Univerza v Ljubljani
Naravoslovnotehniška fakulteta
Oddelek za geologijo

Vodič za predmet

SKLEPNE TERENSKE VAJE

2007 - Severna Italija in Toskana

21.05.–30.05.2007

Avtorja in organizatorja vaj:

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

Ljubljana, 15.05.2007

Trasa ekskurzije v Severno Italijo in v Toskano, 21.5.-30.5.2007

1. Dan: Ponedeljek 21.05 2007

1. Vajont landslide – inženirska geologija, Belluno Bazen – stratigrafija jure in krede (Vodič P22, 14-24)

Spanje: Trento, Giovane Europa

2 Dan: Torek 22.05.2007

1. Monte Baldo: Madona di Corona – stratigrafija in sedimentologija jure
2. Cava Viannini (ammonitico rosso): okrasni kamen

Spanje: Biassa (Cinque Terre), Ostello Tramonti

3. Dan: Sreda 23. 05. 2007

1. Cinque Terre – fliši, ofioliti (Vodič D09, P12)

Spanje: Marina di Massa, Ostello Turimar

4. Dan: Četrtek 24.05. 2007

1. Alpi Apuani - metamorfni kompleks (Vodič P38, B05)
2. Carrara – kamnolom marmorja

Spanje: Marina di Massa, Ostello Turimar

5. Dan: Petek 25.05.2007

1. Larderello – Muzej geotermije (Bertini et al. 2006)

Spanje: Siena, Castle of Selvole (pri kraju Vagliali)

6. Dan: Sobota 26.05. 2007

1. Prost dan v Chiantiju oz. Sieni (Vodič D01)

Spanje: Siena, Castle of Selvole (pri kraju Vagliali)

7. Dan: Nedelja 27.05.2007

1. Le Crete (Vodič P40)
2. Rapolano terme – geokemija, tektonika (Vodič D03, P25)

Spanje: Cetona (pri Chianciano terme)

8. Dan: Ponedeljek 28.05.2007

1. Gubbio – strukturna evolucija narivne strukture Umbria-Marche
2. Assisi – primer popotresne obnove

Spanje: Cetona (pri Chianciano terme)

9. Dan: Torek 29.05. 2007

1. Monte Amiata – vulkanizem (Vodič P25)
2. Lago di Bolsena – vulkanizem (Vodič P09)

Spanje: Ravenna, Youth Hostel Dante

10. Dan: Sreda 30.05.2007

1. Lagune pri Gradežu - recentna sedimentologija (Vodič B24)
2. Tržaški kras

Nazaj v Ljubljano

Seznam udeležencev

Št.	Priimek	Ime
01	Bogunovič	Sergej
02	Borse	Mojca
03	Črne	Alenka Eva
04	Damjanović	Bojan
05	Emeršič	Dejan
06	Gale	Luka
07	Horvat	Jurij
08	Iveković	Aljaž
09	Jamšek	Petra
10	Kastivnik	Jasna
11	Koželj	Lea
12	Lenart	Alenka
13	Leskovar	Nina
14	Marinc	Ana
15	Markelj	Mojca
16	Marn	Anton
17	Martinčič	Darja
18	Medić	Ines
19	Prkič	Matej
20	Purkat	Andraž
21	Ribičič	Mihael
22	Rožič	Boštjan
23	Rupnik	Jaka
24	Skaberne	Dragomir
25	Šmuc	Andrej
26	Štibelj	Kristina
27	Štukovnik	Petra
28	Sušnik	Andreja
29	Tomažin	Andrej
30	Torkar	Anja
31	Udovč	Miran
32	Verbovšek	Timotej
33	Vrhovnik	Petra
34	Zajc	Tina
35	Žibret	Gorazd
36	Žvab	Petra

Zavarovanje CORIS:

št. police: 0438720 (nosilec T. Verbovšek)
24-urna tel. št: +386 1 5192020

ITALIJA

Naša predstavništva

Veleposlaništvo Republike Slovenije
Via Leonardo Pisano 10 00197 Roma
Italy
Tel: (+) 39 06 80 914 310
Fax: (+) 39 06 80 81 471
Elektronska pošta: vri(at)gov.si
Pristojno za: Italija, San Marino, Malta
Nj. eksc. g. dr. Andrej Capuder, veleposlanik

Generalni konzulat Republike Slovenije
Ul./Via S. Giorgio 1 34123 Trst - Trieste
Italy
Tel: (+) 39 040 30 78 55
Fax: (+) 39 040 30 82 66
Elektronska pošta: kts(at)gov.si
g. Jožef Šušmelj, generalni konzul

Konzulat Republike Slovenije
Vialle della Regione 54, 93100 Caltanissetta
Italy
Tel: (+) 39 09 345 981 60
Fax: (+) 39 09 345 570 56
Agostino Ginerva, častni konzul

Konzulat Republike Slovenije
Borgo Pinti, 82/R, 50121 Firenze
Italy
Tel: (+) 39 05 52 420 22
Fax: (+) 39 05 52 450 03
Elektronska pošta: consolato.slovenia(at)libero.it
Piero Tacconi, častni konzul

Tuja predstavništva

Veleposlaništvo Italijanske republike
Snežniška ulica 8
SI-1000 Ljubljana
Slovenija
Tel: (+) 386 1 426 21 94
Fax: (+) 386 1 425 33 02
Elektronska pošta: segreteria.lubiana(at)esteri.it
Nj. eksc. g. Daniele Verga, veleposlanik

Generalni konzulat Italijanske republike
Belvedere 2
SI-6000 Koper
Slovenija
Tel: (+) 386 5 627 37 49
Fax: (+) 386 5 627 37 46
Elektronska pošta: cg.capodistria(at)siol.net
g. Carlo Gambacurta, generalni konzul

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

The Mediterranean region—a geological primer

1 Dept. of Earth and Geoenvironmental Sciences, Univ. of Bologna, Italy. *cavazza@geomin.unibo.it*
2 Institute of Environmental Dynamics, University of Urbino, Italy. *wezel@uniurb.it*

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

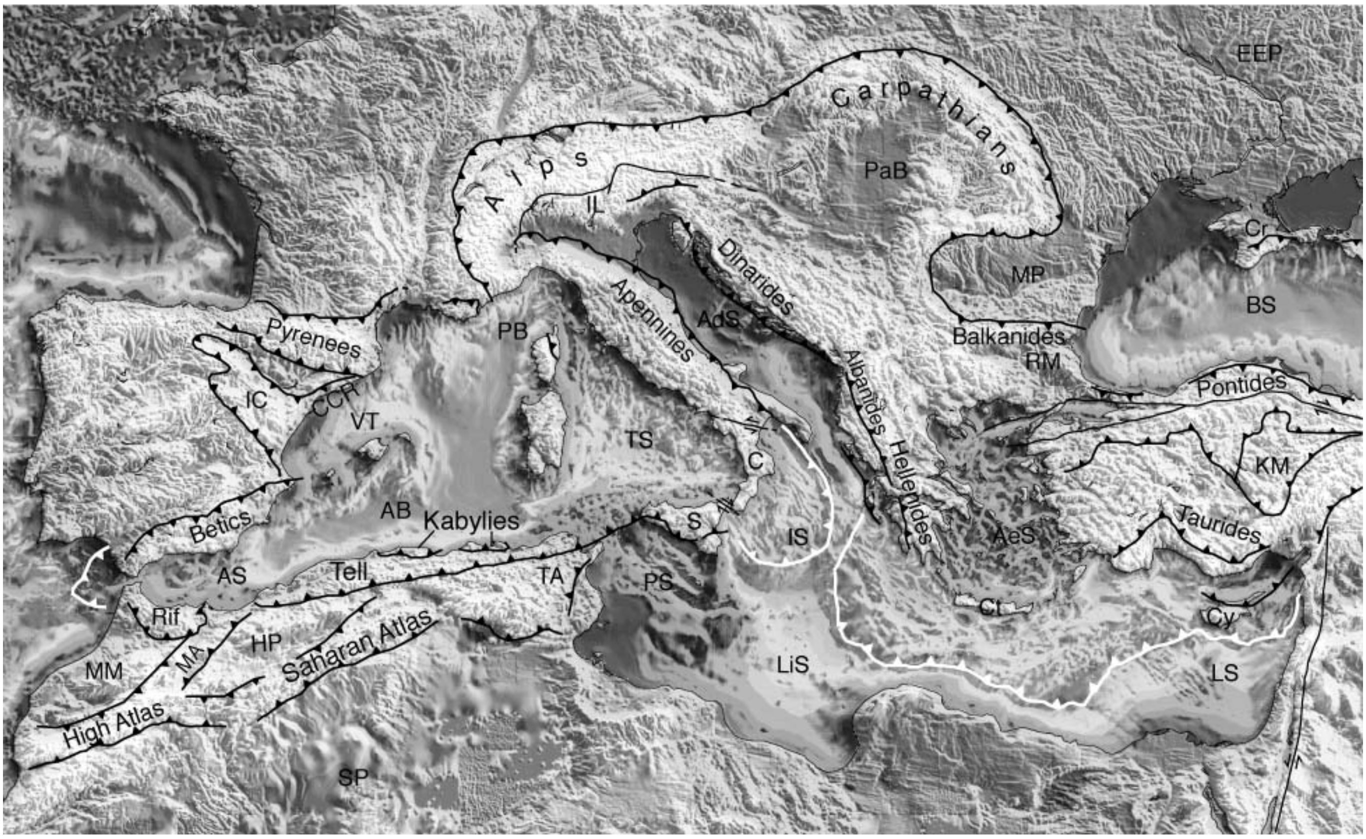


Figure 1 Digital terrain model of the Mediterranean region with major, simplified geological structures. White thrust symbols indicate the outer deformation front along the Ionian and eastern Mediterraean subduction fronts. AB, Algerian basin; AS, Alboran Sea; AdS, Adriatic Sea; AeS, Aegean Sea; BS, 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

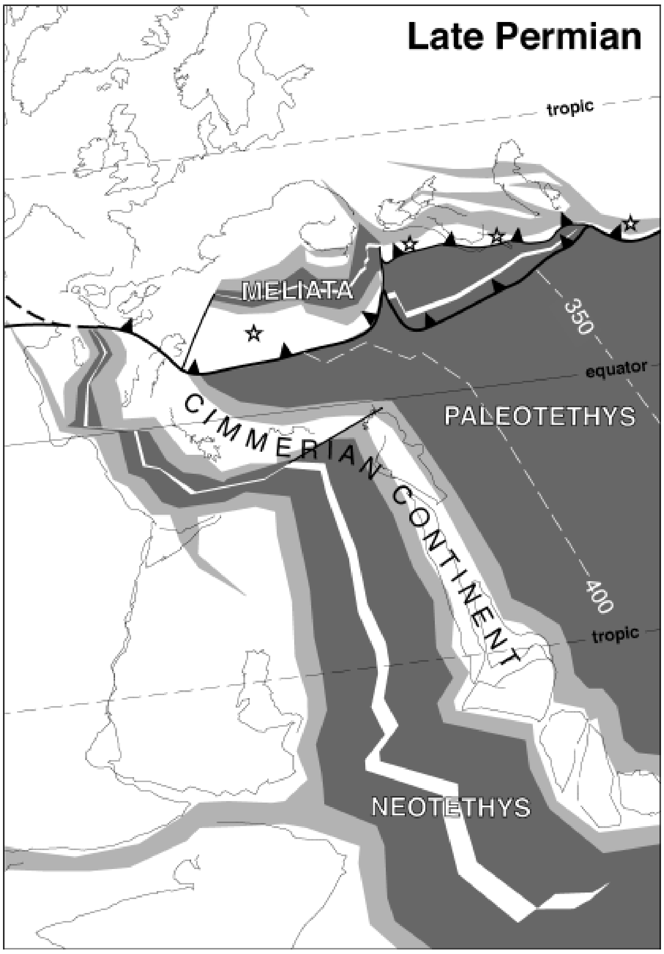


Figure 2 Paleogeographic reconstruction of the western Tethyan area during the late Permian (from Stampfli et al., 2001b, with minor modifications). Stars indicate magmatic activity.

