flysch and its substratum are overprinted by later tectonic detachments, it is still possible to infer the stratigraphic substratum of the Numidian Flysch, represented by both platform and pelagic basin carbonates and by deformed successions of the Sicilide domain.

Epi-Liguride sequence (piggyback basins, Ori & Friend, 1984; satellite basins, Ricci Lucchi, 1986; thrust-top basins, Butler & Grasso, 1993)

This is the well-known Oligocene-Pliocene lithostratigraphic sequence comprising Monte Piano Marls, Ranzano Fm, Antognola Fm, Bismantova Group, Termina Fm and Gessoso-solfifera Fm. The sequence is characterized by relatively deep-marine deposits, with episodes of shallow marine and transitional-continental (lagoon and fan delta) deposition, which unconformably cover the already deformed Liguride and Sub-Liguride allochthon of the Piedmont-Liguride and Emilian thrust belt.

In the central Apennines, correlated thrust-top sequences (Rigopiano, Monte Coppe, Calaturo) of early Pliocene age unconformably cover the carbonatic sequences of the Gran Sasso and Mt. Morrone thrust belt.

In the southern Apennines, many Middle-Late Miocene clastic deposits (e.g., Gorgoglione Fm and Oriolo Fm in Lucania, Anzano Fm in Puglia, Valli Fm in Molise), followed by the Messinian Gessoso-solfifera Fm and lower Pliocene clayey conglomerate sequence (Panni in Puglia, Larino in Molise), unconformably cover both the Sicilide allochthon and its substratum, mainly represented by the Lagonegro-Molise units. Thrust-top basins are also present in the Calabrian Arc (Crotone and Spartivento basins) and Sicily (in the northern part of the Caltanissetta Basin), where the Late Miocene Terravecchia Fm represents a clastic sediment deposited above and adjacent to growing thrusts and folds. Towards the chain, the Terravecchia Fm lies directly above thrust structures, thus representing the infill of one or more basins perched on thrust sheets. Two major Messinian evaporitic successions, separated by regional erosional and/or angular unconformities, and the Early Pliocene Trubi chalks were involved in the thrust-fold belt of central Sicily (Decima & Wezel, 1971; Butler et al., 1995a).

In the northern Apennines and Sicily, these sequences are characterized by several chaotic resedimented breccia bodies related to submarine mass gravity transport of material derived from the Liguride and/or Sicilide substratum (“Argille Brecciate”).

Large-scale tectonic features of the Apennine-Maghrebian thrust belt

The Apennine-Maghrebian chain as a whole is characterized by the superposition of two major geometric units that configurate a regional, east-verging duplex structure separated by a low-angle, west-dipping regional thrust system. This allochthonous edifice tectonically overrides the Adria-Hyblean foreland, as well documented at and below the surface by seismic and drilling exploration (Mostardini & Merlini, 1986).

The uppermost tectonic element consists of allochthonous Liguride, Sub-Liguride and Sicilide nappes, which involve Mesozoic-Cenozoic sedimentary sequences and ophiolitic suites derived from deformation of the internal domains. Prior to thrusting, these units were more or less involved in Alpine tectonics. The upper part of the Liguride complex in the westernmost areas and Elba Island followed a meso-Alpine, European-verging evolution before being thrust above the domains of the Adria continental margin. In turn, the lower complex, mainly represented by External Liguride units, was affected by a Middle Eocene tectonic event. As a whole, the Liguride-Sicilide stack shows a foreland-dipping geometry and a thin-skinned imbricate structure.

The underlying tectonic element is represented by the outer foreland fold-and-thrust belt, consisting of tectonic units derived from the deformation of the Adria margin, i.e., the Tuscan-Umbria-Marche units of the northern Apennines, the Latium-Abruzzo and Lagonegro-Molise units of the central-southern Apennines, and the Panormide-Imerese-Sicanian units of Sicily. The large-scale tectonic structure of the northern Apennines can be clearly observed in the Apuane Alps window, where a complete section of these tectonic units crops out. The lowermost tectonic unit is the low-grade metamorphosed Tuscan unit, represented by a Triassic to Oligocene sedimentary cover involved, along with slices of Paleozoic basement, in large-scale structures. This unit was overridden by the unmetamorphosed Tuscan unit, only represented by the Triassic to Miocene sedimentary cover, which detached along the Triassic evaporites. In the central-southern Apennines and in Sicily, the Tertiary sequences of these external units were decoupled from their Mesozoic substratum and pushed, together with the overlying Sicilide and Liguride units, to form the outermost imbricate thrusts that lie directly above the Bradano-Gela-Catania foredeep and the Apulia-Hyblean foreland.

A further major geometric unit at the top of the Apennine-Maghrebian chain is represented by the extensive klippe of Kabylo-Calabride crystalline exotic terranes derived from deformation of the European passive margin, which overrode both the Liguride-Sicilide nappes and the outer foreland fold-and-thrust belt. These units are submerged in the Tyrrhenian Sea. In the Peloritani Mts. and Calabria (Calabrian Arc), the crystalline nappes and their related Mesozoic-Paleogene carbonate covers are thrust over Cretaceous to Miocene basinal sequences deposited in oceanic and/or thinned continental crust, which was consumed during the early phases of the collision. Most of the arc lies offshore, and its structure and geometry have been mainly reconstructed through the analysis of available multichannel seismic profiles (Finetti, 1982; Finetti and Del Ben, 1986). A series of thrusts, progressively more pronounced in the central sector of the arc, affect the sedimentary sequences of the Ionian Basin. Seismic data highlights a prominent shear surface that progressively deepens toward the inner part of the arc.

All these three major geometric units are dissected by strike-slip and normal faults that post-date thrust structures and in some cases control the opening of minor marine and/or continental basins.

Kinematic reconstruction

The large-scale tectonic evolution of the Apennine thrust belt was firmly constrained by the progressive eastward migration of the outer Apennine front, related to the opening of the Tyrrhenian Basin. The progressive shortening of fold-and-thrust belt is traced by the onset, evolution, deformation and progressive migration of Late Miocene to Early-Middle Pliocene siliciclastic foredeep deposits.

The three main steps in the contractional evolution of the Tyrrhenian-Apennine system have been reconstructed by Patacca et al. (1990).

Late Tortonian-Messinian (in part) rifting in the northern Tyrrhenian Sea, southwestern Tyrrhenian Sea and Gioia Basin was contemporaneous with the eastward shifting of the foredeep-foreland system. This migration can be followed from the Tuscany-Umbria (Macigno, Marnoso-Arenacea) to the Marches (Laga) foredeep basins in the northern Apennines, from the Latium (Frosinone, Torrice Flysch) to the Abruzzo (Laga, Gran Sasso Flysch) foredeep basins in the central Apennines, and from the Campania (Alburno-Cervati) to the Molise-Lucania (S. Elena, Agnone and Masseria Luci Flysch) foredeep basins in the southern Apennines. This foreland fold-and-thrust belt, which represents the lower panel of the Apennine duplex, is overridden by the Liguride, Sub-Liguride and Sicilide nappes, which are unconformably overlain by the thrust-top deposits of the Valli, Oriolo and Gorgoglione Fms of Tortonian-Messinian age.

During the *late Messinian-Pliocene (in part)*, extensional faulting affected the northern Tyrrhenian Basin and the western margin of the Apennine chain, as documented by syntectonic accumulation of Messinian "Lago-Mare" clastic deposits with evaporites, followed by lower Pliocene marine clays in southern Tuscany basins. In this interval, rift processes took place in the central bathyal plain of the southern Tyrrhenian Sea in connection with the opening of the Magnaghi-Vavilov and Issel basins. Extension was accompanied by eastward migration of the Apennine thrust, incorporation into the thrust belt of the former foredeep basinal areas, and eastward shifting of the upper Messinian-Pliocene foredeep siliciclastic deposits. Thrust-top basins filled with clastic deposits of late Messinian-Early Pliocene age developed in the southern Apennines (e.g. Potenza, Ofanto, Ariano Irpino) and Calabria (Crotone, Spartivento basins). During this interval, out-of-sequence thrusting connected with anticlockwise rotations was responsible for several major arcuate structures of the Apennine thrust belt (e.g., the Gran Sasso-Mt. Picca thrust, see Ghisetti & Vezzani, 1991; Olevano-Antrodoco-Sibillini Mts. thrust, see Cipollari & Cosentino, 1992).

During the *Pliocene (in part) -Quaternary*, extensional faulting migrated from the Tyrrhenian Sea to the internal margin of the Apennines, giving rise to the Lunigiana, Valdarno-Valdichiana, Mugello-Casentino, Valtiberina and Rieti basins. In the southern Tyrrhenian Sea, new rifting was responsible for the opening of the Marsili Basin southeast of the central bathyal plain. Along the Tyrrhenian margin of the southern Apennine chain, the eastward migration of extension and downfaulting produced the Volturno, Sele, Crati and Mesima basins, and was accompanied by a parallel migration of the thrust belt-foreland basin system. Several thrust-top basins preserved in structural depressions on the rear of the thrust front follow the arcuate setting of the northern and central Apennine belt from Piedmont to the Marches-Abruzzo. In the southern Apennines, large remnants of thrust top basins are preserved in Molise from Atessa to Larino and in Puglia-Lucania (Panni, S. Arcangelo).

In Sicily, the frontal thrust structures of the Maghrebian chain, involving strongly deformed Miocene to Pliocene sediments, are emplaced above Pliocene-Pleistocene rocks of the foreland margin (Butler et al., 1992). Along the margin of the bulged Hyblean foreland, normal faults accommodate flexural downbending (Figure 4). The Gela-Catania foredeep flanks the northern and western margins of the Hyblean Plateau, and extends offshore south-central Sicily. Within the Gela Nappe, the toe of a regional tectonic wedge coinciding with the Maghrebian thrust belt, compressional tectonics are reflected in folding and in thin-skinned thrusting, which post-dates the deposition of Pliocene sediments. North of the Gela Nappe, the Mt. Iudica imbricate thrusts consist of Mesozoic basinal carbonates and Miocene siliciclastics. Below the Mt. Iudica stack, the top of the impinging Hyblean bulge is no longer recognizable, but there is a dramatic change in the magnetic susceptibility of the basement in relation to a change in the carbonate substratum, i.e. the presence of a deep-seated duplex. Several thrust sheets consisting of Sicilide units and early foredeep deposits (Numidian Flysch), together with slices of their Mesozoic carbonate substratum, are detached from the basement. Small upper Miocene to Pliocene thrust-top basins lie above the thrust sheets. Other deformed Sicilide units are accreted at the junction between the Maghrebian chain and the Calabride-Peloritani units, representing the orogenic hinterland. The Aeolian volcanic arc developed along the southern margin of the Tyrrhenian Basin.

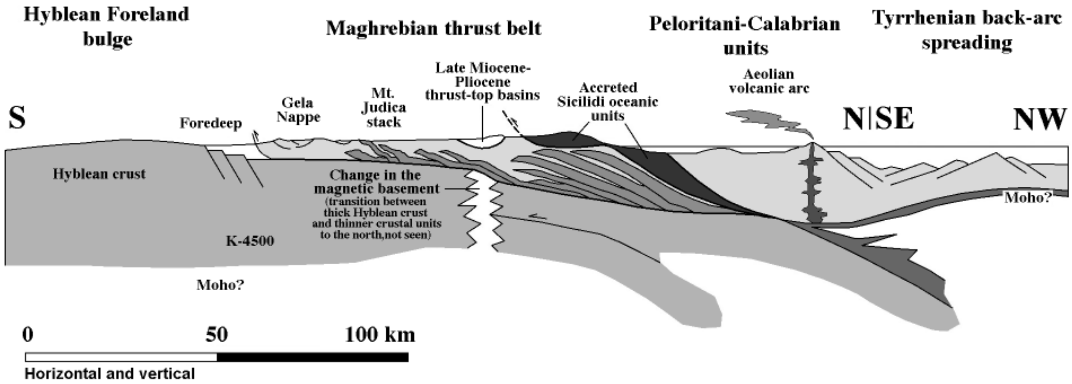

Figure 4 Simplified section across the eastern Sicily, from the Hyblean foreland to the southern Tyrrhenian back-arc basin; K indicates the magnetic susceptibility. For location, see Figure 1. (M. Grasso)

Open issues and discussion

The above traced evolution, which many authors have placed in the simple context of regular forward-migrating piggy-back imbrication of sedimentary units detached from a substantially undeformed crystalline basement (Bally et al., 1986), is in contrast with the observation that the leading thrust faults of the major tectonic units display different orientations, contrasting directions of tectonic transport, rotational emplacement trajectories, and out-of-sequence activation. The joint analysis of deformation styles, displacement gradients and age of shortening reveals that adjacent segments of the Apennine-Maghrebian belt, with contrasting competence, underwent coeval deformation through non-coaxial kinematics.