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

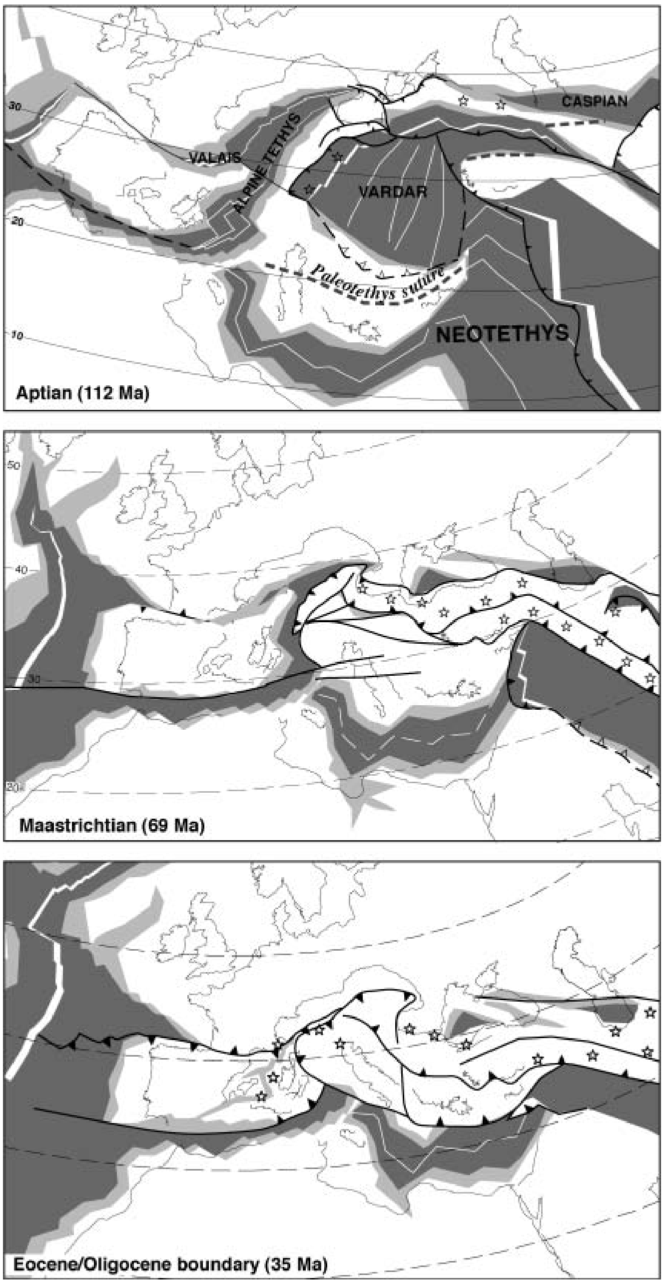


Figure 3 Paleogeographic reconstructions of the western Tethyan area during the Aptian, Maastrichtian and at the Eocene/ Oligocene boundary. Note the progressive narrowing and suturing of the oceanic domains comprised between the Eurasian and Iberia continental blocks to the north and the Africa/Arabia 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 peri-aegean regions (see Durand et al., 1999, and references therein). Overall, Neogene extension in the Mediterranean can be explained as the result of roll-back of the subducting slabs of the Ionian-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).

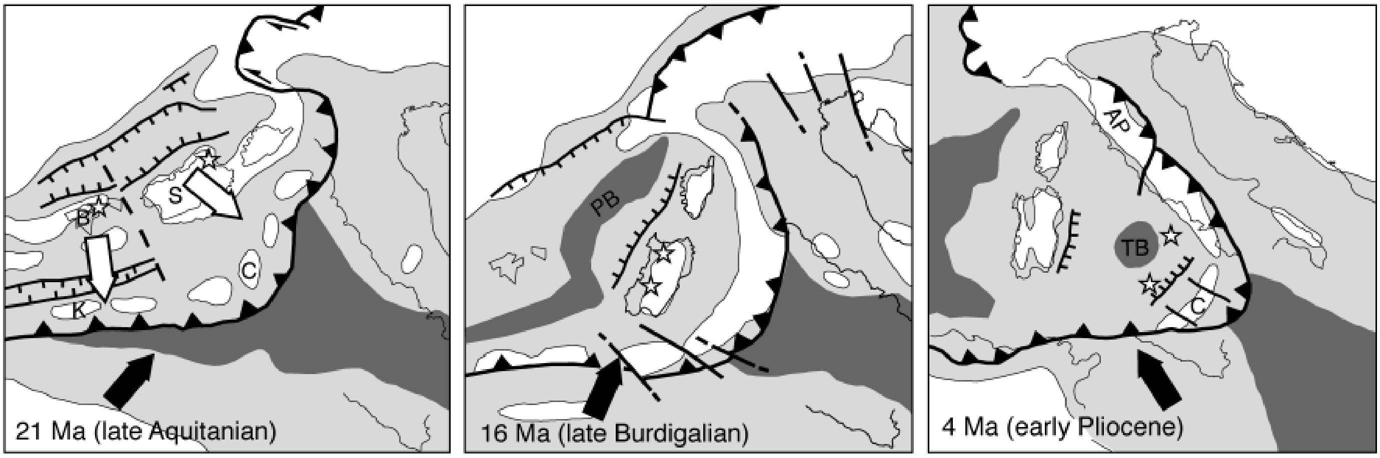


Figure 4 Schematic maps showing the paleotectonic evolution of the W Mediterranean during Neogene time (modified after Bonardi et al., 2001, and Roca, 2001). Only active tectonic elements are shown. White, exposed land; light gray, epicontinental sea; darker gray, oceanic crust. Black arrows indicate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

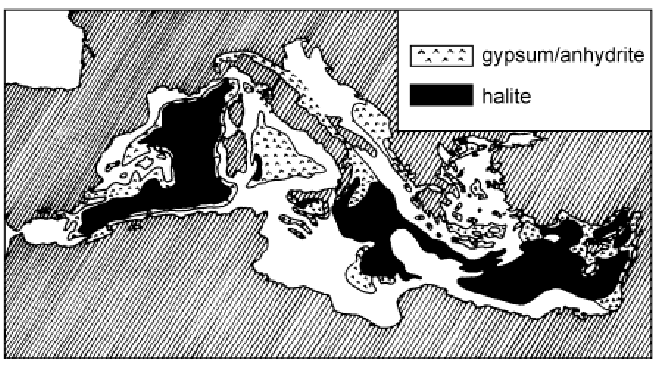


Figure 5 Areal extent of the Messinian evaporites in the Mediterranean region. Modified after Rouchy (1980).

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".

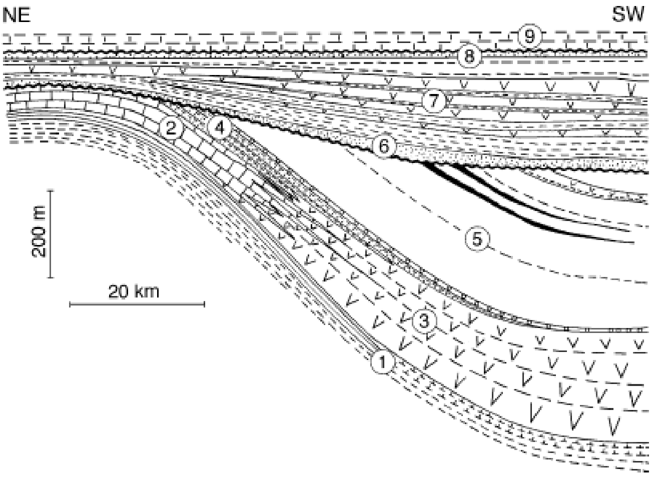


Figure 6 Schematic stratigraphic cross section of the Messinian of Sicily (modified after Decima and Wezel, 1973). 1) Pre-evaporitic clay, marl and diatomite (Tripoli Fm.); 2) evaporitic limestone (Calcare di Base); 3) lower gypsum beds (Gessi di Cattolica); 4) gypsum turbidites; 5) halite and potash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

Acknowledgements

We thank Gerard Stampfli and Gian Battista Vai who reviewed the manuscript.

References

Benson, R.H. and Rakic-El Bied, K., 1995. The Bou Regreg section, Morocco: proposed Global Boundary Stratotype Section and Point of the Pliocene. Notes and Memoires du Service Geologique du Maroc 383, 91 p.

Bonardi, G., Cavazza, W., Perrone, V. and Rossi, S., 2001, Calabria-Peloritani terrane and northern Ionian Sea, *in* Vai, G.B., and Martini, I.P., eds., Anatomy of an Orogen: the Apennines and Adjacent Mediterranean Basins: Dordrecht, Kluwer Academic Publishers, pp. 287-306.

Butler, R.W.H., Lickorish, W.H., Grasso, M., Pedley, H.M. and Ramberti, L., 1995, Tectonics and sequence stratigraphy in Messinian basins, Sicily: constraints on the initiation and termination of the Mediterranean salinity crisis: Geological Society of America Bulletin, v.107, pp. 425-439.

Butler, R.W.H., McClelland, E. and Jones, R.E., 1999, Calibrating the duration and timing of the Messinian salinity crisis in the Mediterranean: linked tectonoclimatic signals in thrust-top basins of Sicily: Journal Geological Society of London, v.156, pp. 827-35.

Cavazza, W. and DeCelles, P.G., 1998, Upper Messinian siliciclastic rocks in southeastern Calabria (southern Italy): paleotectonic and eustatic implications for the evolution of the central Mediterranean region: Tectonophysics, v.298, pp. 223-241.

Cavazza, W., Roure, F., Spakman, Stampfli, G.M. and Ziegler, P.A., eds., 2004, The TRANSMED Atlas: Geological Anatomy of the Mediterranean Region and the Surrounding Areas: Heidelberg, Springer-Verlag, in press.

Channell, J.E.T., D' Argenio, B. and Horvath, F., 1979, Adria, the African promontory, *in* Mesozoic Mediterranean paleogeography: Earth Sciences Reviews, v.15, pp. 213-292

Channell, J.E.T., Rio, D. and Thunell, R.C., 1988, Miocene/Pliocene boundary magnetostratigraphy at Capo Spartivento, Calabria, Italy: Geology, v.16, pp. 1096-1099.

Cita, M.B. and Colombo, L., 1979, Sedimentation in the latest Messinian at Capo Rossello (Sicily): Sedimentology, v.26, pp. 497-522.

Cita, M.B. and Gartner, S., 1973, The stratotype Zanclean foraminiferal and nannofossil biostratigraphy: Rivista Italiana di Paleontologia e Stratigrafia, v.79, pp. 503-558.

Cita, M.B. and McKenzie, J.A., 1986. The terminal Miocene event, *in* Hsu, K.J., ed., Mesozoic and Cenozoic Oceans: Geodynamics Series 15, American Geophysical Union, pp. 123-140.

DeCelles, P. and Cavazza, W., 1995, Upper Messinian fanglomerates in eastern Calabria (southern Italy): response to microplate migration and Mediterranean sea-level changes: Geology, v.23, p.775-778.

Decima, A. and Wezel, F.C., 1973, Late Miocene evaporites of the central Sicilian basin, *in* Ryan, W.B.F., Hsu, K.J. et al., eds., Initial Reports of the Deep Sea Drilling Project, Leg 13: Washington, D.C., U.S. Printing Office, pp. 1234-1240.

de Jong, M.P., Wortel, M.J.R. and Spakman, W., 1993, From tectonic reconstruction to upper mantle model: an application to the Alpine-Mediterranean region: Tectonophysics, v.223, pp. 53-65.

Dercourt, J., Ricou, L.E. and Vrielynk, B., eds., 1993, Atlas Tethys Paleoenvironmental Maps: Editions Gauthier-Villars, Paris, 307 p.

Dercourt, J., Gaetani, M., Vrielynck, B., Barrier, E., Biju-Duval, B., Brunet, M. F., Cadet, J. P., Crasquin, S. and Sandulescu, M., eds., 2000, Peri-Tethys Palaeogeographical Atlas, Paris.

Dixon, J.E. and Robertson, A.H.F., eds., 1984, The Geological Evolution of the Eastern Mediterranean: Geological Society of London Special Publication No.17, 836 p.

Doglioni, C., Busatta, C., Bolis, G., Marianini, L. and Zanella, M., 1996, Structural evolution of the eastern Balkans (Bulgaria): Marine and Petroleum Geology, v.13, pp. 225-251.

Durand, B., Jolivet, L., Horvath, F. and Séranne, M., eds., 1999, The Mediterranean Basins: Tertiary Extension within the Alpine Orogen: Geological Society of London, Special Publication -.156.

Ellouz, N. and Roca, E., 1994, Palinspastic reconstructions of the Carpathians and adjacent areas since the Cretaceous: a quantitative approach, in Roure, F., ed., Peri-Tethyan Platforms: Paris, Editions Technip, pp. 51-78.

Frizon de Lamotte, D., Saint Bezar, B., Bracene, R. and Mercier E., 2000, The two main steps of the Atlas building and geodynamics of west Mediterranean: Tectonics, v.19, pp. 740-761.