All these observations constrain palinspastic reconstruction, section balancing and evaluation of the degree of shortening, and suggest extreme caution in deriving deformational steps and the regional trajectory of stress fields from the kinematics of fault systems.

The amount of extension in the Tyrrhenian Sea, shortening of the Apennine thrust belt, rates of foredeep migration and flexure retreat in the foreland, greater in the southern Apennines than in the northern Apennines, suggest that a single process was responsible for the genesis of the couple Tyrrhenian Sea-Apennine chain.

The Sicilian segment of the chain has a large dextral wrench shear component associated with the opening of the Tyrrhenian Sea during the Neogene, and is affected by relative motion between the African and European plates. The uplifted carbonates exposed in the western segment of the chain suggest that passive-margin sedimentation continued through much of the Paleogene. However, from the Late Oligocene onwards, deposition was predominantly siliciclastic, thus representing a dramatic change to foreland basin sedimentation. The highly rifted nature of the Mesozoic African continental margin during Tethyan spreading and its compartmentalization into a number of sub-basins brought about deposition in foredeep settings which remained deep-marine through much of the early Middle Miocene.

As previously mentioned, at the end of the Oligocene, and especially in the Miocene, the successions of the inner domains and of the platform-basin system of the central Apennine were involved in the progressive development of a thrust belt verging towards Adria. During this process, strongly subsiding sedimentary basins (foredeep) repeatedly developed along the thrust front of the orogen due to the progressive flexure of the foreland margin. The basins were filled with essentially siliciclastic turbidite successions (fed by sectors of the Alpine chain experiencing strong uplift, including magmatites and metamorphites, and by local contributions from the developing Apennine orogen). The diachronism of the turbidite successions highlights the progressive eastward migration of the foredeeps, especially in the Neogene, up to the present Adriatic foredeep. The thrust fronts migrated in the same direction, gradually involving the deposits of the various foredeeps and incorporating them in the chain (Cipollari et al., 1995).

Starting about 7 Ma ago, while the Adriatic side of the chain was building up through compressional structures, intense extensional tectonics began to develop on the Tyrrhenian side. This extensional tectonic regime, which was directly correlated with the development of the Tyrrhenian Basin (further W), began to the west and migrated progressively eastward, involving a good portion of the chain. Its development led to the subsidence of entire sectors of the chain, which had only recently experienced uplift, through generally westward-dipping, high-angle normal fault systems (often reactivating, at great depths, the ramps of earlier thrust surfaces). The tectonic troughs, which consequently developed, accumulated thick marine (shallow water) to continental (fluvial, lacustrine) depositional sequences. Crustal thinning allowed the ascent of magma (both mantle-derived melts and magmas with varying degree of crustal contamination), which fed a chain of impressive volcanic edifices (with melts prevalently high in K) at the site of the western, older and more mature extensional basins.

The presence of an extensional regime in the internal sector of the central-Apennine orogen that compensates compression towards the foreland has often been attributed to continuing lithospheric subduction in the presence of diminished convergence between Europe and Adria. Models propose an upwelling of the asthenosphere and a contemporaneous passive descent of a slab of subducting Adriatic lithosphere, with progressive eastward migration of the subduction hinge. However, some studies suggest that the slab broke away and is sinking. Other researchers believe that there is no conclusive evidence of subduction. They propose, instead, the presence of an asthenolith produced by transformations of the lithospheric mantle and crust induced by thermal anomalies and fluids from deep mantle sources. Whatever the cause, current processes in the Apennine-Maghrebian chain seem to be in relation to general uplift and to the north-western migration of Africa and Adria with respect to stable Europe (Di Bucci & Mazzoli, 2002).

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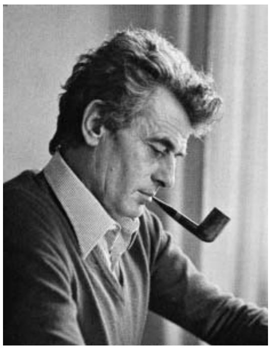

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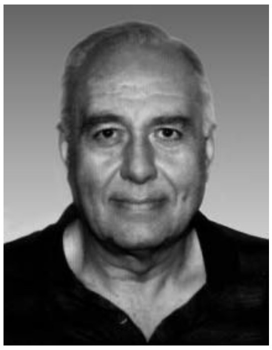

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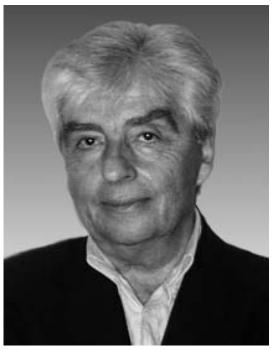


Ligurian ophiolites

Ligurian ophiolites are located in Northern Apennines and are interpreted as remnants of the lithosphere that floored the Jurassic Liguro–Piemontese ocean, a short-lived branch of the Tethys. In the Northern Apennines, ophiolites outcrop in two distinct geological units, the so-called Internal (IL) and External Ligurides (EL). The terms 'External' and 'Internal' refer to the inferred paleogeographic position of these units (i.e. pericontinental versus intraoceanic settings) in the Jurassic oceanic basin.

The Internal Liguride (IL) ophiolitic sequences have been ascribed to intra-oceanic settings. The IL ophiolites are considered remnants of the mature oceanic lithosphere of the Ligurian Tethys. They outcrop in Eastern Liguria and belong to the Bracco–Val Graveglia Unit which comprises ophiolitic sequences with their Upper Jurassic to Paleocene sedimentary cover. These ophiolites consist of depleted mantle peridotites (mostly serpentinites) intruded by a gabbroic body showing MORB affinity. The peridotite–gabbro basement is overlain primarily by N-MORB-type massive and pillow lavas, and a sedimentary cover consisting of ophiolitic breccias, Oxfordian–Callovian radiolarian cherts, Calpionella limestones and shale–limestones. A developed sheeted dike complex is lacking, but discrete dikes intruding serpentinites, gabbros and ophiolitic breccias, are common. The gabbroic body is mainly composed of ultramafic cumulates (plagioclase-bearing dunites), Mg-rich gabbros (troctolites and olivine gabbros), intermediate gabbros (gabbronorites), Fe-rich gabbroids (Fe-gabbros and Fe-diorites) and plagiogranites.

In the External Liguride (EL) Units, ascribed to marginal settings, ophiolites consist mostly of fertile lherzolites and MORB-type volcanics which occur as huge olistoliths within Cretaceous–Eocene flysch formations. The EL units are often associated with continental detritus of Hercynian granites and mafic granulites.

The Ligurian ophiolites are believed to represent an oceanic/continental lithospheric package formed during early rifting stages of the Ligurian Tethys, in a tectonic setting analogous to present-day passive margins (e.g. the Galicia Bank), or embryonic oceans (e.g. the Red Sea). In these regions, subcontinental lithospheric mantle is exposed along with asthenospheric mantle during the initial stages of rifting. Nevertheless, the Internal and External Liguride bodies represent a rare example of an ophiolite associated with MORB geochemistry. They thus constitute samples of the different kinds of mantle that floored the Jurassic Tethyan ocean.

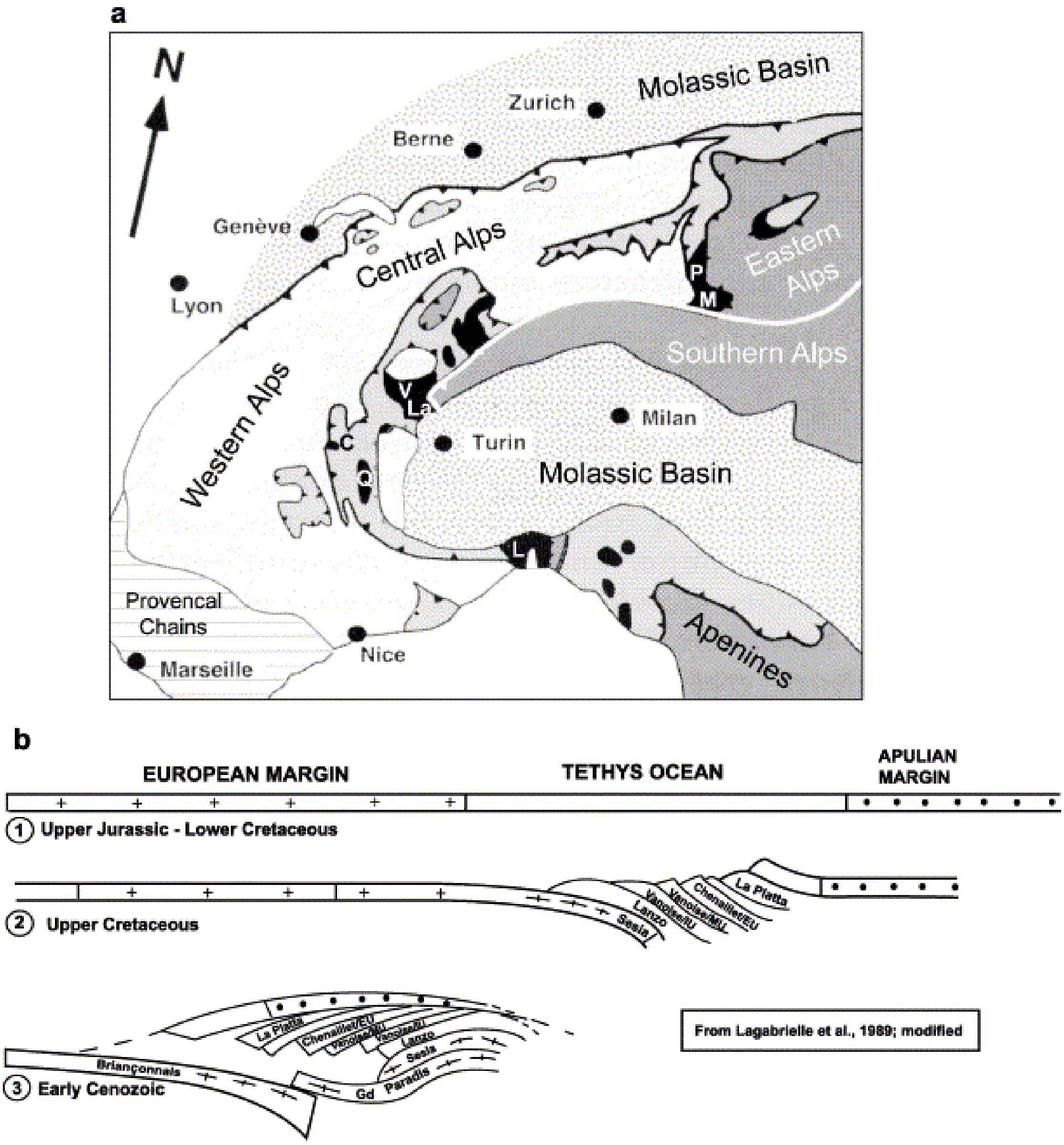


Fig. 1. (a) Schematic structural map of the Alps and the Northern Apennines (after [Lemoine et al]; modified) showing the location of the Tethyan ocean (classical Ligurian–Piemonte area) with the oceanic lithopshere *in black* and the overlaying calc-schists *in middle grey*, and its continental margins: European (Western and Central Alps) *in pale grey* and Apulian (Eastern and Southern Alps, and Northern Apennines) *in dark grey*. In this interpretation, the Sesia basement (to northeast of the Lanzo area) is assumed to belong to the Apulian margin. C: Chenaillet–Montgenèvre (France); Q: Queyras (France); V: Vanoise (France); La: Lanzo area (Italy); L: Ligurides (Italy); M: Val Malenco (Italy); P: La Platta (Eastern Switzerland). Our sampling areas Chenaillet–Montgenèvre (C), Queyras (Q), and Vanoise (V) belong, respectively, to External Units (EU; LP-LT metamorphism), Middle Units (MU for Queyras and western Vanoise) and Internal Units (IU for Eastern Vanoise ; HT-HP eclogite metamorphism) Units (see text for explanations). (b) hypothetic scheme from [Lagabrielle et al] and [Fudral, 1998] for the origin and the emplacement of the different units forming the Ligurian–Piemonte area in the Western Alps.

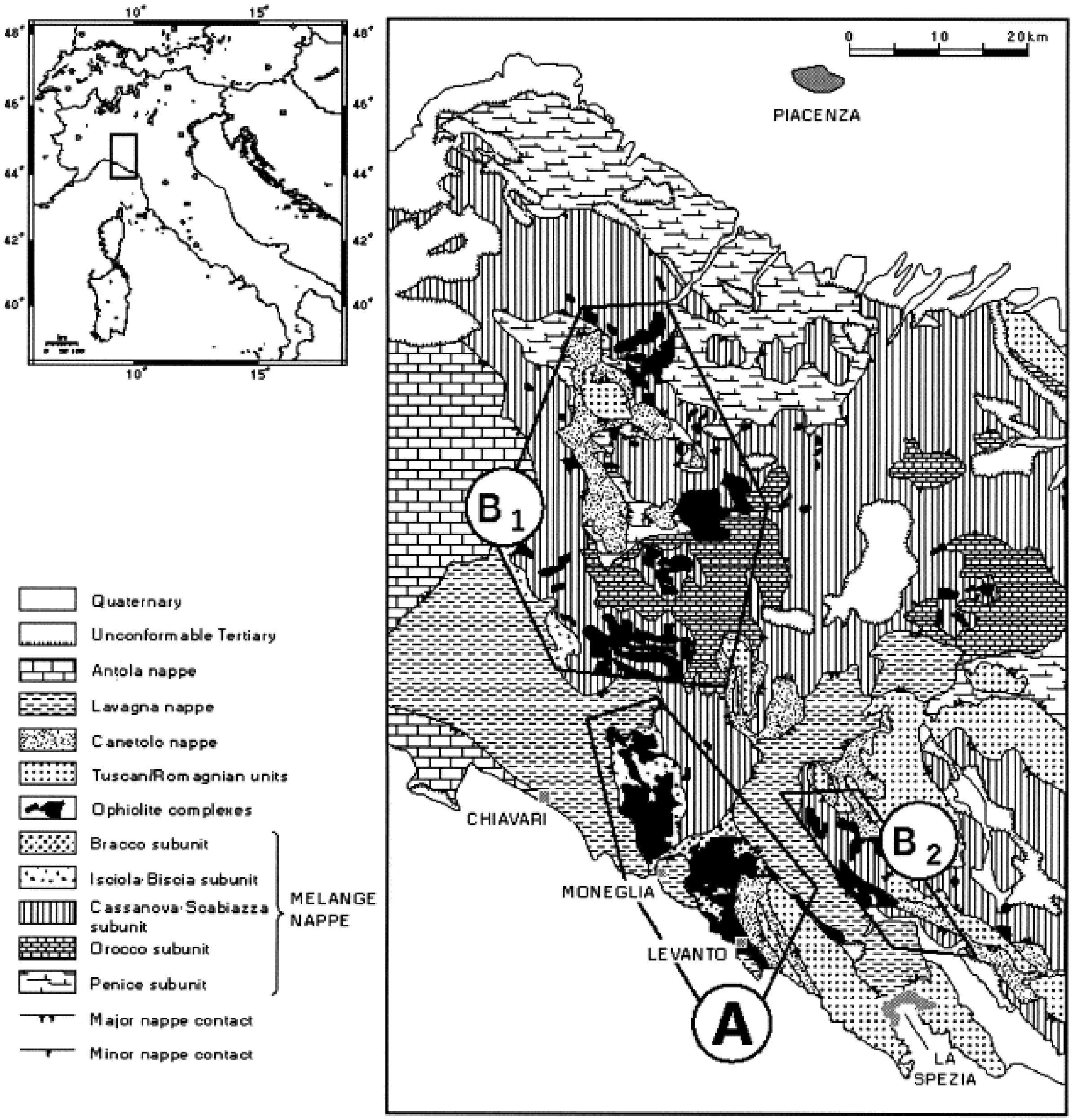


Fig. 2. Location map of the Ligurian peridotite bodies. Modified after Rampone et al. The Internal Liguride is labeled A, the External Liguride is labeled B_1 and B_2.

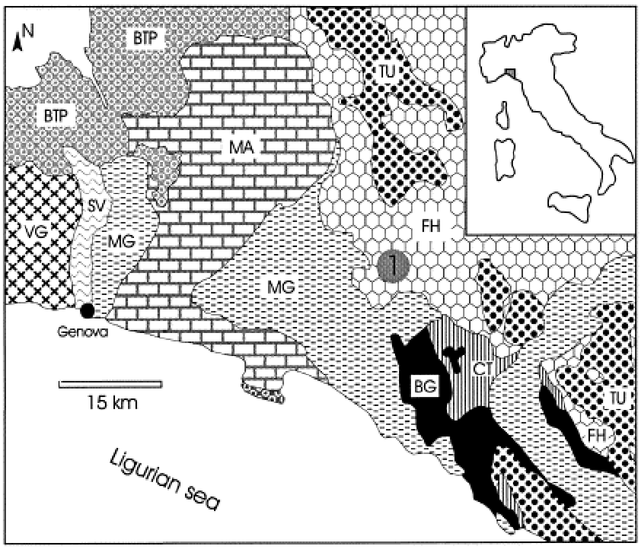


Fig. 3. Tectonic map of Eastern Liguria. *BTP*: Tertiary Piemontese Basin; *VG*: Voltri Group; *SV*: Sestri–Voltaggio Units; *MA*: Mt. Antola Unit; *MG*: Mt. Gottero Unit; *BG*: Bracco–Val Graveglia Unit (ophiolitic Internal Liguride Unit); *CT*: Colli/Tavarone Unit; *FH*: External Liguride Units; *TU*: Tuscan and Canetolo Units. The area labelled (*1*) within the External Liguride Units (FH) indicates the position of the Mt. Aiona peridotite body (sampling locality of the studied EL basaltic dikes). The investigated IL ophiolites are from the Bracco–Val Graveglia (BG) Unit.

Glej članek: *Smith, A. G., 2006: Tethyan ophiolite emplacement, Africa to Europe motions, and Atlantic spreading. Geological Society, London, Special Publications 2006; v. 260; p. 11-34*

Tethyan ophiolite emplacement

Final emplacement of the mid-Jurassic and mid-Cretaceous supra-subduction zone (SSZ) ophiolites onto adjacent continental areas in the Mediterranean region is synchronous with reductions in the rate of motion between Africa and stable Europe. The Apennine-Ligurian-Alpine ophiolites lack SSZ chemistry, are mid-ocean ridge basalt (MORB)-like, range in age from *c.* 169 to 148 Ma, and were emplaced in late Cretaceous and Cenozoic time. The Hellenic-Dinaric SSZ ophiolites include some MORB, ranging from 173 to 168 Ma, and were emplaced, eroded, and covered by younger sediments by *c.* 140 Ma. The creation of the Apennine-Ligurian-Alpine and Hellenic-Dinaric suites is attributed to the motion of Adria, which formed a promontory on Africa, or essentially moved with it, as the central Atlantic opened. Extension to the west of Adria gave rise to the Ligurian Sea, generating MORB crust: to the east, a pre-existing Triassic ocean was subducted, with rollback creating Jurassic SSZ ophiolites that were emplaced onto adjacent continental margins in the later stages of convergence. The two episodes of slower motions between Africa and Europe are attributed to two episodes of attempted subduction of a passive continental margin. This speculation suggests that emplacement of some SSZ ophiolites may exert a significant control on oceanic spreading patterns.

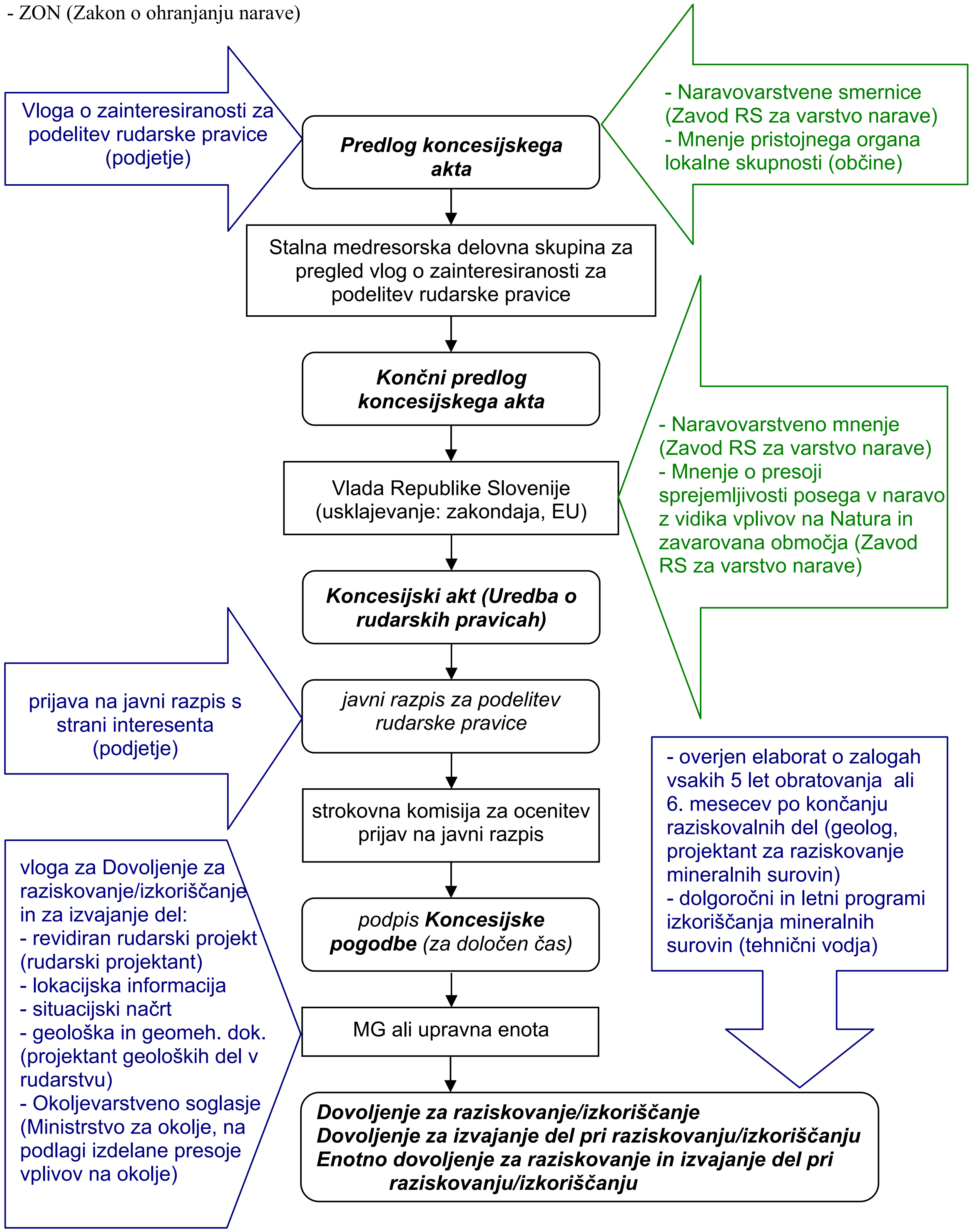
1) Ophiolite creation in the Mediterranean region appears to coincide with the initiation of new ridge segments in the Central and North Atlantic Ocean: the mid-Jurassic MORB and SSZ ophiolites were created shortly after the Central Atlantic began to open; crystallizatiun of the mid-Cretaceous SSZ ophiolites is close in time to the opening of the Labrador Sea.
2) The final stage of mid-Jurassic SSZ ophiolitc emplacement is reflected in a slowdown in the relative motions between Africa and stable Europe: the ophiolites were finally emplaced in latest Jurassic time, very close to the beginning of the first episode of slow relativc velocities; the mid-Cretaceous SSZ ophiolites were finally emplaced in latest Cretaceous time, close to the start of the second episode of slow relative velocities.
3) Both slow episodes are attributed to the attempted subduction of passive continental margins during the final stages of tectonic emplacement. Eventually the subduction zone, which was probably part of the plate margin between Africa and Europe at the time, became inactive and had to move elsewhere. Thus it appears that ophiolite emplacement influenced the spreading pattern in the Central and North Atlantic Ocean.
4) The emplacement of the MORB ophiolites in the Ligurian area is much later. Emplacement sampled ophiolites with a wide range of ages.
5) In the Mediterranean region, the interval between the slowdown or cessation of activity between Africa and stable Europe and the start of a new subduction zone between the two continents is about 15-20 Ma.

Kamnolom marmorja Carrara - osnovni podatki (pripravil Gorazd Žibret, GeoZS):

- svetovno znan marmor, kompleks obsega cca. 250 kamnolomov;
- visokometamorforiziran marmor jurske starosti;
- debelina produktivne plasti znaša več, kot 1000 metrov;
- pridobivanje se vrši površinsko in galerijsko (podzemno), največji kamnolom se imenuje fourier (?);
- proizvodnja blokov je zelo velika, >70.000 blokov/leto, kar je več, kot je bila proizvodnja marmorja v celotni bivši Jugoslaviji;
- bližnje pristanišče La Spezia je največje pristanišče za pretovarjanje naravnega kamna na svetu;
- 2 glavna tipa marmorja: bianco carrara ("cukrasto" bel) in statuario (za kiparstvo, ima maroge);
- težave: sekundarna obarvanja <40 μm piritnih zrnc, ki se ob stiku s kislimi vodami (kamen je bazičen) spremenijo v limonitna ali pa goethitna zrnca in marmor obarvajo;
- zaradi velikega nadkritja so v kamnu prisotne naravne napetosti, ki se ob eksploataciji sprostijo; zato morajo bloke marmorja starati, da se napetosti umirijo in šele nato grejo bloki v razrez; če se bloki ne starajo, se lahko plošče ukrivijo.