Gelati, R., Rogledi, S. and Rossi, M.E., 1987, Significance of the Messinian unconformity-bounded sequences in the Apenninic margin of the padan foreland basin, northern Italy: Memorie Società Geologica Italiana, v.39, pp. 319-323.

Georgiev, G., Dabovski, C. and Stanisheva-Vassileva, G., 2001, East Srednogorie-Balkan rift zone, *in* Ziegler, P.A., Cavazza, W., Robertson, A.H.F. and Crasquin-Soleau, S., eds., Peritethyan Rift/Wrench Basins and Passive Margins: Mémoires du Muséum National d'Histoire Naturelle, Paris, v.186, pp. 259-294

Gutscher, M.A., Malod, J., Rehault, J.P., Contrucci, I., Klingelhoefer, F., Mendes-Victor, L. and Spakman, W., 2002, Evidence for active subduction beneath Gibraltar: Geology, v.30, pp. 1071-1074.

Hilgen, F.J., 1987, Sedimentary rhythms and high-resolution chronostratigraphic correlations in the Mediterranean Pliocene: Stratigraphy Newsletter, v.17, pp. 109-127.

Hsü, K. J., 1972, Origin of saline giants: a critical review after the discovery of the Mediterranean evaporite. Earth Sciences Reviews, v.8, pp. 371-396.

Hsü, K.J., Montadert, L. et al., 1978, History of the Mediterranean salinity crisis, *in* Hsu, K.J., Montadert, L. et al., eds., Initial Reports Deep Sea Drilling Project, v.42, p.1, U.S. Government Printing Office, Washington D.C., pp. 1053-1078.

Jolivet, L., 2001, A comparison of geodetic and finite strain pattern in the Aegean, geodynamic implications: Earth and Planetary Science Letters, v.187, pp. 95-104

Kastens, K., Mascle, J. et al., eds., 1990, Proceeding Ocean Drilling Program, Scientific Results, vol.107: College Station, Texas, 772 p.

Krijgsman, W., Hilgen, F.J., Raffi, I., Sierro, F.J. and Wilson, D.S., 1999, Chronology, causes and progression of the Messinian salinity crisis: Nature, v.400, pp. 652-655.

Linzer, H.G., 1996, Kinematics of retreating subduction along the Carpathian arc, Romania: Geology, v. 24, pp. 167-170.

Lonergan, L. and White, N., 1997, Origin of the Rif-Betic mountain belt: Tectonics, v.16, p.504-522.

Malinverno, A. and Ryan, W.B.F., 1986, Extension in the Tyrrhenian Sea and shortening in the Apennines as result of arc migration driven by sinking of the lithosphere: Tectonics, v.5, pp. 227-245.

Mattei, M., Cipollari, P., Cosentino, D., Argentieri, A., Rossetti, F., Speranza, F. and Di Bella, L., 2002, The Miocene tectono-sedimentary evolution of the southern Tyrrhenian Sea: stratigraphy, structural and paleomagnetic data from the on-shore Amantea basin (Calabrian Arc, Italy): Basin Research, v.14, pp. 147-168.

Mazzoli, S. and Helman, M., 1994, Neogene patterns of relative plate motions for Africa-Europe: some implications for recent central Mediterranean tectonics: Geologische Rundschau, v.83, pp. 464-468.

Montadert, L., Letouzey, J. and Mauffret, A., 1977, Messinian event: seismic evidence, *in* Ross, D.A., Neprochnov, Y.P. et al., eds., Initial Reports, Deep Sea Drilling Project, Initial Reports Deep Sea Drilling Project, v.42, p.1, U.S. Government Printing Office, Washington D.C., pp .1037-1050:

Moores, E.M. and Fairbridge, R.W., eds., 1997, Encyclopedia of European and Asian Regional Geology: London, Chapman & Hall, 804 p.

Muttoni, G., Garzanti, E., Alfonsi, L., Cirilli, S., Germani, D. and Lowrie, W., 2001, Motion of Africa and Adria since the Permian: paleomagnetic and paleoclimatic constraints from Northern Lybia: Earth and Planetary Science Letters, v.192, pp. 159-174.

Okay, A., Sengor, A.M.C. and Gorur, N., 1994, Kinematic history of the Black Sea and its effect on the surrounding regions: Geology, v.22, pp. 267-270.

Patacca, E., Sartori, R. and Scandone, P., 1993, Tyrrhenian basin and Apennines: kinematic evolution and related dynamic constraints, *in* Boschi, E., Mantovani, E. and Morelli, A., eds., Recent Evolution and Seismicity of the Mediterranean Region: Kluwer Academic Publishers, pp. 161-171.

Pfiffner, O.A., et alii, eds., 1996, Deep Structure of the Swiss Alps - Results of NFP/PNR 20: Basel, Birkhauser A.G.

Ratschbacher, L., Frisch, W. and Linzer, H.-G., 1991, Lateral extrusion in the Eastern Alps, part 2: structural analysis: Tectonics, v.10, pp. 257-271.

Rio, D., Sprovieri, R. and Thunell, R., 1991, Pliocene-Pleistocene chronostratigraphy: a re-evaluation of Mediterranean type sections: Geological Society of America Bulletin, v.103, pp. 1049-1058.

Robinson, A.G., ed., 1997, Regional and Petroleum Geology of the Black Sea and Surrounding Region: American Association of Petroleum Geologists Memoir 68, 385 p.

Roca, E., 2001, The Northwest Mediterranean Basin (Valencia Trough, Gulf of Lions and Liguro-Provençal basins): structure and geodynamic evolution, *in* Ziegler, P.A., Cavazza, W., Robertson, A.H.F. and Crasquin-Soleau, S., eds., Peritethyan Rift/Wrench Basins and Passive Margins: Mémoires du Muséum National d'Histoire Naturelle, Paris, v. 186, pp. 671-706.

Rouchy, J.M., 1980, La génèse des évaporites messiniennes de Méditerranée: un bilan: Bulletin Centre Recherche Pau, v.4, p.511-545.

Roure, F., ed., 1994, Peri-Tethyan Platforms: Paris, Editions Technip, 275 p.

Roveri, M., Bernasconi, A., Rossi, M.E. and Visentin, C., 1992, Sedimentary evolution of the Luna Field area, Calabria, southern Italy, in Spencer, A.M., ed., Generation, accumulation and production of Europe Hydrocarbons II. Spec. Publ. Eur. Ass. Petr. Geosc. No.2, Springer-Verlag, Berlin, pp. 217-224.

Royden, L.H., 1993, Evolution of retreating subduction boundaries formed during continental collision: Tectonics, v.12, pp. 629-638.

Royden, L.H. and Horvath, F., eds., 1988, The Pannonian Basin: a Study in Basin Evolution: American Association of Petroleum Geologists Memoir 45, 394 p.

Ryan, W.B.F., Hsu, K.J. et al., 1973, Initial Reports, Deep Sea Drilling Project, v.13, parts 1-2, U.S. Government Printing Office, Washington D.C., 1447 pp.

Salas, R., Guimerà, J., Mas, R., Martin-Closas, C., Meléndes, A. and Alonso, A., 2001, Evolution of the Mesozoic Central Iberian Rift System and its Cenozoic inversion (Iberian chain), *in* Ziegler, P.A., Cavazza, W., Robertson, A.H.F. and Crasquin-Soleau, S., eds., Peritethyan Rift/Wrench Basins and Passive Margins: Mémoires du Muséum National d'Histoire Naturelle, Paris, v. 186, pp. 145-186.

Sengor, A.M.C., 1979, Mid-Mesozoic closure of Permo-Triassic Tethys and its implications: Nature, v.279, pp. 590-593.

Sengor, A.M.C., 1984, The Cimmeride Orogenic System and the Tectonics of Eurasia: Geological Society of America Special Paper 195, 82 p.

Sengor, A.M.C., 1997, Turkey, *in* Moores, E.M. and Fairbridge, R.W., eds., Encyclopedia of European and Asian Regional Geology: London, Chapman & Hall, pp. 732-743.

Spakman, W., van der Lee, S. and van der Hilst, R.D., 1993, Travel-time tomography of the European-Mediterranean mantle down to 1400 km: Physics of the Earth and Planetary Interiors, v.79, pp. 2-74.

Stanley, D.J. and Wezel, F.-C., eds., 1985, Geological Evolution of the Mediterranean Sea: Heidelberg, Springer-Verlag, 589 p.

Stampfli, G.M., Borel, G., Cavazza, W., Mosar, J. and Ziegler, P.A., eds., 2001a, The Paleotectonic Atlas of the PeriTethyan Domain (CD-ROM): European Geophysical Society.

Stampfli, G.M., Borel, G., Marchant, R. and Mosar, J., 2002, Western Alps geological constraints on western Tethyan reconstructions: Journal Virtual Explorer, v.8, pp. 77-106.

Stampfli, G.M., Mosar, J., Favre, P., Pillevuit, A. and Vannay, J.-C., 2001b, Permo-Mesozoic evolutionof the western Tethys realm: the Neo-Tethys East Mediterranean basin connection, *in* Ziegler, P.A., Cavazza, W., Robertson, A.H.F. and Crasquin-Soleau, S., eds., Peritethyan Rift/ Wrench Basins and Passive Margins: Mémoires du Muséum National d'Histoire Naturelle, Paris, v. 186, pp. 51-108.

Vai, G.B., and Martini, I.P., eds., 2001, Anatomy of an Orogen: the Apennines and Adjacent Mediterranean Basins: Dordrecht, Kluwer Academic Publishers, 632 p.

Vai, G.B., 2003, Development of the palaeogeography of Pangea from Late Carboniferous to Early Permian: Paleogeography, Paleoclimatology, Paleoecology, v.196, pp. 1-31.

Vai, G.B., and Ricci Lucchi, F., 1977, Algal crusts, autochthonous and clastic gypsum in a cannibalistic evaporite basin: a case history from the Messinian of northern Apennines: Sedimentology, v.24, pp. 211-244.

Van Couvering, J.A., Castradori, D., Cita, M.B., Hilgen, F.J. and Rio, D., 2000, The base of the Zanclean Stage and of the Pliocene Series: Episodes, v.23, pp. 179-187.

Vially, R. and Trémolières, P., 1996, Geodynamics of the Gulf of Lions: implications for petroleum exploration, *in* Ziegler, P.A. and Horvath, F., eds., Structure and Prospects of Alpine Basins and Forelands: Mémoires du Muséum National d'Histoire Naturelle, Paris, v.170, pp. 129-158.

Wortel, M.J.R. and Spakman, W., 2000, Subduction and slab detachment in the Mediterranean-Carpathian region: Science, v.290, pp. 1910-1917.

Yilmaz, P.O., Norton, I.O., Leary, D. and Chuchla, R.J., 1996, Tectonic evolution and paleogeography of Europe, *in* Ziegler, P.A., and Horvath, F., eds., Structure and Prospects of Alpine Basins and Forelands: Mémoires du Muséum National d'Histoire Naturelle, Paris, v.170, pp. 47-60.

Ziegler, P.A., 1988, Evolution of the Arctic-North Atlantic and Western Tethys: American Association of Petroleum Geologists Memoir 43, 198 p.

Ziegler, P.A., 1990, Geological Atlas of Western and Central Europe (2nd edition): Shell International Petroleum Mij. B. V., The Hague, 232 p.

Ziegler, P.A., Cavazza, W., Robertson, A.H.F. and Crasquin-Soleau, S., eds., 2001, Peritethyan Rift/Wrench Basins and Passive Margins: Mémoires du Muséum National d'Histoire Naturelle, Paris, v. 186, 762 p.

Ziegler, P.A. and Roure, F., 1996, Architecture and petroleum systems of the Alpine orogen and associated basins, *in* Ziegler, P.A. and Horvath, F., eds., Structure and Prospects of Alpine Basins and Forelands: Mémoires du Muséum National d'Histoire Naturelle, Paris, v.170, pp. 15-45.