Shematski postopek pridobivanja dovoljenj od ideje do začetka izkoriščanja. Vhodno dokumentacijo na levi strani (modra barva) priskrbi interesent za izkoriščanje, dokumentacijo na desni (zelena barva) pa priskrbi ustrezni organ, ki vodi postopek.

<u>Zakoni, ki uravnavajo to področje:</u>
- ZUureP (Zakon o urejanju prostora)
- ZRud (Zakon o rudarstvu)
- ZGO (Zakon o graditvi objektov)
- ZVO (Zakon o varstvu okolja)
- ZON (Zakon o ohranjanju narave)



by Fausto Batini[1], Andrea Brogi[2], Antonio Lazzarotto[3], Domenico Liotta[4], and Enrico Pandeli[5,6]

Geological features of Larderello-Travale and Mt. Amiata geothermal areas (southern Tuscany, Italy)

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This paper summarises the geological features of the Larderello-Travale and Monte Amiata areas, where the world's most ancient exploited geothermal fields are located. In both geothermal areas, three regional tectonostratigraphic elements are distinguished, from the top: (a) Late Miocene-Pliocene and Quaternary, continental to marine sediments; (b) the Ligurian and Sub-Ligurian complexes, which include remnants of the Jurassic oceanic realm and of the transitional area to the Adriatic margin, respectively; (c) the Tuscan Unit (Tuscan Nappe), composed of sedimentary rocks ranging in age from Late Triassic to Early Miocene. The substratum of the Larderello and Monte Amiata areas is referred to as the Tuscan Metamorphic Complex. This is mainly known through drilling of geothermal wells. This complex is composed of two metamorphic units: the upper Monticiano-Roccastrada Unit and the lower Gneiss Complex. The Monticiano-Roccastrada Unit consists of (from top to bottom): the Verrucano Group, the Phyllite-Quartzite Group and the Micaschist Group. The Gneiss Complex consists only of pre-Alpine polymetamorphic gneiss. The Tuscan Metamorphic Complex is affected by contact metamorphism by Plio-Quaternary granitoids and their dyke swarms. Hydrothermal phenomena still occur in both geothermal fields. The Larderello-Travale and Mt. Amiata geothermal fields are located in the inner Northern Apennines, in an area that has been subject to extension since the ?Early-Middle Miocene. Two main extensional events are well expressed in the structures of the geothermal areas. The first extensional event (?Early-Middle Miocene) determined the tectonic delamination of the Ligurian Units and Tuscan Nappe. The second extensional event (Late Miocene–Present) is characterized by high-angle normal faults bounding the Neogene tectonic depressions of southern Tuscany.

Introduction

Continental extensional tectonic environments with high heat flow are often affected by geothermal systems, independently from the geodynamic context in which they are located (Barbier, 2002 and references therein). Extension also characterises southern Tuscany (inner Northern Apennines), where the most important geothermal fields of Italy are located (Figure 1).

The structural and stratigraphic setting of southern Tuscany derives from two different deformational processes: the first one is linked to the convergence between the European margin and the Adria microplate (Cretaceous-Early Miocene), producing the stacking of the Northern Apennines nappes; the second is related to the post-collisional extensional tectonics which have affected the inner zone of the Northern Apennines since the Early-Middle Miocene (Carmignani et al., 1994; Brunét et al., 2000 and references therein). This latter process is reflected by: (a) the present crustal and lithospheric thicknesses of about 22 km and 30 km respectively

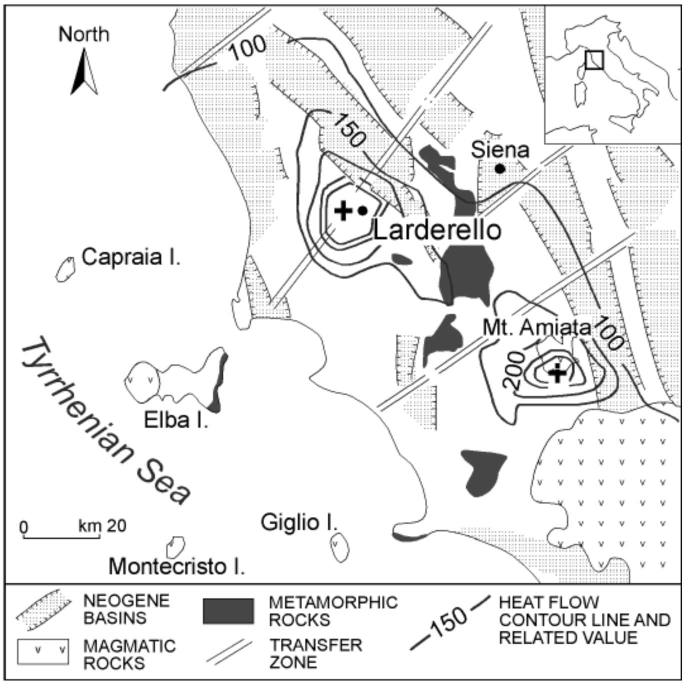

***Figure 1** Structural sketch map of southern Tuscany with the regional heat flow contour lines (equidistance: 50 mW/m^2). The Larderello-Travale and Mt. Amiata geothermal fields are located in areas where heat flow reaches 1 W/m^2 and 0.6 W/m^2, respectively. (after Baldi et al., 1995)*

Figure 2 The Larderello Valley ("Devil's Valley") in a 19th century print.

(Calcagnile and Panza, 1981); (b) the high heat flow (Baldi et al., 1995) that characterises southern Tuscany (120 mW/m^2 on average, with local peaks up to 1000 mW/m^2); (c) the anatectic to subcrustal magmatism that has affected southern Tuscany during the Late Miocene to Pleistocene time period (Serri et al., 1993). The Tuscan magmatism is coupled with Pliocene-Quaternary hydrothermal mineralization and widespread geothermal vents.

This paper summarises the geological features of the Larderello-Travale and Mt. Amiata geothermal areas, the most ancient exploited geothermal fields in the world. Particularly, the Larderello field has been industrially exploited since 1818 (Figure 2), when the Montecerboli Count, Francesco de Larderel, extracted boric acid from the geothermal vents. In 1904, the Larderello geothermal fluids were used to produce electricity by Prince Piero Ginori Conti. In contrast, the Mt. Amiata area has been exploited since the early 1960s, when the first electrical power plant was activated.

Today the endogenous fluids, intercepted at depth by boreholes, feed the Larderello-Travale and Mt. Amiata power plants belonging to the Enel GreenPower Electric Company. Present production is more than 700 MW, corresponding to about 2% of the total electricity production in Italy (Cappetti et al., 2000).

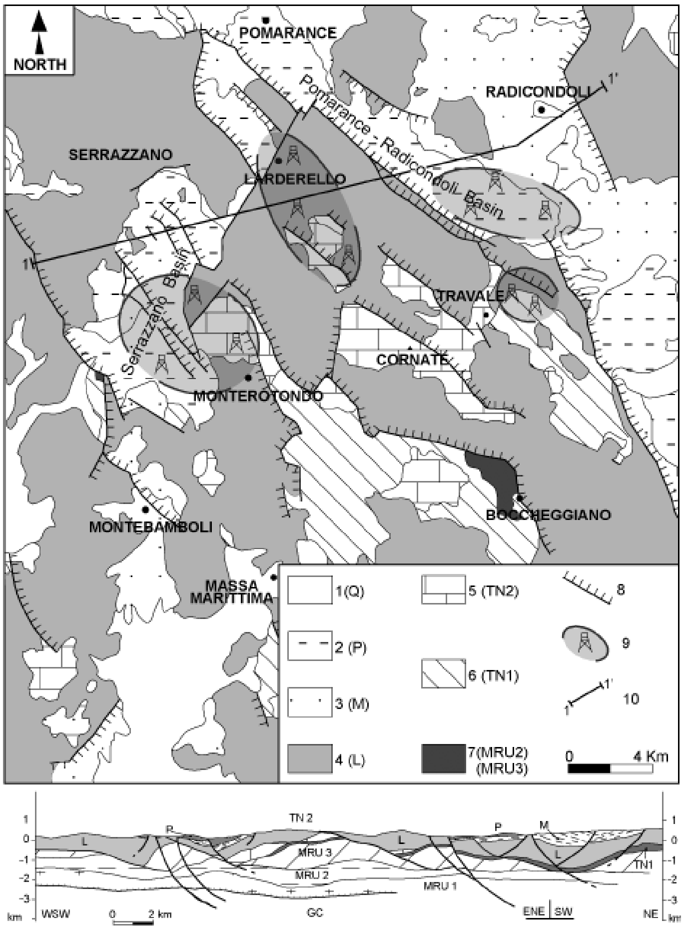

Figure 3 Geological sketch map of the Larderello-Travale area. Key: Neogene and Quaternary deposits: 1—Quaternary continental sediments; 2—Pliocene marine sediments; 3—Miocene continental and marine sediments; 4—Ligurian units l.s. (Jurassic- Eocene); 5—Tuscan Nappe: Late Triassic-Early Miocene sedimentary sequence; 6—Tuscan Nappe: Late Triassic basal evaporite (Burano Fm.); 7—Palaeozoic Phyllite-Quartzite Group (MRU$_2$) and Triassic Verrucano Group (MRU$_3$); 8—Normal faults; 9—Main geothermal fields; 10—Trace of geological cross-section. (MRU$_1$)—Palaeozoic Micaschist Group; (GC)–Gneiss Complex.

Geological features of the Larderello-Travale and Monte Amiata areas

The geological evolution of the Northern Apennines is well expressed in the structure of the Larderello-Travale and Monte Amiata geothermal fields.

Larderello-Travale Area

In the Larderello-Travale area three regional tectonostratigraphic elements crop out (Figure 3). These are, from top to bottom:

(1) Neogene and Quaternary deposits: Late Miocene to Pliocene and Quaternary, continental to marine sediments, filling up the extensional tectonic depressions which, in the geothermal areas, unconformably overlie the pre-Neogene substratum (Figure 4).

(2) The Ligurian Complex l.s.. This includes the Ligurian units s.s. and the Sub-Ligurian Unit. The Ligurian units are composed of remnants of the Jurassic oceanic basement and its pelagic sedimentary cover. The Sub-Ligurian Unit ("Argille e calcari" Unit) belongs to a palaeogeographical domain interposed between the Ligurian Domain and the Tuscan Domain (Figure 5). The Ligurian

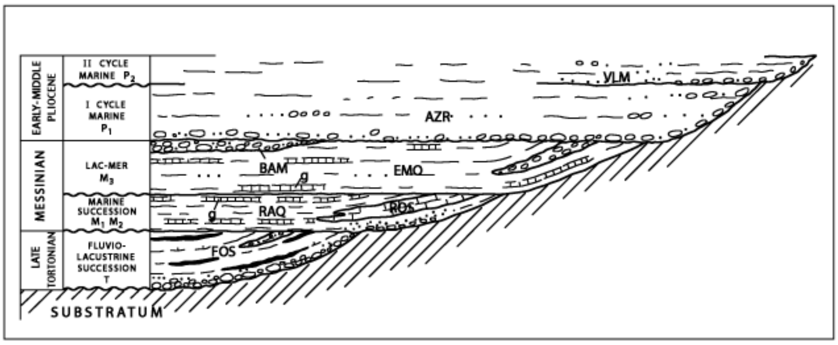

Figure 4 Stratigraphic relationships among the Neogene formations cropping out in the surroundings of the Larderello-Travale geothermal area; RAQ: Raquese Stream Fm. (Early Messinian), ROS: Rosignano Limestone (Early Messinian), EMO: Clays and gypsum of Era Morta River (Late Turolian), BAM: Montebamboli Conglomerate (Late Turolian); AZR: Blue Clays (Late-Middle Pliocene); VLM: Villamagna Fm. (Middle Pliocene).

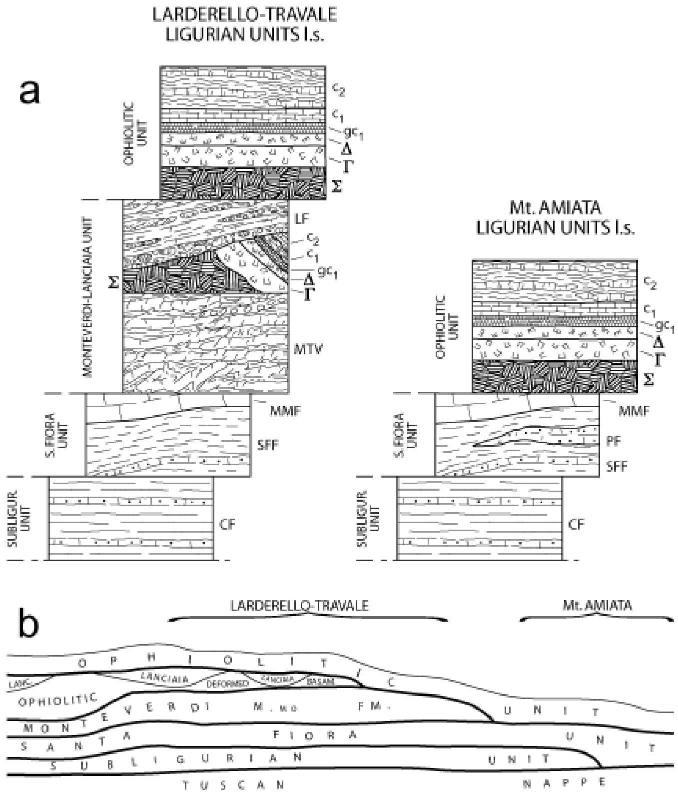

Figure 5 ***a) Structural and stratigraphic relationships between the Ligurian and Subligurian Units in the Larderello-Travale and Mt. Amiata geothermal areas. Ophiolitic Unit: Δ, Γ, ∑: Ophiolites (serpentinites, gabbros, basalts—Middle-Late Jurassic), gc_1: Mt. Alpe cherts (Late Jurassic), c_1: Calpionella limestone (Early Cretaceous), c_2: Palombini shales (Early Cretaceous); Monteverdi-Lanciaia Unit: MTV: Monteverdi Marittimo Fm. (Cretaceous-Early Palaeocene), LF: Lanciaia Fm. (Early-Middle Eocene); S. Fiora Unit: SFF: Santa Fiora Fm. (Late Cretaceous), Pf: Pietraforte Fm. (Late Cretaceous), MMF: Monte Morello Fm. (Paleocene-Eocene); Subligurian Unit: CF: Canetolo Fm. (Paleocene-Eocene).***

b) Reconstructed relationships among the Ligurian, Subligurian Units and Tuscan Nappe at the end of the collisional stage (Late Oligocene-Early Miocene).

l.s. Complex was thrust eastwards over the Tuscan Domain during latest Oligocene to Early Miocene times.

(3) The Tuscan Unit (Tuscan Nappe). This is related to part of the Late Triassic-Early Miocene sedimentary cover of the Adria continental palaeomargin (Figure 6). The Tuscan Nappe was detached from its substratum along the Triassic evaporite level and was thrust over the outer palaeogeographical domains during the Late Oligocene-Early Miocene compression.

The substratum of the Larderello-Travale area is referred to as the Tuscan Metamorphic Complex. This is mainly known through drillings of the geothermal fields, some of these penetrating down to about 4.5 km. This Complex is composed of two metamorphic units (Bertini et al., 1994): the upper Monticiano-Roccastrada Unit and the lower Gneiss Complex.

The Monticiano-Roccastrada Unit consists of three groups (Figure 7):

- The Verrucano Group. This is made up of Carnian phyllites and metacarbonates, related to marine littoral facies, and Middle–Early Triassic continental quartzites and quartz conglomerates. The Verrucano Group is imbricated in duplex structures, often separated by Late Triassic evaporites and Early-Late Palaeozoic phyllites and quartzites.

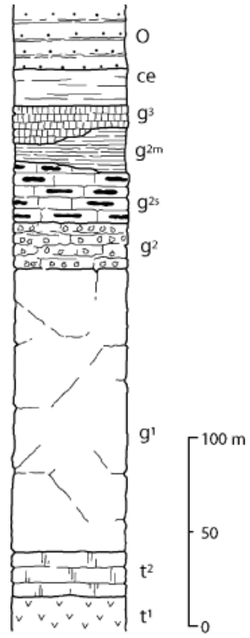

Figure 6 Stratigraphic succession of the Tuscan Nappe.
Symbols: O: Macigno Fm. (Late Oligocene-Early Miocene); ce: Scaglia toscana Fm. (Cretaceous-Oligocene); c^1: Maiolica Fm. (Early Cretaceous); g^3: Diaspri Fm. (Malm); g^{2m}: Marne a Posidonia (Dogger); g^{2s}: Calcare Selcifero Fm. (Middle-Late Liassic); g^2: Calcare Rosso ammonitico (Early-Middle Liassic); g^1: Calcare Massiccio (Early Liassic); t^2: Calcare a Rhaetavicula contorta (Rhaetic); t^1: Burano Fm. and Calcare cavernoso (Noric-Rhaetic).

- The Phyllite-Quartzite Group. This mainly consists of Palaeozoic phyllite and quartzite, affected by the Alpine greenschist metamorphism which overprints a previous Hercynian blastesis. Layers of anhydritic dolomites and basic metavolcanites in lenses can occur.
- Micaschist Group. This includes Palaeozoic rocks (garnet-bearing micaschists and quartzites with amphibolite zones) affected by Alpine and Hercynian deformations. Particularly, the micaschists were affected by a synkinematic Hercynian metamorphism and by an Early Permian thermal event (Del Moro et al., 1982; Pandeli et al., 1994 and references therein).

The Gneiss Complex consists of pre-Alpine polymetamorphic gneiss and paragneiss with intercalations of amphibolites and orthogneiss. In contrast to the Monticiano-Roccastrada Unit, the effects of the Alpine orogeny are not recorded in the Gneiss Complex (Elter and Pandeli, 1990). At different depths, deep boreholes encountered granitoids and felsic dykes (3.8–2.25 Ma, Villa & Puxeddu, 1994; Gianelli and Laurenzi, 2001) whose emplacement

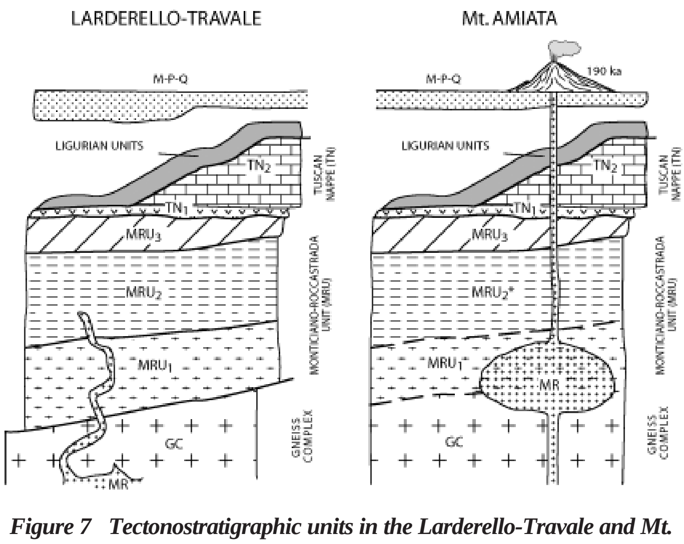

Figure 7 Tectonostratigraphic units in the Larderello-Travale and Mt. Amiata areas. Q-P-M: Quaternary, Pliocene and Miocene sediments; MR—Magmatic rocks; Tuscan Nappe (TN): TN_2—Early Miocene-Rhaetian sequence; TN_1—Late Triassic evaporite (Burano Fm.); Monticiano-Roccastrada Unit (MRU): MRU_3—Triassic Verrucano Group; MRU_2—Palaeozoic Phyllite-Quartzite Group; MRU_1—Palaeozoic Micaschist Group; GC: Palaeozoic Gneiss Complex.

gave rise to contact aureoles in the metamorphic host rocks (Elter and Pandeli, 1990; Musumeci et al., 2002 and references therein). Moreover, hydrothermal mineral associations (Gianelli, 1994), locally no older than 270,000 years and no younger than 10,000 years (Bertini et al., 1996), partially or totally fill the fractures affecting the Larderello metamorphic rocks.

Mt. Amiata Area

The geological framework of Mt. Amiata (Figure 8) is characterised by the trachitic-latitic Mt. Amiata volcano (0.3–0.2 Ma: Ferrari et al., 1996 and references therein). The outcropping units belong to the already mentioned Ligurian and Sub-Ligurian units (Figure 5) and to the Tuscan Nappe (Figure 6). The Monticiano-Roccastrada Unit does not crop out in the Mt. Amiata area, but it has been encountered by geothermal wells (Figure 9). This Unit is made up of very low-grade metamorphic sequences (Elter & Pandeli, 1991 with references therein) with: (a) Triassic Verrucano Group (MRU3 in Figure 7); (b) graphitic phyllite and metasandstone of probable Carboniferous age (Formation a); (c) ?Devonian hematite-rich and anhydrite-bearing chlorite phyllite, metasandstone with dolostone levels (Formation b); (d) Late Permian fusulinid-bearing crystalline limestone and dolostone with intercalations of graphitic phyllite (Formation c) (MRU2* in Figure 7).

Relicts of micaschists and gneisses have been discovered as xenoliths in the Mt. Amiata lavas (Van Bergen, 1983) (MRU1-GC in Figure 7), suggesting their occurrence at depth. Also the metamorphic rocks of the Mt. Amiata geothermal area are affected by the thermometamorphism and hydrothermalism linked to the recent magmatism (Gianelli et al., 1988).

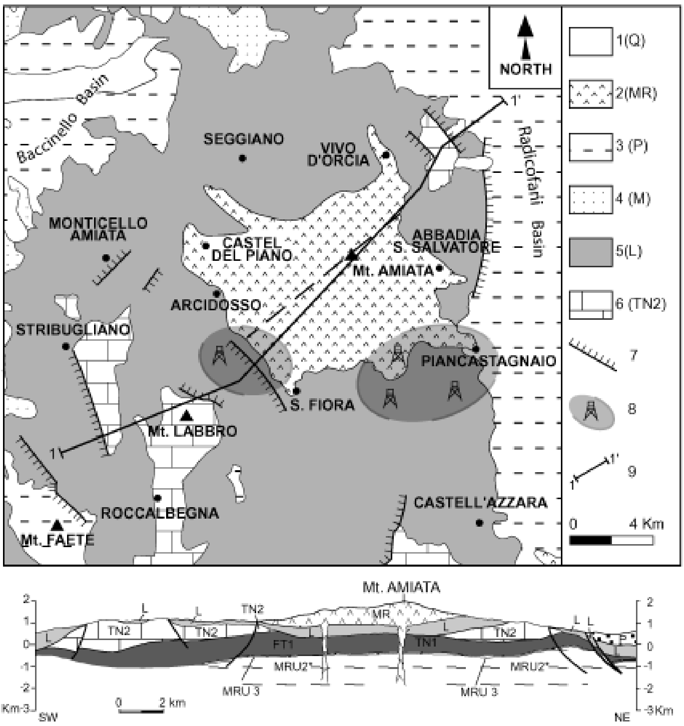

Figure 8 Geological sketch map of the Mt. Amiata area. Keys: 1—Quaternary continental sediments; 2—Magmatic rocks; 3—Pliocene marine sediments; 4—Miocene continental, brackish and marine sediments; 5—Ligurian Units l.s. (Jurassic-Eocene); 6—Tuscan Nappe (Late Trias-Early Miocene); 7—normal faults; 8—Main geothermal fields; 9—Trace of the geological cross-section; (TN_1)–Tuscan Nappe: Late Triassic basal evaporite (Burano Fm.); (MRU_3)—Triassic Verrucano Group; (MRU_2)—Palaeozoic phyllite Group (stratigraphic details are shown in Figure 9).*

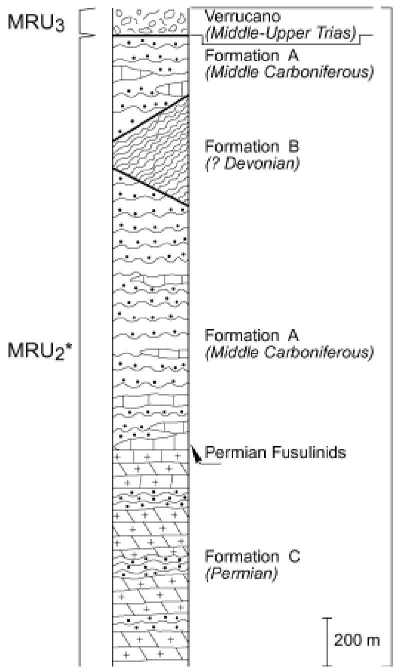

Figure 9 Relationship among the Triassic and Palaeozoic formations belonging to Monticiano-Roccastrada Unit encountered by geothermal wells in the Mt. Amiata area (after Elter & Pandeli, 1991).