William Cavazza *is Professor of Stratigraphy and Sedimentology at the University of Bologna (Italy). He received a Ph.D. from UCLA in 1989. His research focuses on the combined application of structural geology, sedimentology, and stratigraphy to the analysis of ancient and modern sedimentary basin fills. Other research interests include fission-track analysis, strontium-isotope stratigraphy, and the paleogeographic-paleotectonic evolution of the Mediterranean region. He has published on the geology of the Eastern Alps, the Rio Grande rift, the Rocky Mountains, the Mojave Desert, the northern Apennines, southern Italy and Corsica. He is currently the chairperson of the Mediterranean Consortium for the 32nd International Geological Congress.*

Forese Carlo Wezel *is Professor of Stratigraphy at Urbino University (Italy). His research activities are concerned with global geology, the geology of the central Mediterranean region (both on-land and offshore), mass biotic extinctions, and Holocene paleoclimatology. He served as chief scientist of many research cruises and as scientist of DSDP Leg 13. He has been member of the editorial board of Tectonophysics and Terra Nova and editor of volumes of proceedings of national and international workshops. He is one of the founders of the European Union of Geosciences (EUG) of which was appointed secretary for the period 1991–93. He is currently corresponding member of the Italian National Academy (of Lincei) and chairperson of the Advisory Board for the 32nd International Geological Congress.*

by Giorgio V. Dal Piaz, Andrea Bistacchi, and Matteo Massironi

Geological outline of the Alps

Dipartimento di Geologia, Paleontologia e Geofisica, Università di Padova, via Giotto 1, 35137 Padova, Italy.
E-mail: *giorgio.dalpiaz@unipd.it*

The Alps were developed from the Cretaceous onwards by subduction of a Mesozoic ocean and collision between the Adriatic (Austroalpine-Southalpine) and European (Penninic-Helvetic) continental margins. The Austroalpine-Penninic wedge is the core of the collisional belt, a fossil subduction complex which floats on the European lower plate. It consists of continental and minor oceanic nappes and is marked by a blueschist-to-eclogite-facies imprint of Cretaceous-Eocene age, followed by a Barrovian overprint. The collisional wedge was later accreted by the Helvetic basement and cover units and indented by the Southalpine lithosphere, which in turn was deformed as an antithetic fold-and-thrust belt.

Introduction

The Alps are the typical example of a collisional belt, the mountain range where the nappe theory was conceived and rapidly consolidated (see Dal Piaz, 2001, and Trümpy, 2001, for historical reviews). This belt was generated by the Cretaceous to present convergence of the Adriatic continental upper plate (Argand's African promontory) and a subducting lower plate including the Mesozoic ocean and the European passive continental margin. Complete closure (Eocene) of the ocean marked the onset of the Adria/Europe collision. The collisional zone is represented by the Austroalpine-Penninic wedge, a fossil subduction complex, showing that even coherent fragments of light continental crust may be deeply subducted in spite of their natural buoyancy.

In a map view, the Alps extend from the Gulf of Genoa to Vienna, through the French-Italian western Alpine arc and the east-west-trending central and eastern Alps (Figure 1). South of Genoa, the Alpine range disappears, because it collapsed and was fragmented during the Late Neogene opening of the Tyrrhenian basin (southern segments of the Alpine belt are preserved in Corsica and Calabria). To the east, the former connection between the Alpine and Carpathian belts is buried below the Neogene fill of the Vienna and Styria (Pannonian) basins. The maximum elevations of the Alps are the Mont Blanc (4888 m) and some dozen of summits which exceed 4000 m, whereas most of the Alpine orogen extends below the surface, to a depth of nearly 60 km. Large-wave undulations coupled with orogen-parallel denudation by low-angle normal faults and differential uplift expose the 20–25 km thick upper part of the nappe edifice, going from structural depressions, where the capping Austroalpine units are preserved, to the core of the deepest Penninic Ossola-Ticino window. The remaining buried part has been imaged by deep reflection seismic profiles and other geophysical soundings (Roure et al., 1990; Pfiffner et al., 1997; Transalp Working Group, 2002).

Our aim is a synthetic overview of the structural framework and geodynamic evolution of the Alps, mainly addressed to geoscientists from far-off countries. Tectonic units and essential lithology are represented in the northern sheets (1–2) of the Structural Model of Italy, scale 1:500,000 (Bigi et al., 1990; edited by SELCA, Firenze, e-mail: *selca@selca-cartografie.it*). These maps facilitate readers' approach to the complex geology of the Alps. Due to space limitations, only a few special publications and regional syntheses with extended references are quoted here, concerning the French-Italian Alps (Roure et al., 1990; Michard et al., 1996; Dal Piaz, 1999), Switzerland (Trümpy et al., 1980; Pfiffner et al., 1997), Austria (Flügel and Faupls, 1987; Plöchinger, 1995; Neubauer and Höck, 2000), Southern Alps (Bertotti et al., 1993; Castellarin et al., 1992), tectonics (Coward et al., 1989, Ratschbacher et al., 1991), pre-Mesozoic geology (von Raumer and Neubauer, 1993), metamorphic features (Frey et al., 1999) and geochronology (Hunziker et al., 1992). Daniel Bernoulli and Gabriel Walton are warmly acknowledged for reviews.

Structural framework

According to the direction of tectonic transport, the Alps may be subdivided into two belts of differing size, age and geological meaning: 1) the Europe-vergent belt, a thick collisional wedge of Cretaceous-Neogene age, consisting of continental and minor oceanic units radially displaced towards the Molasse foredeep and European foreland; 2) the Southern Alps, a minor, shallower (non-metamorphic) and younger (Neogene) thrust-and-fold belt displaced to the south (Adria-vergent), which developed within the Alpine hinterland of the Adriatic upper plate, far from the oceanic suture. These belts are separated by the Periadriatic (Insubric) lineament, a major fault system of Oligocene-Neogene age.

From top to bottom and from the internal to the external side, the principal Europe-vergent tectonic domains are (Figure 1): i) the Austroalpine composite nappe system, derived from the distal (ocean-facing) part of the Adriatic passive continental margin, which mainly developed during the Cretaceous (Eoalpine) orogeny; ii) the Penninic zone, a stack of generally metamorphic nappes scraped off the subducting oceanic lithosphere and European passive continental margin (distal part), mainly accreted during the Paleogene; its outer boundary is the Penninic frontal thrust; iii) the Helvetic zone, consisting of shallower basement slices and décolled cover units derived from the proximal part of the European margin, mainly imbricated from the Oligocene onwards. The vertical nappe sequence and their deformation age generally reflect the outward propagation of the orogenic front.

The Helvetic zone is thrust over the Molasse foredeep, a northward-thinning sedimentary wedge which developed from the Oligocene to the Late Miocene, with repeated alternations of shallow marine and freshwater deposits. Its imbricated inner zone (Subalpine Molasse) was buried to a distance of over 20 km below the frontal thrust belt. In the outer French-Swiss Alpine arc, the Molasse basin is bounded by the thin-skinned Jura fold-and-thrust belt of Late Miocene-Early Pliocene age.

The anatomy of the Alps has been explored by the deep seismic experiments mentioned above, identifying two distinct Moho surfaces, i.e., the Adriatic and the deeper European Moho, gently bending from the Alpine foreland to the deep base of the collisional wedge (Figure 2). This means that the overall setting of the Alps is asymmetric, the orogen was dominated by Europe-vergent displacements, and the antithetic Southalpine belt is only a superficial feature

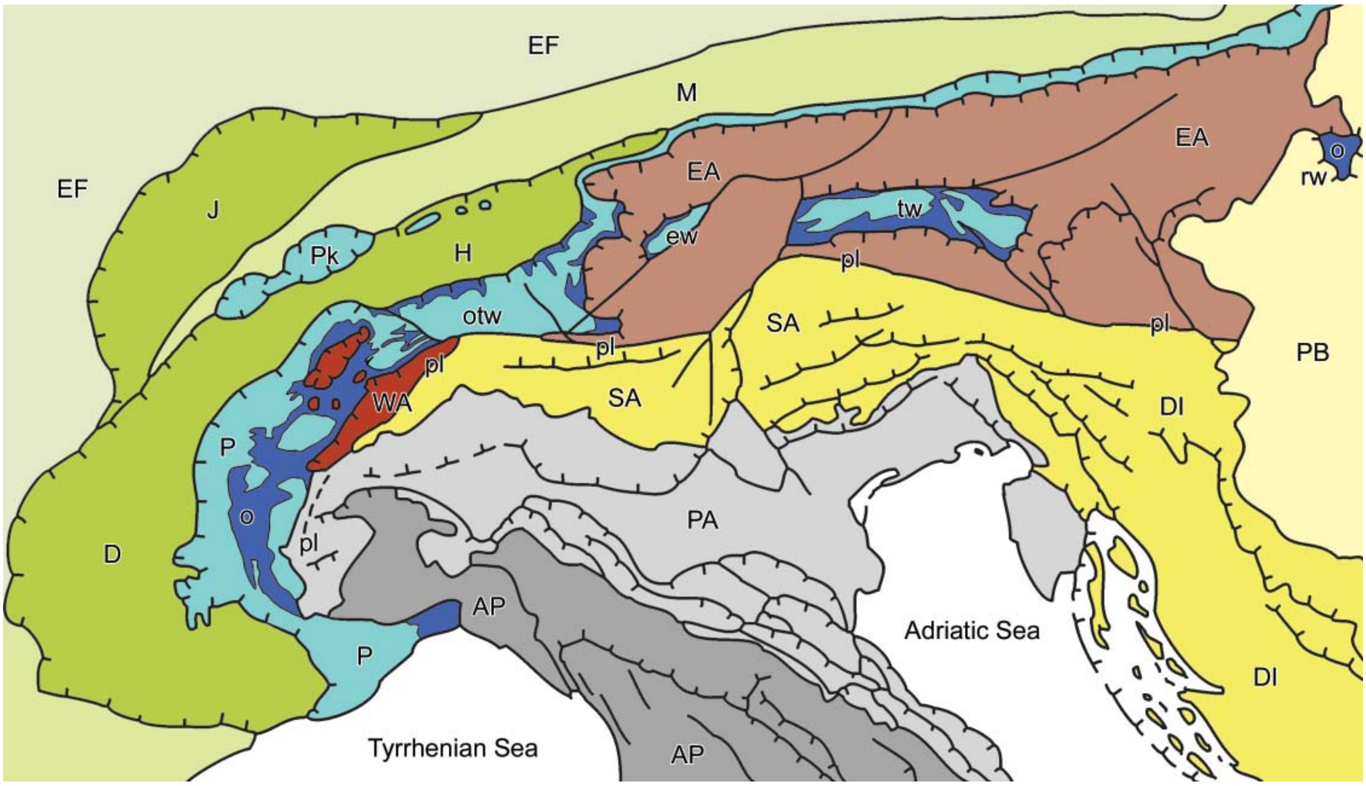


Figure 1 Tectonic map of Alps - (1) Europe-vergent collisional belt: i) Western (WA) and Eastern (EA) Austroalpine; ii) Penninic domain: continental and ophiolitic (o) nappes in western Alpine arc (P) and tectonic windows (otw: Ossola-Ticino, ew: Engadine, tw: Tauern, rw: Rechnitz); Prealpine klippen (Pk); iii) Helvetic-Dauphinois (H-D) domain; iv) Molasse foredeep (M); v) Jura belt (J). (2) Southern Alps (SA), bounded to the north by the Periadriatic lineament (pl). Pannonian basin (PB), European (EF) and Po Valley-Adriatic (PA) forelands, Dinaric (DI) and Apenninic (AP) thrust-and-fold belts.

within the Adriatic upper plate. If we integrate surface geology with interpretation of seismic images, the Europe-vergent belt is a mantle-free crustal wedge which tapers to the north, floats on top the European lower plate and is indented, to the south, by the present Adriatic (Southern Alps) lithosphere (Figure 2). Both continental plate margins originally extended way into the Penninic-Helvetic and Austroalpine domains presently incorporated into the collisional belt. This wedge groups the Austroalpine, Penninic and Helvetic units, and may be subdivided into two diachronous parts: i) the internal, older part (Austroalpine-Penninic), which forms now the axial zone of the Alps, is a fossil subduction complex which includes the Adria/Europe collisional zone; it is marked by one or more ophiolitic units (in different areas) and displays polyphase metamorphism evolving from blueschist or eclogite facies imprint (Cretaceous-Eocene subduction), locally coesite-bearing, to a Barrovian overprint (mature collision, slab break-off) of Late Eocene-Early Oligocene age (Frey et al., 1999); ii) the outer, younger part (Helvetic) is made up of shallower basement thrust-sheets and largely detached cover units derived from the proximal European margin, which escaped the low-T subduction regime and, from the Oligocene, were accreted in front of the exhumed Austroalpine-Penninic wedge.

In the following, we outline the essential features of the Europe-vergent Austroalpine, Penninic and Helvetic tectonic domains and the antithetic Southern Alps.

The Austroalpine thrust units

The Austroalpine is subdivided into two sectors (western and eastern), based on contrasting distribution, structural position, and main deformation age.