Extensional structural features

Two different extensional events affected southern Tuscany after the emplacement of the Northern Apennines units. These are well expressed in the structures of the Larderello-Travale and Mt. Amiata geothermal areas (Figures 3 and 8). The first extensional event produced low-angle normal faults which soled out in the Late Triassic evaporites or in the Palaeozoic phyllites. According to some authors (Baldi et al., 1994; Carmignani et al., 1994) this first extensional event is related to ?Early-Middle Miocene on the basis of both stratigraphic considerations and mineral cooling ages linked to the exhumation of the Alpi Apuane core complex (Kligfield et al., 1986). The second extensional event (Late Miocene-Present) is characterized by high-angle normal faults which dissected all the previous structures and defined the Neogene tectonic depressions.

Reflection seismic features

Information on deeper structures derives from seismic reflection surveys carried out by Enel S.p.a. for geothermal exploration in the Larderello-Travale and Mt. Amiata areas. The seismic sections show a poorly reflective upper and a highly reflective mid–lower crust, particularly in the Larderello-Travale area (Cameli et al., 1993; Brogi et al., 2003). The top of the reflective crust is marked by a rather continuous reflector of high amplitude and frequency called the K-horizon (Batini et al., 1978), which locally exhibits bright spot features (Batini et al., 1985). The K-horizon ranges in depth from 3 to 8 km (Cameli et al., 1998 and references therein) both in the Larderello-Travale and Mt. Amiata fields (Figures 10 and 11). Present–Pliocene normal faults tend to flatten at the K-horizon depth or just below it (Cameli et al., 1993). The origin of the reflectivity at the K-horizon and in the zone below has been discussed by several authors (see Gianelli et al., 1997 for a review). The occurrence of fluids can explain the observed high contrast in acoustic impedance (Liotta and Ranalli, 1999 and references therein). Gianelli et al. (1997) hypothesised that a granite carapace and associated wall rocks, probably delimited by overpressurised horizons, could give rise to the K-reflector. Furthermore, temperature data, hypocentral distributions and rheological predictions led to the explanation of the K-horizon as the top of an active shear zone, located at the brittle/ductile transition (Cameli et al., 1993; 1998; Liotta & Ranalli, 1999). In this framework,

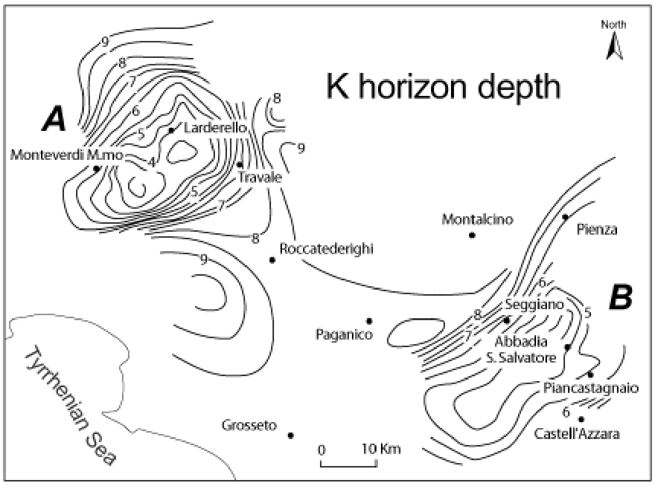

Figure 10 Contour lines in kilometres (equidistance: 0.5 km) of the K-horizon depth. A and B show respectively the Larderello-Travale and Mt. Amiata geothermal areas (after Cameli et al., 1998).

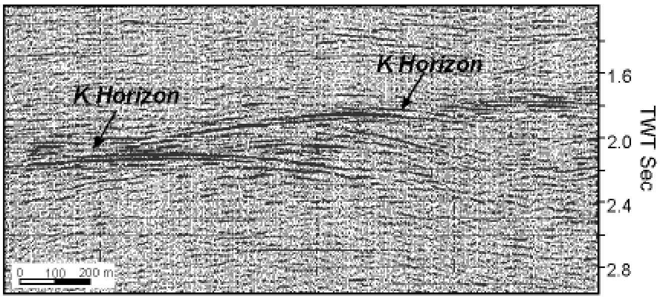

Figure 11 The K-horizon in the Larderello-Travale geothermal area. Vertical axis: TWT seconds.

the K-horizon would not represent a lithological boundary, even if it locally coincides with the roof of a magmatic body.

Geothermal reservoirs and fluids

Both in the Larderello-Travale and Mt. Amiata fields, there are two geothermal reservoirs recognised and industrially exploited (Cappetti et al., 2000; Bertini et al., 1995): one more superficial, located in the cataclastic horizon corresponding to the Late Triassic evaporites and the overlying Jurassic carbonatic formations; and a deeper one, located in fractured metamorphic rocks at depths ranging between 2000 and 4500 m. The Cretaceous–Early Miocene terrigenous formations of the Tuscan Nappe, the Ligurian Units l.s. and the Miocene-Pliocene sediments represent the impervious cover of the more superficial geothermal reservoir.

The geothermal fluids are mainly made up of a mixture of meteoric water with thermometamorphic and magmatic fluids (Minissale, 1991; Manzella et al., 1998).

The Larderello-Travale geothermal field produces high enthalpy geothermal fluids (T = 150˚–260˚C; P = 2–15 bar) which mainly comprise superheated steam and minor gases (max 15% by weight) essentially made up of CO_2 and H_2S. The average flow rate of the wells is 25 t/h of dry steam (max 350 t/h). In the deeper reservoir, pressure and temperature increase with depth, up to values of 70 bar and 350˚C.

The Monte Amiata geothermal area has two water-dominated fields (Bagnore and Piancastagnaio fields). In the deeper reservoir, P = 200–250 bars and T = 300˚–360˚C. The resulting fluids are two-phase mixtures with T = 130˚–190˚C and P = 20 bars. Fluids are characterised by a TDS content of about 10–12 g/l (mainly alkaline chlorides and, to a lesser extent, alkaline earth bicarbonates) and a gas percentage similar to that of the Larderello field.

Concluding remarks

Field information integrated with borehole and seismic data allow the reconstruction of the structural and stratigraphic features of the Larderello-Travale and Mt. Amiata geothermal areas. We emphasise two main points:

(a) These geothermal areas are located in a regional extensional context whose development favoured the localization of fractured zones, magmatism and high heat flow.

(b) Deep fractured zones in the metamorphic rocks and cataclasites in the Triassic evaporite levels represent the reservoirs in both described geothermal areas. In principle, in geothermal areas the permeability is time-dependent, since the circulation of geothermal fluids favours the deposition of hydrothermal minerals. However, fractures are mantained open only if microseismicity occurs, as is the case in both geothermal areas (Cameli et al., 1993; 1998).

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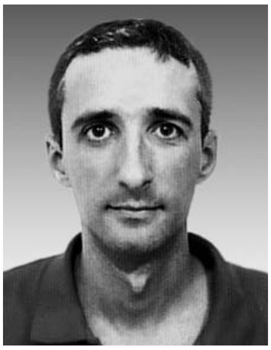

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Plio-Quaternary magmatism in Italy

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Plio-Quaternary magmatism in Italy exhibits an extremely variable composition, which spans almost entirely the spectrum of magmatic rocks occurring worldwide. Petrological and geochemical data provide a basis for distinguishing various magmatic provinces, which show different major element and/or trace element and/or isotopic compositions. The Tuscany province (14–0.2 Ma) consists of silicic magmas generated through crustal anatexis, and of mantle-derived calcalkaline to ultrapotassic mafic rocks. The Roman, Umbria, Ernici-Roccamonfina and Neapolitan provinces (0.8 Ma to present) are formed by mantle-derived potassic to ultrapotassic rocks having variable trace element and isotopic compositions. The Aeolian arc (?1 Ma to present) mainly consists of calcalkaline to shoshonitic rocks. The Sicily province contains young to active centers (notably Etna) with a tholeiitic to Na-alkaline affinity. Finally, volcanoes of variable composition occur in Sardinia and, as seamounts, on the Tyrrhenian Sea floor. Magmas in the Aeolian arc and along the Italian peninsula have a subduction-related geochemical character, whereas the Sicily and Sardinia provinces display intraplate signatures. Intraplate and orogenic volcanics coexist on the Tyrrhenian Sea floor.

The geochemical and isotopic complexities of Plio-Quaternary magmatism reveal that the upper mantle beneath Italy consists of various domains, spanning both orogenic and anorogenic compositions. Isotopic data suggest that compositional heterogeneity originated from mixing between various mantle reservoirs, and between these and subduction-related crustal material. This probably occurred during the Cenozoic-Quaternary geodynamic evolution of the western Mediterranean.

Introduction

The Italian peninsula is one of the most complex geodynamic settings on Earth (e.g. Wezel, 1985; Doglioni et al., 1999 and references therein). One expression of this complexity is the wide variety of Plio-Quaternary volcanic rocks, which range from subalkaline (tholeiitic and calcalkaline) to Na- and K-alkaline and ultra-alkaline, from mafic to silicic, and from oversaturated to strongly undersaturated in silica. Trace element contents and isotopic signatures are also highly variable, covering both mantle and crustal values, and ranging from typical intra-plate to orogenic compositions. This extreme magmatic diversity requires the occurrence of a complexly zoned mantle, which reveals an unusual geotectonic setting for the Italian region. Understanding the origin and evolution of the mantle beneath Italy is a challenge for igneous petrology, geochemistry, and geodynamics.

This paper describes the most important geochemical and petrological characteristics of the Plio-Quaternary volcanism in Italy, with the aims of (i) clarifying the first-order processes of magma genesis and evolution and (ii) providing constraints for models of geodynamic evolution of the Italian peninsula and adjoining regions.

Petrological characteristics of Plio-Quaternary magmatism in Italy

The Plio-Quaternary magmatism in Italy occurs along a belt parallel to the Tyrrhenian Sea border, in Sicily and Sicily Channel, on the Tyrrhenian Sea floor, and in Sardinia (Figure 1). The erupted volcanic rocks exhibit a large compositional variability, which is best illustrated by the Total Alkali vs. Silica diagram (TAS) shown in Figure 2. It is evident that Recent magmatism in Italy ranges from ultrabasic to acid, and from sub-alkaline to ultra-alkaline, covering

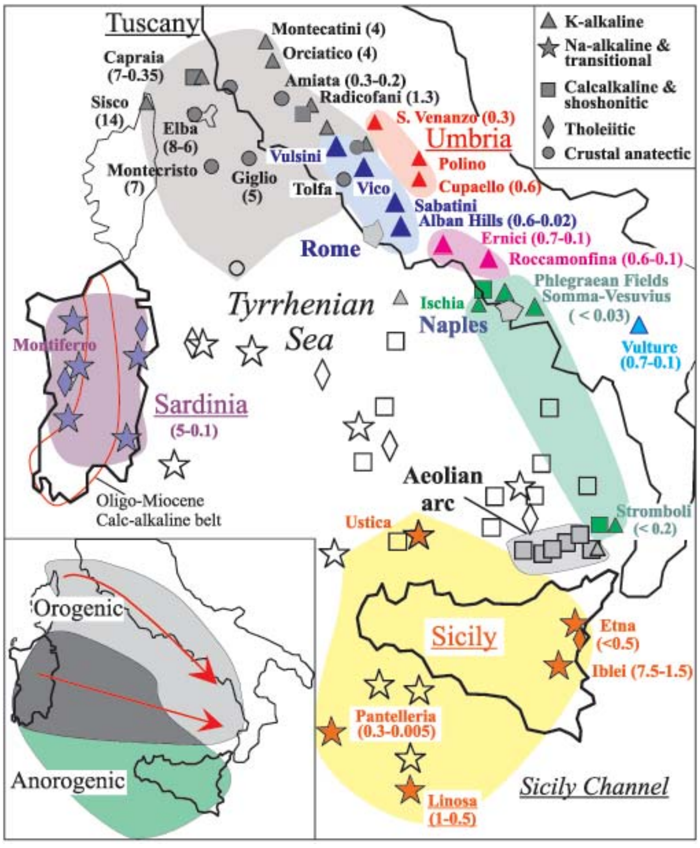


***Figure 1** Distribution of Recent magmatism in Italy. Open symbols indicate seamounts. Ages (in Ma) are given in parentheses. Different colours denote various magmatic provinces. Inset: schematic distribution of orogenic and anorogenic volcanism: red arrows indicate migration of orogenic magmatism with time.*

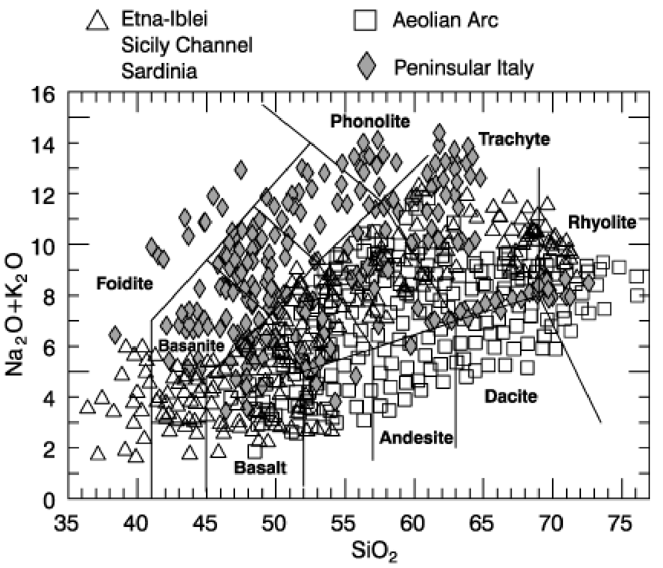


Figure 2 Total alkali vs. silica classification diagram for Italian Plio-Quaternary magmatic rocks. For source of data see Peccerillo (2002).

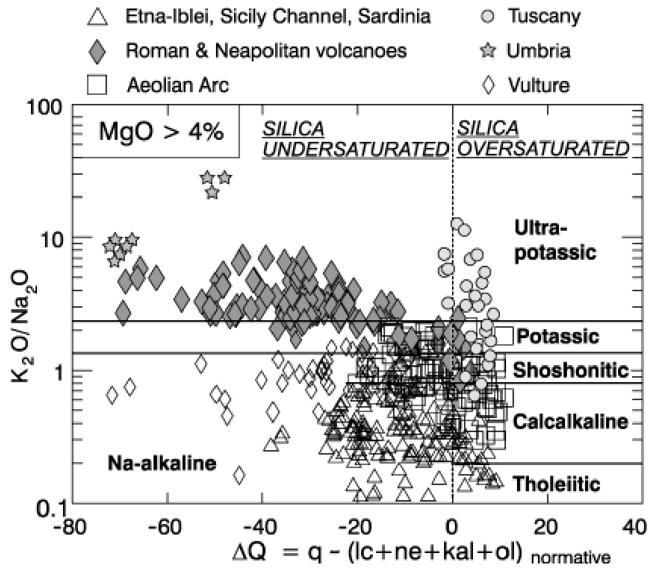


Figure 3 ΔQ vs. K_2O/Na_2O classification diagram for Plio-Quaternary mafic (MgO > 4%) volcanic rocks from Italy. ΔQ is the algebraic sum of normative quartz (q), minus leucite (lc), nepheline (ne), kalsilite (kal) and olivine (ol). Silica oversaturated rocks have $\Delta Q > 0$, whereas silica undersaturated rocks have $\Delta Q < 0$.

almost entirely the compositional field of igneous rocks occurring worldwide. Similarly, large variations are also observed for trace elements and isotopes, as discussed below.

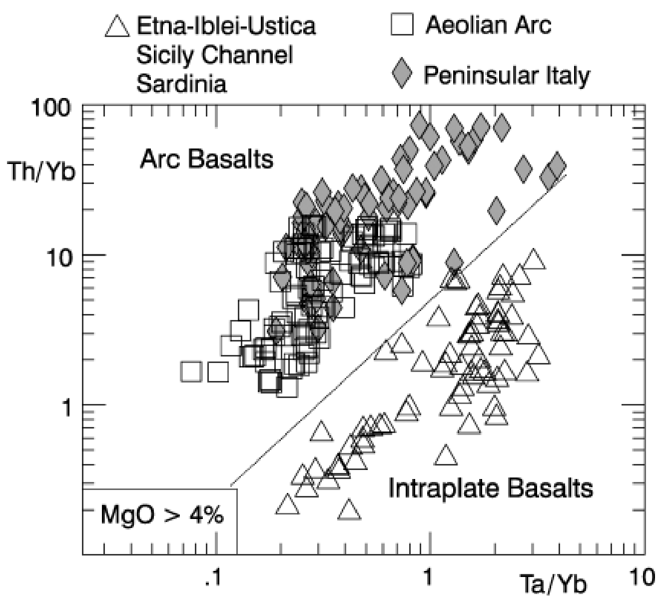
A large proportion of Italian Plio-Quaternary volcanic rocks have high-silica, low-MgO compositions. However, mafic rocks (MgO>3–4 wt%) deserve particular attention, since they are the closest relatives of primary mantle-derived magmas that were parental to erupted lavas, and can furnish the maximum of the information on mantle sources. Figure 3 is a classification diagram (Peccerillo, 2002), which shows that Italian mafic volcanics range from compositions that are strongly undersaturated to oversarturated in silica, from tholeiitic, calcalkaline, and shoshonitic to Na-alkaline, potassic, and ultrapotassic.

Regional distribution of magma types

There is a strong correlation between petrological characteristics of recent magmas and their regional distribution (Figure 1). Tholeiitic rocks occur in western Sicily (e.g. older Etna and Iblei), Sardinia, and on the Tyrrhenian sea floor (MORB and island arc tholeiites). Calcalkaline and shoshonitic rocks are concentrated in the Aeolian arc, although they are also found in the Naples area and in Tuscany (e.g. Capraia). Other calcalkaline and shoshonitic volcanoes occur as seamounts on the Tyrrhenian Sea floor, where they show an age decreasing south-eastward, from the Oligo-Miocene calcalkaline volcanic belt of Sardinia to the active Aeolian islands and seamounts (e.g. Beccaluva et al., 1989; Santacroce et al., 2003 and references therein). Na-alkaline and transitional rocks occur at Etna, Iblei, in the Sicily Channel (e.g. Pantelleria), in the Tyrrhenian Sea (Ustica and some seamounts) and extend to Sardinia (Lustrino et al., 2000). Potassic and ultrapotassic rocks represent the most typical compositions in central Italy. These occur over a large belt, from southern Tuscany to the Naples area (Vesuvius, Ischia, Phlegraean Fields); some potassic rocks occur at Vulcano and Stromboli in the Aeolian arc. Note, however, that potassic and ultrapotassic rocks from Tuscany differ from potassium-rich rocks from central-southern Italy on the basis of their silica saturation and K_2O/Na_2O ratios (Figure 3). Moreover, ultrapotassic volcanoes in Umbria are characterised by extremely high K_2O/Na_2O and very low degrees of silica undersaturation. Finally, undersaturated alkaline rocks, which are rich in both Na and K, with variable K_2O/Na_2O ratios, occur at Mount Vulture, east of Vesuvius.

Regional variation of trace element and Sr-Nd-Pb-Oxygen isotope compositions of mafic rocks

The mafic rocks from Italy have variable abundances and ratios of trace elements. Large Ion Lithophile Elements (LILE, e.g. K, Rb, Th) generally have high concentrations in calcalkaline, potassic, and ultrapotassic rocks. High Field Strength Elements (HFSE, e.g. Ta, Nb, Zr, Ti) have high concentration in Na-alkaline rocks, and low values in calcalkaline and potassic volcanics. Trace elements ratios (especially LILE/HFSE) are useful to distinguish intraplate and subduction-related basalts. The Th/Yb vs. Ta/Yb discriminant diagram of Wood et al., 1979 (Figure 4) is used here to show that mafic rocks from eastern Sicily, Sicily Channel, Ustica, and Sardinia fall in the



Figures 4 Th/Yb vs. Ta/Yb diagram for Plio-Quaternary mafic rocks from Italy, discriminating between intraplate and arc basalts.

field of intraplate (anorogenic) basalts, whereas the magmas occurring in the Aeolian arc and along the Italian peninsula have clear island-arc (i.e. orogenic) signatures. Subduction-related and intraplate volcanics coexist on the Tyrrhenian Sea floor (Figure 1, inset).

Additional petrogenetic information can be obtained by other trace element ratios and isotopes (Figures 5, 6). These highlight important variations that are heavily correlated to regional distribution, and are rather independent on the major petrological characteristics. For instance, calcalkaline and shoshonitic rocks from Tuscany fall in a distinct field with respect to rocks of equivalent petrologic composition from the Aeolian arc (Peccerillo, 1999, 2002).

The variation of $^{87}Sr/^{86}Sr$ vs. $^{143}Nd/^{144}Nd$ ratios of mafic rocks (Figure 6) show that the Italian volcanics define a curved trend between typical mantle compositions (MORB, Etna, Sicily channel, etc.) and upper crust values. Moreover, there is an overall increase of $^{87}Sr/^{86}Sr$ and a decrease of $^{143}Nd/^{144}Nd$ from south to the north, and the various regions display distinct isotopic compositions. Similar trends are shown by Pb isotope ratios (Conticelli et al., 2001 and references therein).

Oxygen isotopic data are also variable in the volcanic rocks from central-southern Italy. The lowest values are found in the south (e.g. $\delta^{18}O \approx +5.5$ to 6‰, in the mafic rocks from the Aeolian arc). Higher values ($\delta^{18}O \approx +7$ to +8‰) are found on mafic potassic and ultrapotassic rocks and separated minerals from central Italy (Harmon and Hoefs, 1995 and references therein).

Magmatic provinces in central-southern Italy: a new classification scheme

Plio-Quaternary magmatism of central-southern Italy has been classically subdivided into various magmatic provinces, represented by Tuscany, the Roman-Neapolitan area (the so-called Roman Comagmatic Province), the Aeolian arc, the Sicily and Sicily Channel (Etna, Iblei, Ustica, Pantelleria, Linosa), and Sardinia. Major, trace element and isotopic data reported above (Figures 3–6) provide evidence for a much more varied magmatic setting. These data permit subdivision of the Italian volcanism into several provinces that exhibit distinct major element compositions and/or incompatible trace element ratios and/or radiogenic isotope signatures (Peccerillo, 1999, 2002). These differences reveal distinct petrogenetic histories. The newly-established magmatic provinces are indicated in Figure 1. Their petrological characteristics and ages are summarised in Table 1.

Petrogenesis

Low-pressure magma evolution

As stated earlier, the largest proportion of Recent volcanism in Italy consists of high-silica lavas, such as andesites, rhyolites, trachytes and phonolites. Except for the Tuscany acid rocks, which are of crustal anatectic origin, these intermediate to silicic magmas were derived predominantly through fractional crystallisation from mafic parents. Mixing between various types of magmas and assimilation of crustal rocks also played an important role in magmatic compositional evolution for some volcanoes (Peccerillo, 2002, and references therein).

However, it is unlikely that such evolutionary processes, including contamination through magma-crust interaction, are responsible for the range of petrological, geochemical and isotopic variations observed in mafic volcanic rocks along the Italian peninsula. It is pertinent to recall that the high concentration of incompatible trace elements (e.g. Th, Sr, REE, etc.) of Italian rocks effectively buffers modifications of trace element and isotope ratios during magma evolution. This holds also true for mafic melts whose evolution degree is low to moderate (see discussion in Conticelli et al., 2001; Peccerillo, 1999, 2002). Therefore, the large geochemical and isotopic variations observed in Italy basically reflect compositional characteristics of mantle sources.

Genesis of mafic magmas

The variable petrological characteristics of Italian recent magmatism require a wide variety of mantle compositions and petrogenetic processes (i.e. degrees and pressure of partial melting, mantle mineral compositions, fluid pressure, etc.) to be generated (see Peccerillo, 2002). The potassic nature of most of the mafic Italian magmas require that a K-rich mineral, such as phlogopite, was present in the upper mantle and melted to produce the potassic magmas. The variable potassium contents probably reflect melting of different amounts of phlogopite. However, phlogopite is not a typical mantle mineral and its presence in the upper mantle reveals compositional anomalies. These can be generated at different spatial scales by introduction of K-rich fluids or melts: this process is known as mantle metasomatism. The large amount of potassic magma within the Italian peninsula requires very extensive mantle metasomatism (Peccerillo, 1999).

Isotopic data furnish further insight into mantle metasomatic processes. The curved trend of Sr-Nd isotope ratios (Figure 6) clearly suggests that the magmatism in central-southern Italy results

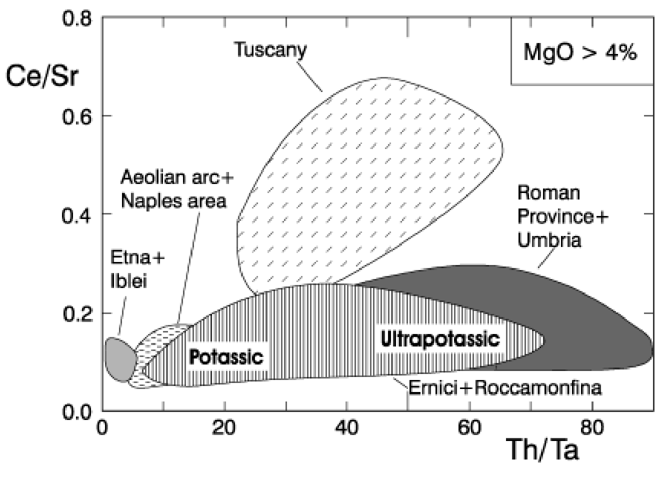


Figure 5 Variation of trace element ratios in Plio-Quaternary Italian mafic rocks. Note strong regional variation.