The western Austroalpine consists of the Sesia-Lanzo zone and numerous more external thrust units traditionally grouped as Argand's Dent Blanche nappe. These units override and are partly tectonically interleaved with the structurally composite ophiolitic Piedmont zone, the major remnant of the Mesozoic ocean. Two groups of Austroalpine units are identified: i) the upper outliers (Dent Blanche-Mt. Mary-Pillonet) and the Sesia-Lanzo inlier occur on top of the collisional nappe stack; they overlie the entire ophiolitic Piedmont zone and display a blueschist to eclogite facies metamorphism of Late Cretaceous age; ii) the Mt. Emilius and other lower outliers are interleaved with the Piedmont zone, along the tectonic contact between the upper (Combin) and lower (Zermatt-Saas) ophiolitic nappes, and display an eclogitic imprint of Eocene age. Therefore, these groups of nappes originated from different structural domains, were diachronously subducted to various depths, and finally juxtaposed during their later exhumation.

In the central Alps, east of the Ossola-Tessin window, the western Austroalpine may be correlated to the Margna nappe (Staub's interpretation), which is thrust over the Malenco-Avers ophiolite and overlain by the Platta ophiolite, both being potential homologues of the Piedmont zone. The Platta nappe is in turn the tectonic substratum of the eastern Austroalpine system. This means that the western Austroalpine and Margna nappes are presently located at a structural level lower than that of the capping eastern Austroalpine.

The eastern Austroalpine is a thick pile of cover and basement nappes which extends from the Swiss/Austrian border to the Pannonian basin (Figure 1). Its allochthony with respect to the Penninic zone is documented by Mesozoic and ophiolitic units exposed in the Engadine, Tauern and Rechnitz windows. To the north, the Austroalpine overrides the outer-Penninic Rheno-Danubian flysch belt; to the south, it is juxtaposed to the Southalpine basement along the Periadriatic fault system. Part of the Austroalpine displays an eclogitic to Barrovian metamorphism dated as early-mid Cretaceous (Eoalpine; Frey et al., 1999). In addition, thrust surfaces are sealed by Gosau beds (Coniacian-Eocene intramontane basins), testifying that the principal tectono-metamorphic history of the eastern Aus-

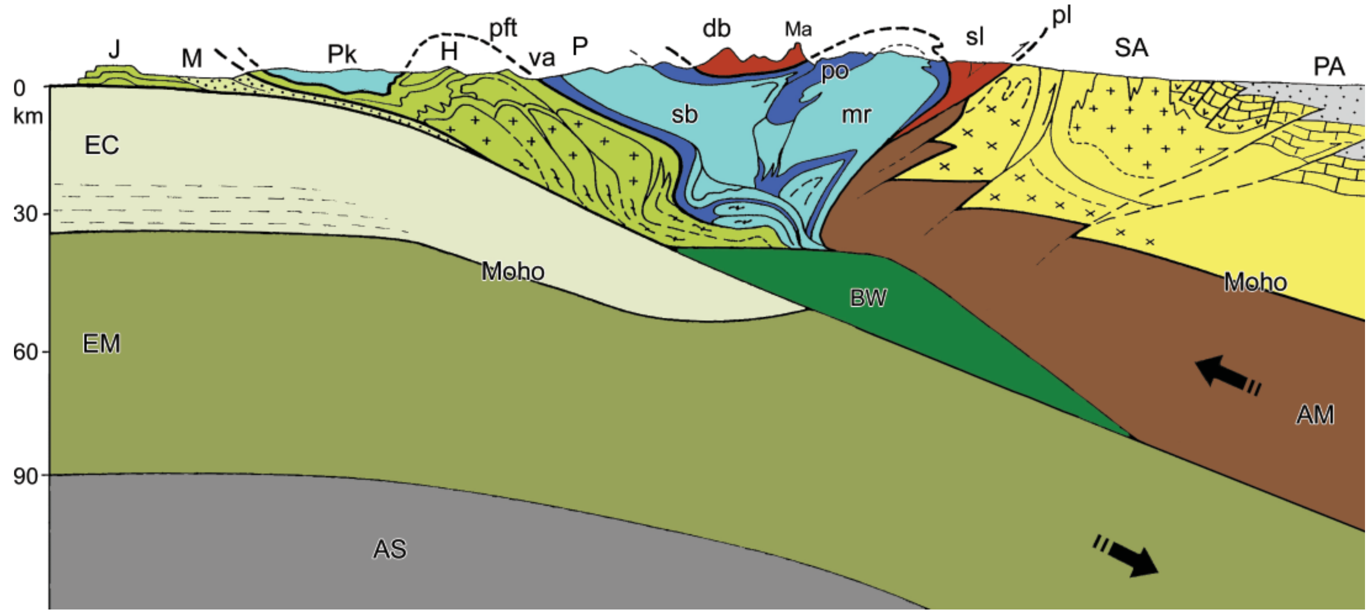


Figure 2 Lithospheric section of north-western Alps - 1) Austroalpine: Sesia-Lanzo inlier (sl) and Dent Blanche nappe s.l. (db), including Matterhorn (Ma); 2) Penninic domain (P): Piedmont ophiolitic units (po), Monte Rosa (mr) and Grand St. Bernard (sb) nappes, underlain by lower Penninic and outer Penninic Valais zone (va), Penninic klippen (Pk), Penninic frontal thrust (pft); 3) Helvetic basement slices and cover nappes (H); 4) Molasse foredeep (M); 5) Jura belt (J); 6) buried wedge (BW) of European mantle or eclogitized crustal units; 7) European lithosphere: continental crust (EC) and mantle (EM); asthenosphere (AS); 8) Adriatic lithosphere: antithetic belt of Southern Alps (SA) and mantle (AM); Periadriatic fault system (pl); 9) Padane-Adriatic foreland (PA).

troalpine is older (pre-Late Cretaceous) than that of the western Austroalpine (Late Cretaceous-Eocene).

The eastern Austroalpine is subdivided into two (Structural Model of Italy) or three (Austrian literature) main groups of nappes. The Upper Austroalpine encompasses the Northern Calcareous Alps and some phyllitic basement nappes occurring west (Steinach klippe), south-east (Gurktal nappe, Graz Paleozoic) and north (Graywacke zone) of the Tauern window. The Northern Calcareous Alps are an imbricated pile of décollement cover nappes made up of Permian-Mesozoic clastic to carbonate deposits, including platform (Hauptdolomit) and basin (Hallstatt) sequences, mainly detached from the Graywacke zone along evaporite-bearing shales. The Middle Austroalpine groups most of the basement and minor cover units of the eastern Alps. The Silvretta, Oetztal and Ortler-Campo nappes occur west of the Tauern window, followed to the south by the Ulten-Tonale nappe. The latter is a fragment of Variscan lower continental crust with eclogitic relics and slices of garnet-spinel peridotite. Similar basement and cover nappes occur east of the Tauern window, including the Speick ophiolite (Variscan) and some basement units (Koralpe-Saualpe, Sieggraben) which display an Eoalpine eclogite-facies imprint in Permian mafic protoliths (Thöni, in Frey et al., 1999). The Lower Austroalpine includes some cover and basement units exposed along the western (Err-Bernina), central (Innsbruck Quartz-phyllite, Radstatt system) and eastern edge (Semmering-Wechsel) of the Austroalpine ranges. The Innsbruck Phyllite (Paleozoic) is overthrust by the Reckner nappe, a Mesozoic ophiolite which displays a blueschist facies imprint of Eocene age.

The Penninic zone

Penninic is the classic name used to group the continental and oceanic nappes which issued from the distal European continental margin and the Mesozoic ocean (one or more branches), all belonging to the subducting lower plate. The original position of the ophiolitic units with respect to the spreading center (now lost) is unknown.

In the western Alps, the Penninic zone includes, from top to bottom: i) the ophiolitic Piedmont zone; ii) the inner-Penninic Dora-Maira, Gran Paradiso and Monte Rosa continental basement nappes; iii) the middle-Penninic Grand St. Bernard (Briançonnais) composite nappe system; iv) the lower-Penninic nappes of the Ossola-Ticino window, and outer-Penninic Valais zone, including ophiolitic units and/or flysch nappes, bounded by the Penninic frontal thrust; v) the Prealpine klippen, a stack of décollement cover nappes in the French-Swiss Alps which, at the onset of subduction, were detached from various units of the Austroalpine-Penninic wedge and later displaced over the Helvetic domain. In the central Alps, the Ossola-Ticino window (lower Penninic) is overlain, to the east, by the Tambo and Suretta continental nappes (middle-inner Penninic), capped in turn, as previously seen, by the Malenco-Avers ophiolite, Margna nappe and Platta ophiolite.

The outer-Penninic extends from the Valais zone (northwestern Alps) through the Grisons to the Rheno-Danubian flysch belt (eastern Alps), constituting the frontal part of the Penninic wedge. It is composed of décollement units, mainly Cretaceous (western side) or Cretaceous-Eocene siliciclastic to carbonatic turbidites, locally with pre-flysch sequences. A few ophiolitic fragments point to the oceanic origin of these deposits.

In the eastern Alps, the Penninic zone is exposed in the Engadine, Tauern and Rechnitz windows. The Tauern nappe stack consists of the ophiolitic Glockner nappe and the underlying basement and cover nappes of European origin (mid- and/or inner-Penninic), i.e., i) the Venediger-Zillertal and Tux, forming the core of two gigantic antiforms in the western side of the window; ii) the Granatspitz dome in the central window; iii) the Sonnblick, Siglitz, Hochalm-Ankogel, Gössgraben and Mureck units in the southeastern side.

The ophiolitic Piedmont zone and its eastern extension are subdivided into blueschist and eclogite facies units. Other differences concern the lithostratigraphic setting, varying between: i) carbonate to terrigenous flysch-type metasediments (calcschists s.l.), often including multiple interleavings of metabasalt and major ophiolitic bodies; ii) large slices of normal to anomalous oceanic lithosphere, consisting of antigorite serpentinites (from mantle peridotite), in places mantled by ophicarbonate-ophicalcite breccias (western Alps, Platta) and/or intruded by discontinuous metagabbro bodies, and overlain by massive to pillow tholeiitic metabasalts, manganiferous metacherts (Middle-Late Jurassic), impure marbles, syn-orogenic deposits, and subduction mélanges. Disregarding the metamorphic imprint, the former association roughly recalls the External Ligurides (Northern Apennines), which are characterized by mélanges and olistolith-rich flysch sequences, whereas the latter may be correlated with the slices of oceanic lithosphere of the Internal Ligurides.

Continental nappes of the Penninic zone are décolled cover units and large, generally thin basement slices, in places still carrying complete or partial cover sequences. The basement includes Variscan and locally older metamorphic units, intruded by Upper Paleozoic granitoids. The post-Variscan sedimentary cover begins with Upper Paleozoic and/or Lower Triassic clastic deposits (e.g., Grand St. Bernard, Tauern), followed by Triassic platform and Jurassic platform to basinal carbonate sequences, locally extending to the Cretaceous (internal Penninic) or Eocene (Briançonnais) synorogenic deposits. The entire zone is marked by a severe Alpine metamorphic overprint, with the exception of the Prealpine klippen. The internal Penninic basement in the western and central Alps displays eclogitic metamorphism (coesite-bearing in Dora Maira; Chopin, 1984, in Frey et al., 1999) of Eocene age, also recorded in a few lower Penninic basement nappes (e.g., Adula-Cima Lunga), whereas a blueschist facies imprint is shown by the Grand St. Bernard system. In contrast, the continental nappes of the Tauern window are dominated by a greenschist to amphibolite facies Barrovian overprint (collisional metamorphism), which obliterated most of the previous high-P features.

The Helvetic-Dauphinois zone

The Helvetic and Dauphinois zone (French part) consists of prominent crystalline duplexes, sedimentary cover units, and décollement nappes. Updomed basement thrust-sheets of metamorphic and granitoid composition are exposed in the Argentera-Mercantour, Pelvoux (Haut-Dauphiné), Belledonne-Grandes Rousses, Aiguilles Rouges-Mont Blanc and Aar-Gotthard external “massifs”. Polymetamorphic (Variscan and older) and monometamorphic (only Variscan) basement units may be distinguished, evolving from an Ordovician subduction cycle, through Variscan collision, nappe stacking and regional metamorphism, to Carboniferous erosion, orogenic collapse, later intrusions and wrench faulting. The Variscan basement is unconformably covered by thick sedimentary sequences of Late Carboniferous to Eocene/Oligocene age, characterized by early Mesozoic asymmetric fault-bounded rift basins and passive-margin sequences.