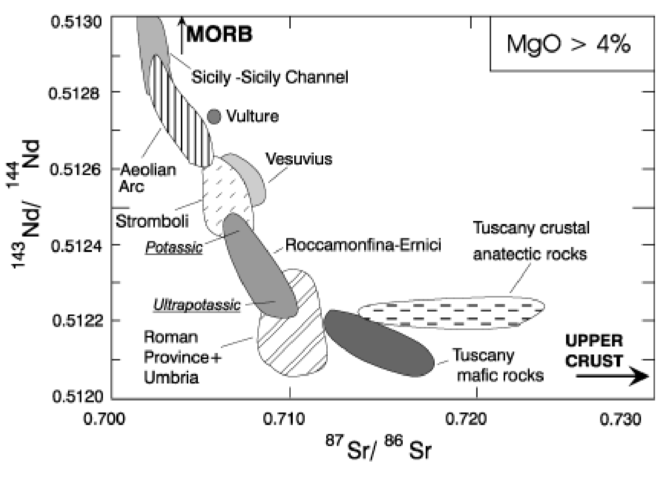


Figure 6 Sr vs. Nd isotope diagram for Plio-Quaternary mafic volcanic rocks from Italy. Note strong regional variation.

Table 1 Petrological characteristics and ages of Plio-Quaternary volcanic provinces in Italy.

MAGMATIC PROVINCE (age in Ma)	MAIN MAGMATIC CENTERS AND AGES (in Ma)	MAIN ROCK TYPES AND VOLCANIC STRUCTURES
TUSCANY (14-0.2)	***Acid intrusions***: Elba (8-6), Montecristo (7), Giglio (5), Campiglia-Gavorrano (5-4). ***Acid volcanics***: San Vincenzo (4.5), Roccastrada (2.5), Amiata (0.3-0.2), Cimini (1.4-1.1), Tolfa (3.8-1.8). ***Mafic centers***: Sisco (14), Capraia (7-3.5), Orciatico and Montecatini val di Cecina (4), Cimini (0.9), Radicofani (1.3), Torre Alfina (0.8)	***Crustal anatectic rocks***: Granitoid intrusions, aplites, pegmatites. Monogenic lava flows and domes, and stratovolcanoes (Mt. Amiata, Cimini Mts.). ***Mafic rocks***: monogenic extrusive and subvolcanic bodies with potassic and ultrapotassic (*lamproites*) composition; calcalkaline and shoshonitic rocks at Capraia.
UMBRIA (0.6-0.3)	San Venanzo (0.3), Cupaello (0.6-0.5), Polino (0.3)	Monogenic pyroclastic centers and lava flows with an ultrapotassic melilititic (*kamafugites*) composition.
ROMAN PROVINCE (0.6-0.02)	Vulsini (0.6-0.15), Vico (0.4-0.1), Sabatini (0.6-0.04), Alban Hills (0.6-0.02)	Large volcanoes formed by potassic (trachybasalt, latite, trachyte) and ultrapotassic (leucite-tephrite, leucitite, phonolite) lavas and pyroclastics.
MONTI ERNICI – ROCCAMONFINA (0.7-0.1)	Ernici: Pofi, Ceccano, Patrica, etc. (0.7-0.1) Roccamonfina (0.6-0.1)	Monogenic cinder cones and lava flows (Ernici), and a stratovolcano with caldera (Roccamonfina) formed by ultrapotassic (leucite-tephrite to phonolite) and potassic (trachybasalt to trachyte) rocks.
CAMPANIA – STROMBOLI (0.8 – Present)	Somma-Vesuvius (0.03-1944 AD), Phlegraean Fields (0.05-1538 AD), Ischia (0.13-1302 AD), Procida (0.05-0.01), Ventotene (0.8-0.1), Stromboli (0.2 – Present)	Stratovolcanoes with calderas formed by calcalkaline, shoshonitic, potassic (trachybasalts to trachytes) and ultrapotassic (leucite-tephrite to phonolites) rocks.
VULTURE (0.7 - 0.1)	Vulture, Melfi	Stratovolcano with caldera formed by Na-K-rich tephrites, phonolites, foidites with abundant hauyne. Carbonatite(?)
AEOLIAN ARC (1(?) – Present)	Panarea (0.15-0.05), Vulcano (0.12-1888 AD), Lipari (0.2-580 AD), Salina (0.5-0.13), Filicudi (1(?)-0.04), Alicudi (0.06-0.03)	Stratovolcanoes with dominant calcalkaline (basalt-andesite-rhyolite) and shoshonitic compositions.
SICILY (7.5 – Present)	Etna (0.5-Present), Iblei (7.5-1.5), Ustica (0.7-0.1), Pantelleria (0.3-0.005), Linosa (1-0.5)	Tholeiitic basalts to Na-alkaline rocks (basanite, hawaiite, trachyte, peralkaline trachyte and rhyolite) forming stratovolcanoes, diatreme, small plateau, etc.
SARDINIA (5.3 - 0.1)	Capo Ferrato (5), Montiferro (4-2), Orosei-Dorgali (4-2), Monte Arci (~ 3), Logudoro (3-0.1)	Tholeiitic basalts to Na-alkaline rocks (basanite, hawaiite, trachyte, alkaline trachyte and rhyolite) forming stratovolcanoes, basaltic plateau and monogenic centres.
TYRRHENIAN SEA FLOOR (7 – Present)	Magnaghi (3), Marsili (1.7-0), Vavilov, Anchise, Lametini, Palinuro, Pontine Islands (?) (~4-1), etc.	Coexisting intraplate (oceanic tholeiites, Na-transitional and alkaline) and arc (arc-tholeiitic, calcalkaline and shoshonitic) rocks.

from mixing between mantle and crustal end-member, revealing input of crustal material into the mantle (mantle contamination). The increase in crustal signatures from Sicily to Tuscany (increase of $^{87}Sr/^{86}Sr$ and decrease of $^{143}Nd/^{144}Nd$) reveals an enhancement in the amount of crustal contaminant going northward. The mantle-like isotopic signatures of Sicily and Sardinia magmatism indicate that the sources of these magmas were not subjected to significant compositional modification by input of crustal material, and probably represent largely pristine and uncontaminated mantle reservoirs.

Geodynamic significance

Much of the discussion on the geodynamic significance of the Recent Italian magmatism has addressed the problem of whether it relates to subduction processes or it represents an intraplate magmatism (e.g. Ayuso et al., 1997). The hypothesis that the variable and anomalous composition of volcanism in the Italian peninsula reflects addition of crustal material to the upper mantle, inevitably leads to the conclusion that at least the magmatism occurring from the Aeolian arc to Tuscany is indeed related to subduction processes. By contrast, the volcanoes in the Sicily and Sardinia provinces and some Tyrrhenian seamounts are intraplate and reflect derivation from mantle source unmodified by subduction. Therefore, the answer to the old question of whether Italian magmatism is subduction-related or not, is simply answered by saying that some volcanoes are subduction-related, whereas other volcanoes are not (Figure 1, inset).

This concept is well explained by a $^{87}Sr/^{86}Sr$ vs. $^{206}Pb/^{204}Pb$ diagram (Figure 7). This shows that the Italian volcanics define two main trends, both emanating from a high $^{206}Pb/^{204}Pb$ and low $^{87}Sr/^{86}Sr$ mantle composition: these mantle reservoirs are called "HIMU" (high-μ, where μ = Th/Pb ratio) and FOZO (Focal Zone) by isotope geochemists (e.g. Zindler and Hart, 1986). One trend includes the Aeolian arc and peninsular Italy, and points to moderately low $^{206}Pb/^{204}Pb$ and high $^{87}Sr/^{86}Sr$ compositions, which are typical of the upper crust. A second trend includes Etna-Iblei, Sardinia and some Tyrrhenian seamounts, and points to a mantle reservoir characterised by low $^{206}Pb/^{204}Pb$ and $^{87}Sr/^{86}Sr$: this is called EM1 (Enriched Mantle 1). The first trend is suggestive of mantle

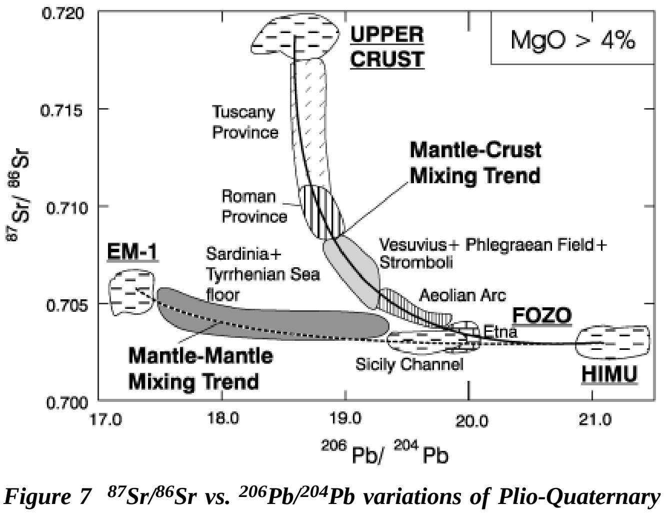


***Figure 7** $^{87}Sr/^{86}Sr$ vs. $^{206}Pb/^{204}Pb$ variations of Plio-Quaternary Italian mafic rocks. Central Italy orogenic magmatism falls along a mantle-crust mixing trend involving HIMU-FOZO and Upper Crust. Sicily, Sicily Channel, Sardinia and some Tyrrhenian Sea seamounts (anorogenic magmatism) plot along a mantle-mantle mixing trend involving at least two end members (HIMU-EM1).*

(HIMU or FOZO) contamination by upper crustal material transported into the zone of magma genesis by subduction processes. The second trend suggests interaction between different types of mantle reservoirs.

Important problems to address are those dealing with the timing of mantle contamination event(s) beneath peninsular Italy (i.e. the age of subduction processes), and with the significance of HIMU, FOZO and EM1 mantle reservoirs. Although the problem of contamination timing is still debated, geophysical and isotopic evidences favour young events by recent to active subduction. Mantle tomography (Spakman et al., 1993) and S-waves velocity studies (e.g. Panza and Mueller, 1979) have shown that a rigid body occurs within the mantle beneath the Apennines. This mass is actively subducting beneath the eastern Aeolian arc, where deep-focus earthquakes are recorded. Shifting of this subduction zone, from Corsica-Sardinia toward its present position in the southern Tyrrhenian Sea, is responsible for orogenic volcanism inside the Tyrrhenian Sea basin and its time-related migration toward south-east (Beccaluva et al. 1989). Young contamination does not conflict with isotopic evidence, since mafic rocks from single provinces have poorly variable $^{87}Sr/^{86}Sr$ with changing Rb/Sr ratios (see Peccerillo, 2002 for discussion). The significance of HIMU, FOZO, EM1 and other mantle reservoirs are still much debated (see Hofmann, 1997). HIMU compositions are generally believed to represent mantle plumes, whereas EM1 may represent old metasomatised mantle lithosphere. Therefore, the overall picture of the Plio-Quaternary magmatism in Italy would be that of deep mantle material uprising as plumes, mixing with EM1, impinging in an ongoing subduction process and contaminated by subduction-related upper crustal material (Gasperini et al., 2002). Research is actively going on to shed further light on these issues.

Conclusions

The Plio-Quaternary volcanism in Italy shows strong compositional variations, which reveal heterogeneous compositions and complex evolution processes of mantle sources. Both subduction-related and intraplate signatures are observed.

The hypothesis that best explains this complex magmatic setting is continent-continent convergence in which the leading edge of African plate is subducted beneath the Italian peninsula to generate heterogeneous mantle sources that then produced the wide variety of volcanic rocks (from calcalkaline to ultrapotassic) with subduction-related geochemical signatures. Mantle end-member could be partially represented by plume material, on the basis of isotopic evidence. Mixing among various mantle reservoirs generated anorogenic volcanism in Sardinia, Sicily, Sicily Channel and for some Tyrrhenian seamounts. The coexistence of orogenic and anorogenic seamounts on the Tyrrhenian Sea floor reflects both the southeastward migration of the subduction zone, and the mantle uprise beneath the Tyrrhenian Sea basin.

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