The Helvetic-Dauphinois domain was strongly deformed from the Late Oligocene onwards, when the orogeny propagated onto the proximal European margin. Rift faults were largely reactivated and inverted. Basement and cover units were accreted in front of the exhumed Austroalpine-Penninic collisional wedge, and partly recrystallized in anchizonal (deep burial diagenesis) to greenschist, locally amphibolite facies conditions (southern Gotthard).

The Helvetic and Ultrahelvetic nappes are décolled cover sheets and minor recumbent folds, mainly consisting of Mesozoic carbonates and Paleogene flysch which were detached along Triassic evaporites and Middle Jurassic and/or Lower Cretaceous shales. Similar cover sheets occur in the Subalpine Ranges (French Alps), west (Chartreuse) and south (Devoluy-Ventoux) of the Belledonne and Pelvoux massifs, where the Dauphinois sedimentary cover was detached and extensively deformed.

At the Swiss-Austrian boundary, the Helvetic zone dramatically narrows and, in the Eastern Alps, is reduced to some décollement cover sheets discontinuously exposed in front and below the Rheno-Danubian flysch belt.

Southern Alps

The Southern Alps are the typical example of a deformed passive continental margin in a mountain range (Bertotti et al., 1993). Until the Oligocene, this Adriatic domain was the gently deformed retro-wedge hinterland of the Alps, intensively reworked only at its eastern edge by the Paleogene Dinaric belt. From the Neogene, the Southalpine thrust-and-fold belt developed and progressively propagated towards the Adriatic foreland, mainly reactivating Mesozoic extensional faults (Castellarin et al., 1992). Its front is mainly buried beneath the alluvial deposits of the Po Plain and sealed by Late Miocene to Quaternary deposits. To the north, the Southern Alps are bounded by the Periadriatic lineament.

A complete crustal section of the Southern Alps is exposed at the surface: thick cover successions are dominant in the central (Lombardy) and eastern sector (Dolomites), whereas the basement is nearly continuous from the central sector (upper-intermediate crust: Orobic Alps and area of the Como-Maggiore lakes) to the western edge (Ivrea zone), where the lower continental crust crops out.

The crystalline basement includes various kinds of Variscan metamorphic rocks derived from sedimentary and igneous protoliths, later intruded by igneous bodies of Permian age. Among them is the famous Ivrea gabbro batholith, which was emplaced at the base of an attenuated gneissic crust (Kinzigitic complex). Below the Variscan unconformity regional metamorphism increases from very low-grade (Carnian Alps), to greenschist facies (Venetian region, east of Adamello), and medium- to high-grade conditions (central and western Southern Alps). This imprint predates exhumation, extensive erosion and the discordant deposition of a Westphalian (Lombardy, Ticino) to Lower Permian clastic and volcanic sequences. A new sedimentary cycle developed in the Late Permian, marked by continental deposits grading eastwards into shallow marine sediments. In the Triassic, the Southalpine domain was flooded and characterized by carbonate platform and basin systems, with regional evidence of andesitic-shoshonitic magmatism, mainly Ladinian in age. Rifting developed from the Norian to the early Middle Jurassic, leading to the opening of the Piedmont-Ligurian ocean, when the Austroalpine and Southalpine domains became the subsiding passive continental margin of Adria. Pre-existing structures were reactivated as normal faults and persisted to the Middle Jurassic, when pelagic deposition became dominant. The Cretaceous-Paleogene sequences are discontinuously preserved pelagic and flysch deposits, whereas most of the subsequent succession was eroded during the Oligocene-Present orogenic evolution and related uplift.

Geological history

The Alpine-Meditarranean area is a mobile zone which, from the Precambrian, was reworked and rejuvenated by recurring geodynamic processes. The pre-Alpine history may be reconstructed in the Southern Alps and, to various extents, also in areas of the Austroalpine, Helvetic and Penninic domains which are weakly overprinted by the Alpine orogeny.

Variscan and older evolution

The Paleozoic orogeny and Variscan collision gave rise to Pangea by the merging of the Gondwana and Laurasia megacontinents and the consumption of intervening oceans. The future Alpine domains were located along the southern flank of this orogen. The classic “Variscan” term was coined to define the Carboniferous collision in central Europe, but earlier events of Ordovician to Devonian age were later documented, suggesting the existence of an essentially continuous Paleozoic orogeny. Traces of older orogeny are locally preserved. As a whole, the pre-Permian evolution of the Alps may be summarized as follows:

1) U-P data on zircon and Nd model ages document a Precambrian history. The oldest zircons found in various polymetamorphic basement units refer to Precambrian clastic material eroded from extra-Alpine sources. The occurrence of Proterozoic-Early Cambrian ocean-floor spreading, island-arc activity, and bimodal volcanism is documented in the European and Adriatic basement, with debated traces of Precambrian amphibolite-eclogite facies metamorphism (Silvretta). Cambrian fossils are occasionally found.

2) Early Paleozoic northward subduction of the ocean flanking Gondwana to the north is recorded in eastern Austroalpine and Helvetic basement units, with recycled Precambrian rocks, mafic-ultramafic ophiolites and marginal basin remnants. Subduction is inferred from the accretion of a Paleozoic orogenic wedge, eclogitic relics in mafic and felsic rocks, and calc-alkaline island-arc magmatism (460–430 Ma): these traces are mainly preserved in the Variscan metamorphic basement of a few Southalpine, Austroalpine and Helvetic-Dauphinois units.
3) The Silurian-Early Carboniferous continental collision (classic Variscan orogeny) generated crustal thickening by nappe stacking, low- to high-grade regional metamorphism in relaxed or thermally perturbated conditions, anatectic processes, post-nappe deformation, flysch deposition, and syn-orogenic igneous activity (350–320 Ma). In the Late Carboniferous, the collapsed Variscan belt was sealed by clastic deposits (Variscan unconformity) and intruded by post-orogenic plutons.

Permian-Mesozoic evolution

Variscan plate convergence ended around the Carboniferous-Permian boundary, when transcurrent and transtensive tectonics became dominant on the scale of the Eurasian plate. Asthenospheric upwelling, thermal perturbation and lithosphere attenuation marked the Early Permian onset of a new geotectonic regime in the future Adriatic domain. The Permian evolution was characterized, on a lithospheric scale, by extensional detachments, asymmetric extension (with Adria as an upper plate) and widespread igneous activity from asthenospheric sources. In the Austroalpine and Southalpine basement, igneous activity began with underplating of Early Permian gabbro batholiths, emplaced below and within rising segments of attenuated continental crust, and then recrystallizing under granulitic conditions. The heated crustal roof generated anatectic melts which partly migrated to upper crustal levels. This cycle is recorded by shallower granitoids and fault-bounded basins filled by clastic sediments and volcanic products.

A calc-alkaline to shoshonitic igneous pulse developed in the Middle Triassic, mainly in the Southern Alps, and was produced by extensional partial melting of previously enriched mantle sources (Variscan subduction). From the Late Triassic, continental rifting between Adria (Africa) and Europa generated the Alpine Tethys, a deep-water seaway marked first by listric faults, half-grabens and syn-rift deposits. Rifting ended in the Middle Jurassic when the Mesozoic ocean began to spread. This age is constrained by deposition of radiolarian cherts on subsiding continental blocks in late syn-rift Early Bajocian times, and the evolution of oceanic crust from the Middle Bathonian onwards, coeval with the oldest occurrences in the Central Atlantic. The Austroalpine-Southalpine domains became the distal and proximal parts of the Adriatic continental passive margin, opposite the European margin formed by the Penninic and Helvetic-Dauphinois domains. The Adriatic margin is well recorded by the sedimentary successions in the Northern Calcareous Alps and the less deformed Southern Alps; the European margin by the Prealpine klippen, the metamorphic Briançonnais cover, and the better preserved Helvetic-Dauphinois sedimentary sequences.

Continental rifting was generated by simple shear mechanisms, probably with Europa as the upper plate (opposite to the Permian setting). The continent-to-ocean evolution is particularly complex. From some central and western Alpine ophiolites, the local exposure on the ocean floor of an exhumed and altered peridotitic basement (e.g., Aosta, Malenco and Platta areas) may be envisaged. This hypothesis is corroborated by ophicarbonate breccias and continental detritus deposited on top of mantle serpentinites, recalling modern exposures along ocean-continent transitions (Manatschal and Bernoulli, 1999). In this view, coherent continental remnants of the extremely thinned extensional upper plate may have been lost within the Tethyan ocean, as isolated allochthons and potential sources for the Austroalpine and Penninic continental nappes presently inserted between ophiolitic units. As previously seen, other ophiolitic units recall either fragments of normal oceanic lithosphere, or tectonic slices and olistoliths of oceanic suites inside dominantly turbiditic and other mass-flow deposits.

Restoration of the Tethyan ocean is a long and intriguing problem, mainly due to the occurrence in the central Alps of multiple ophiolitic units within the collisional zone. Indeed, the complex multilayer of the Alps may represent two or more oceanic branches, or may be merely the ultimate result of orogenic dispersal by polyphase folding and transposition. The Piedmont zone is the largest ophiolite in the Alps. It extends over most of the western Alps and reappears beyond the Ossola-Ticino window in the central (Malenco-Avers, Platta) and eastern Alps (Glockner, Rechnitz), below the eastern Austroalpine. Minor ophiolites, generally associated with flysch-type metasediments, are located at lower structural levels, mainly in the external Penninic domain from the north-western (Valais zone, Ossola-Ticino) to the central Alps (Grisons) and Engadine window. By classic kinematic inversion of the nappe stack, these ophiolitic units are thought to be derived, respectively, from the Piedmont (South-Penninic) ocean and a northern basin (North-Penninic), supposedly separated by the Briançonnais microcontinent. Alternative reconstructions include a single Jurassic ocean with ribbon continents and/or variously-sized extensional allochthons, or a younger development of the North-Penninic basin, supposedly opening during the closure of the Piedmont ocean.

Alpine orogeny

The Alpine orogeny began in the eastern Austroalpine and finally involved, step by step, the entire Alpine Tethys, gradually progressing from internal to external domains.

The earliest Alpine orogeny developed in the eastern Austroalpine and was accomplished before the deposition of the Late Cretaceous Gosau beds: it is tentatively related to the closure of a western branch of the Triassic Vardar ocean, possibly extending into the eastern Austroalpine domain through the Carpathians (Meliata ophiolite) and leading to a pre-Gosau continental collision. This reconstruction does account for the eclogitic (subduction) to Barrovian (collisional) metamorphism of Eoalpine (Early-Mid Cretaceous) age and wedge generation, although the oceanic suture is poorly documented and the axial trend of the Triassic ocean (oblique or transversal to the future Alpine belt) is uncertain.

The subsequent orogeny developed in the entire Alpine belt from the Late Cretaceous (western Austroalpine) onwards, and was closely related to the subduction of the Piedmont (South-Penninic) oceanic lithosphere below the Adriatic active continental margin, leading to Eocene collision between Europe and Adria. The first stage of Alpine contraction was dominated by a subduction-related low thermal regime which initiated with the onset of oceanic subduction (Mid Cretaceous ?): this is revealed by the oldest (Late Cretaceous) high-P peak in the western Austroalpine, and lasted until the Eocene syncollisional subduction of the proximal European margin, clearly recorded by the eclogitic to blueschist facies Penninic continental units. This stage was characterized by the growth of a pre-collisional to collisional (Austroalpine-Penninic) wedge at the Adria active margin. Since the beginning, it was devoid of a proper lithospheric mantle, being first underlain by the subducting oceanic lithosphere and, after ocean closure, by the passive margin of the European lower plate undergoing syn-collisional subduction and accretion. Wedge dynamics are enigmatic and are interpreted by: i) accretion of delaminated fragments of lithospheric microcontinents separated by oceanic channels; ii) tectonic erosion of the Adriatic active margin, inferred from the debated Cretaceous age of the subduction metamorphism also in the internal Penninic continental nappes; iii) accretion, by tectonic underplating, of originally thin crustal fragments resulting from an extensional upper plate (asymmetric rifting). In any case, exhumation of the high-P Penninic nappes was assisted by periodic extension in the wedge suprastructure, associated with nappe underplating at depth and active plate contraction.

From the late or latest Eocene (in differing areas), the cool, subduction-related regime was replaced by relaxed and perturbated thermal conditions. Indeed, the subduction complex was exhumed to shallower structural levels and overprinted by a Barrovian metamorphism of Late Eocene-Early Oligocene age (called Mesoalpine), characterized by a thermal gradient of 35 to 50°C/km (Frey et al., 1999). Soon after, a post-collisional magmatic cycle developed and was rapidly exhausted during the Oligocene (32–30 Ma). It is widely recorded along the Periadriatic fault system, from the lower Aosta valley to the eastern edge of the Alps (Bigi et al., 1990). Older magmatic products (42–38 Ma) only occur in the southern part of the Adamello batholith. The Periadriatic magmatism is represented by calc-alkaline to ultrapotassic plutons and dykes, which cut the northern part of the Southern Alps and the inner part of the Austroalpine-Penninic wedge. These bodies were generated by partial melting of lithospheric mantle sources previously modified during the Cretaceous-Eocene subduction. Generation and ascent of Periadriatic melts to upper crustal reservoirs were linked to slab break-off and related thermal perturbation, coupled with extension and rapid uplift of the wedge during active plate convergence.

The Periadriatic magmatism ceased in the Late Oligocene, when renewed collisional shortening disactivated the magmatic sources. Continuing plate convergence progressed externally, mainly through bilateral frontal accretion, coupled with vertical and horizontal extrusion and cooling of the Austroalpine-Penninic wedge. Indeed, segments of the foreland were accreted in front of the collisional wedge, as shown by the Helvetic basement slices and décollement cover nappes, displaced over the sinking Molasse foredeep. An opposite-vergent thrust-and-fold belt developed in the Southalpine upper crust, mainly generated by indentation of the Adria mid-lower lithosphere moving against the rear of the wedge. In the meantime, the overthickened Austroalpine-Penninic nappe stack underwent orogen-parallel tectonic denudation along low-angle detachments (e.g., Ratschbacher et al., 1991). Seismicity, GPS measurements and foreland subsidence give evidence of still active Adria/Europe convergence, with extensional and/or contractional tectonics in different sectors of the belt.

References

Bertotti, G., Picotti, V., Bernoulli, D. and Castellarin, A., 1993, From rifting to drifting: tectonic evolution of the South-Alpine upper crust from the Triassic to the Early Cretaceous: Sedimentary Geol., v. 86, pp. 53-76.

Bigi, G., Castellarin, A., Coli, M., Dal Piaz, G.V., Sartori, R., Scandone, P. and Vai, G.B., 1990, Structural Model of Italy, sheets 1-2: CNR, Progetto Fin. Geodinamica, SELCA Firenze.

Castellarin, A., Cantelli, L., Fesce, A. M., Mercier, J. L., Picotti, V., Pini, G. A., Prosser, G., Selli, L., 1992, Alpine compressional tectonics in the Southern Alps. Relationships with the Apennines. Annales Tectonicae 6, 62-94.

Coward, M.P., Dietrich, D., and Park, R.G., 1989, Alpine Tectonics: Geol. Soc. London Spec. Publ., no. 45, 449 pp.

Dal Piaz, G.V., 1999, The Austroalpine-Piedmont nappe stack and the puzzle of Alpine Tethys, *in* G. Gosso et al., eds, Third Meeting on Alpine Geol. Studies: Mem. Sci. Geol., v. 51, pp. 155-176.

Dal Piaz, G.V., 2001, History of tectonic interpretations of the Alps. J. Geodynamics, v. 32, pp. 99-114.

Flügel, H.W. and Faupl, P., eds, 1987, Geodynamics of the Eastern Alps: Deutikle, Vienna, 418 pp.

Frey, M., Desmons, J. and Neubauer, F., eds, 1999, The new metamorphic map of the Alps: Schweiz. mineral. petrogr. Mitt., v. 79, pp. 1-230.

Hunziker, J.C., Desmons, J., Hurford, A.J., 1992. Thirty-two years of geochronological work in the Central and Western Alps: a review on seven maps. Mem. Geol., Lausanne, v. 13, 59 pp.

Manatschal, G. and Bernoulli, D., 1999, Architecture and tectonic evolution of non-volcanic margins: present-day Galicia and ancient Adria. Tectonics, v. 18, pp. 1099-1119.

Michard, A., Goffé, B., Chopin, Ch., and Henry, C., 1996, Did the Western Alps develop through an Oman-type stage? Eclogae Geol. Helv., v. 89, pp. 43-80.

Neubauer, F. and Höck, V., eds, 2000, Aspects of Geology in Austria: Mitt. Oesterr. Geol. Ges., v. 92, 335 pp.

Pfiffner, O.A., Lehner, P., Heitzmann, P., Mueller, St. and Steck, A., eds, 1997, Deep structure of the Swiss Alps: Birkhäuser Verlag, Basel, 380 pp.

Plöchinger, B., 1995, Tectonics of the Northern Calcareous Alps: a review. Mem. Sci. Geol., v. 47, pp. 73-86.

Ratschbacher, L., Frisch, W., Linzer, H.G. & Merle, O., 1991, Lateral extrusion in the Eastern Alps, 2: structural analysis. Tectonics, v. 10, pp. 257-271.

Roure, F., Heitzmann, P. and Polino, R., eds, 1990, Deep structure of the Alps: Mém. Soc. géol. France, v. 156, 367 pp.

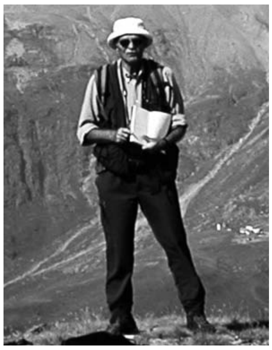
Transalp Working Group, 2002, First deep seismic images of the Eastern Alps reveal giant crustal wedges. Geophys. Res. Lett., v. 29, pp. 10.1029-10.1032.

Trümpy, R., 2001, Why plate tectonics was not invented in the Alps. Int. J. Earth Sci., v. 90, pp. 477-483.

Trümpy, R. and coworkers, 1980, Geology of Switzerland, a guide-book: Schweiz. Geol. Komm., Wepf & Co., Basel,. 334 pp.

Von Raumer, J. and Neubauer, F., eds, 1993, Pre-Mesozoic geology in the Alps: Springer-Verlag, 677 pp.

***Giorgio V. Dal Piaz**, full professor of geology in the Faculty of Sciences of the University of Padova, Italy, was born in 1935 and his research activity focuses on field survey, tectonics, subduction metamorphism and hard rock geology in the Alps.*



***Andrea Bistacchi** is PhD in Earth Sciences (University of Padova) and his work concentrates on post-nappe ductile to brittle tectonics in the Alps.*

***Matteo Massironi** is PhD in Space Science and Technology and Lecturer of geological mapping in the University of Padova. He mainly works on remote sensing and tectonics in the Alps.*

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

Structural setting of the Apennine-Maghrebian thrust belt

1 Dep. of Earth Sciences, University of Pisa, Via S. Maria 53, I-56126 Pisa, Italy.
2 Dep. of Geological Sciences, University of Catania, Corso Italia 55, I-95129 Catania, Italy.
3 Dep. of Geological Sciences, University "Roma Tre", Largo S. Leonardo Murialdo 1, I-00146 Rome, Italy.
4 Dep. of Earth Sciences, University of Torino, Via Accademia delle Scienze 5, I-10123 Torino, Italy.

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-

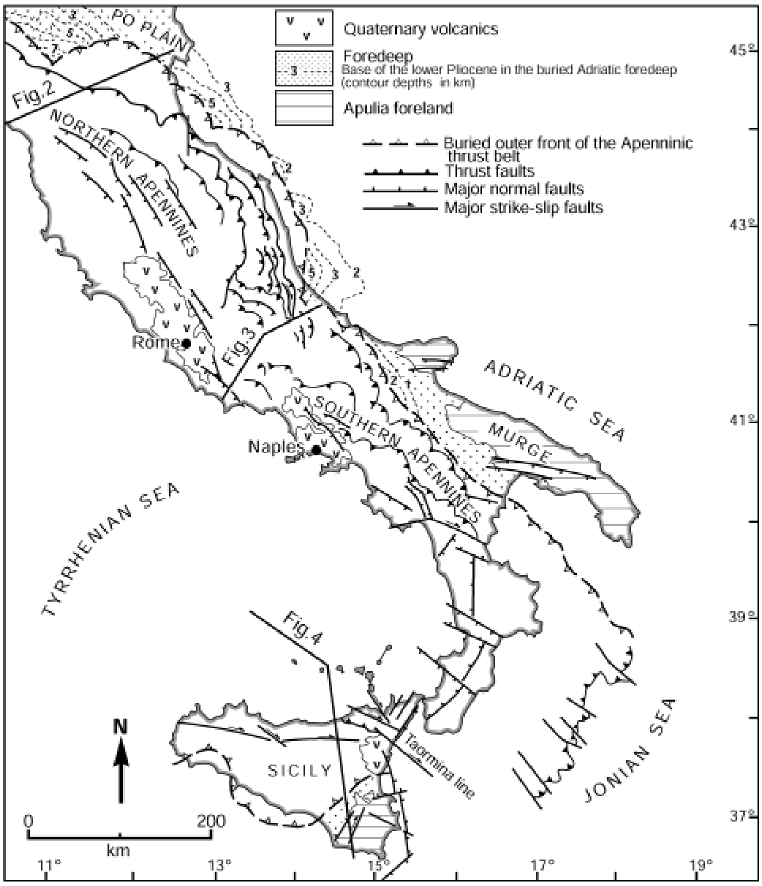


***Figure 1** General structural map of the Apennine-Maghrebian chain. Fig.2, Fig.3, Fig.4: locations of the cross-sections of Figures 2–4. (After Ghisetti and Vezzani, 1999, modified)*

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-

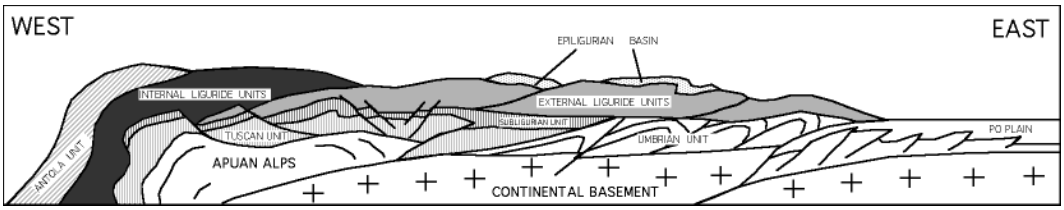


***Figure 2** Schematic cross-section showing the geometric relations among the major structural units of the northern Apennine. For location, see Figure 1. (P. Elter)*

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

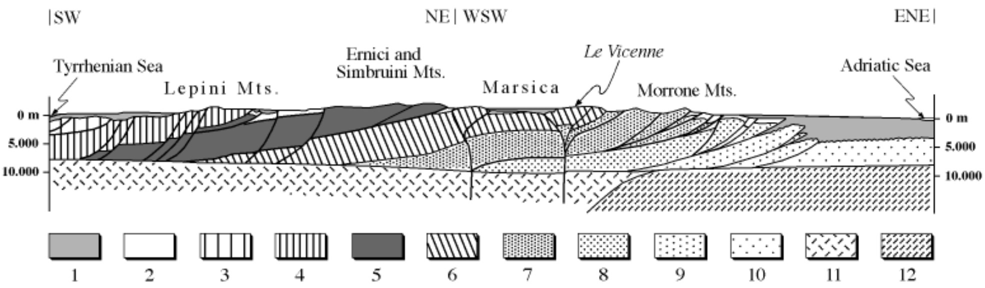


***Figure 3** Schematic cross-section of the central Apennine thrust belt. For location, see Figure 1. 1. marine and continental post-orogenic sedimentary cover (Tyrrhenian side) and intermountain basins (Late Messinian-Quaternary); marine syn- and post-orogenic deposits (Adriatic side: early Pliocene-Quaternary); 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 multi-channel 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 anti-clockwise 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.

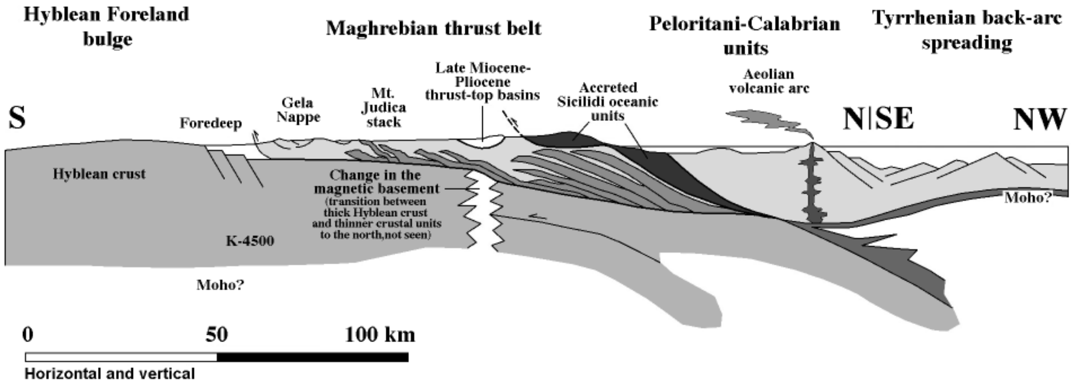


Figure 4 Simplified section across the eastern Sicily, from the Hyblean foreland to the southern Tyrrhenian back-arc basin; K indicates the magnetic susceptibility. For location, see Figure 1. (M. Grasso)

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).

Acknowledgement

Contributions by Michele Marroni, Boris Behncke and Giovanni Sturiale and the critical review of this paper by Antonio Praturlon and Daniela Di Bucci are gratefully acknowledged.

References

Bally, A.W., Burbi, L., Cooper C., and Ghelardoni, R., 1988, Balanced sections and seismic profiles across the central Apennine: Mem. Soc. Geol. It., v. 35, pp. 257-310.

Baruffini, L., Lottaroli, F., Torricelli, S., and Lazzari. D., 2000, Stratigraphic revision of the Eocene Albidona Formation in the type locality (Calabria, Southern Italy): Riv., Ital., Pal., Strat., v. 106, pp. 73-98.

Bonardi, G., Ciampo, G., and Perrone, V., 1985, La Formazione di Albidona nell'Appennino calabro-lucano: ulteriori dati stratigrafici e relazioni con le unità esterne appenniniche: Boll. Soc. Geol. It., v. 104, pp. 539-549.

Bonardi, G., Cavazza, W., Perrone, V., and Rossi, S., 2001, Calabria-Peloritani Terrane and Northern Ionian Sea, *in* Vai, G.B., and Martini, I.P., eds, Anatomy of an Orogen: the Apennines and adjacent Mediterranean Basins, Kluwer Academic Publishers, pp. 255-286.

Butler, W.H.R., and Grasso, M., 1993, Tectonic controls on base-level variation and depositional sequences within thrust-top and foredeep basins: examples from the Neogene thrust belt of central Sicily: Basin Research, v. 5, pp. 137-151.

Butler, W.H.R., Grasso, M., and La Manna, F., 1992, Origin and deformation of the Neogene-Recent Maghrebian foredeep at Gela Nappe, SE Sicily: Journ. Geol. Soc. of London, v. 149, pp. 547-556.

Butler, W.H.R., Lickorish, W.H., Grasso, M., Pedley, M., and Ramberti, L., 1995, Tectonics and sequence stratigraphy in Messinian basins, Sicily: constraints on the initiation and termination of the Mediterranean salinity crisis: Geological Society of America Bulletin, v. 107, pp. 425-439.

Cipollari, P., and Cosentino, D., 1992: La linea Olevano-Antrodoco: contributo della biostratigrafia alla sua caratterizzazione cinematica: Studi Geologici Camerti, spec. v. 1991/2, pp. 143-149.

Cipollari, P., Cosentino, D., and Parotto, M., 1995, Modello cinematico-strutturale dell'Italia centrale: Studi Geologici Camerti, spec. v. 1995/2, pp. 135-143.

Cipollari, P., Cosentino, D., Esu, D., Girotti, O., Gliozzi, E., and Praturlon, A., 1999, Thrust-top lacustrine-lagoonal basin development in accretionary wedges: late Messinian (Lago-Mare) episode in the central Apennines (Italy): Paleogeo., Paleoclim., Paleoecol., v. 151, pp. 149-166.

Decima, A., and Wezel, F., 1971, Osservazione sulle evaporiti messiniane della Sicilia centro-meridionale: Rivista Mineraria Siciliana, v. 130-132, pp. 172-187.

Di Bucci, D., and Mazzoli, S., 2002, Active tectonics of the Northern Apennines and Adria geodynamics: new data and discussion: Journ. Geodynamics, v. 34, pp. 687-707.

Finetti, I, 1982, Structure, stratigraphy and evolution of Central Mediterranean: Boll. Geof. Teor. Appl., v. 24, no 96, pp. 274-315.

Finetti, I, and Del Ben, A., 1986, Geophysical study of the Tyrrhenian opening: Boll. Geof. Teor. Appl., v. 28, no 110, pp. 75-155.

Ghisetti, F., and Vezzani, L., 1991, Thrust belt development in the central Apennines: northward polarity of thrusting and out-of-sequence deformations in the Gran Sasso chain (Italy): Tectonics, v. 10, pp. 904-919.

Ghisetti, F., and Vezzani, L., 1999, Depth and modes of Pliocene-Pleistocene crustal extension of the Apennines (Italy): Terra Nova, v. 11, pp. 67-72.

Mostardini, F., and Merlini, S., 1986, Appennino centro-meridionale: sezioni geologiche e proposta di modello strutturale: Mem. Soc. Geol. It., v. 35, pp. 342-361.

Ori, G.G., and Friend, P.F., 1984, Sedimentary basins formed and carried piggyback on active thrust sheets:. Geology, v. 12, pp. 475-478.

Parotto, M., and Praturlon, A., 1975, Geological summary of the Central Apennines: Quad. Ric. Sci., v. 90, pp. 257-311.

Patacca, E., Sartori, R., and Scandone P., 1990, Tyrrhenian basin and apenninic arcs: kinematic relations since Late Tortonian times. Mem. Soc. Geol. It., v. 45, pp. 425-451.

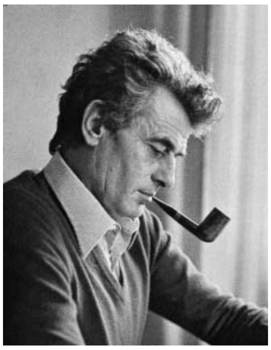
Ricci Lucchi, F., 1986, The Oligocene to Recent foreland basins of the northern Apennines, *in* Allen, P.A., and Homewood, P., eds, Foreland Basins: I.A.S. Spec. Publ., v. 8, pp. 105-139.

Royden, L., Patacca, E., and Scandone, P., 1987, Segmentation and configuration of subducted lithosphere in Italy: an important control on thrust-belt and foredeep-basin evolution: Geology, v. 15, pp. 714-717.

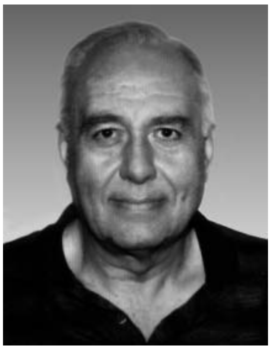
Sestini, G., ed, 1970, Development of the Northern Apennines geosyncline: Sedimentary Geol., v. 4, no 3/4, pp. 1-637 pp.

Vezzani, L.,1970, Il Flysch di Albidona nell'area del confine tra Calabria e Lucania: Geol. Rom., v. 9, pp. 101-126.

***Piero Elter**, born in 1927, retired Professor of Geology at the University of Pisa. He received the Degree in Geology at University of Geneva in 1954. Piero Elter was Director of the Centro di Geologia Strutturale e Dinamica dell'Appennino of the CNR (Italian National Council of Research) from 1972 to 1988. His main research interests have been focused on the tectonic evolution of the Alps and the Northern Apennine. In particular, he has investigated the ophiolites from Northern Apennine and the implications for the origin of the oceanic lithosphere in the western Tethys.*

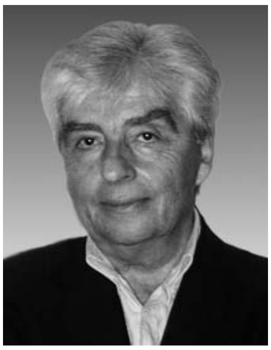


***Maurizio Parotto** is teaching Introduction to Geodynamics and Historical Geology at University "Roma Tre" (Italy). His research concentrates on stratigraphy and structural setting of central Apennine; at present he is Director of sub-project CROP 11 (CROsta Profonda, Deep Crust), a part of a CNR (Italian National Council of Research) project which involves the integration of crustal NVR seismic profiles with surface and subsurface geology in central Italy, from Tyrrhenian Sea to Adriatic Sea.*



***Mario Grasso** is Professor of Structural Geology at Catania University. Research expertise: stratigraphy, synsedimentary tectonics and geomorphological evolution of the Sicily-Calabrian region, regional field mapping, crustal structure of the Mediterranean region. He is leader of several National scientific projects concerning geological mapping and environmental risk assessment and member of the National Geological Committee.*

***Livio Vezzani** is Professor of Geology at Torino University. Expertise: Stratigraphic and structural analyses of fold-and-thrust belts, field mapping, editing and compilation of geological and regional tectonic maps, regional geology of the Mediterranean region, tectonic geomorphology, Quaternary geology, neotectonics, seismotectonics and seismic hazard assessment. Research activity has been ficused on the geodynamic evolution of the central and southern Apennine chain, Calabrian Arc and Sicily.*



by Carlo Bartolini[1], Nicola D'Agostino[2], and Francesco Dramis[3]

Topography, exhumation, and drainage network evolution of the Apennines

1 Department of Earth Sciences, University of Florence, Via Giorgio La Pira, 4 - 50121 Florence, Italy.
2 INGV - National Institute of Geophysics and Volcanology, Via di Vigna Murata, 605 - 00143 Rome, Italy.
3 Department of Geological Sciences, "Roma Tre" University, Largo San Leonardo Murialdo, 1 - 00146 Rome, Italy.

The present-day topography of the Italian peninsula results from the interactions between crustal-mantle and surface processes occurring since the Late Miocene. Analysis of exhumation and cooling of crustal rocks, together with Quaternary drainage evolution, helps to unravel the tectonic-morphologic evolution of the Apennines by distinguishing end-member models, and hence describing the orogenic belt evolution. The pattern of regional topography, erosional history and present-day distribution of active deformation suggests that the eastward migrating extensional-compressional paired deformation belts may still control the topography of the northern Apennines, albeit at slower rates than in the past. Conversely, Quaternary drainage evolution in the central and southern Apennines suggests that the topography of these regions underwent a Quaternary regional arching, which is only partly consistent with the persisting migration of the compressional-extensional pair.

Introduction

Structural, sedimentological and volcanological observations show that the Neogene-Quaternary geological history of the Apennines was dominated by the coexistence of paired, eastward-migrating extensional and compressional deformational belts (Elter et al., 1975) driven by the passive sinking rollback of the Adriatic-Ionian lithosphere (Malinverno and Ryan, 1986). It is now widely recognized that interactions between surface and crustal processes is a first-order factor controlling the evolution of orogenic belts (Zeitler et al., 2001). Detailing the changes in the morphology and the associated surface processes acting in these belts throughout their evolution is thus important for understanding the underlying driving mechanisms. Exhumation and rock uplift rates, changes in drainage patterns and depositional environments constrain conceptual geodynamical models and can be used to quantitatively test numerical models.

Two end-member views have been outlined in the literature for the topographic evolution of the Apennines:

i) In the first, the Apennines have existed as a self-similar, eastward-migrating topographic high throughout Neogene-Quaternary times. The present morphology thus represents the present-day expression of a continuous process. This concept implies the view that crustal shortening to the east is still active all along the Apennines and kinematically related to back-arc crustal extension to the west of the drainage divide (Cavinato and DeCelles, 1999). This view has strongly influenced geomorphologic studies and is deeply rooted in the literature (e.g. Mazzanti and Trevisan, 1978; Alvarez, 1999).

ii) An alternative hypothesis views the Apennine topography as decoupled from the tectonic deformation of the Apennine orogenic wedge. This view is supported by various structural, geophysical and morphological lines of evidence, such as the significant enlargement and arching of the Italian peninsula, the free-air gravity anomaly spectral signature and the lack of significant crustal shortening active east of the topographic high, particularly in the central and southern Apennines (Carminati et al., 1999; D'Agostino et al., 2001).

The conceptual models outlined above predict very different patterns of erosion and drainage evolution through time which allows us to focus here on two main topics. The first is the distribution and intensity of erosion rates in the northern Apennines over the Neogene-Quaternary time-scale, as determined by thermochronological data, while the second focuses on the evolution of the Quaternary intramontane basins and their associated drainage network.

Tectonic setting

The Apennines are formed by a Neogene thrust and fold belt system which developed during the eastward retreat of west-dipping subduction of the Adriatic-Ionian lithosphere. Foredeep subsidence and thrust stacking in the eastern external parts coexisted with western back-arc extension and volcanism (Serri et al., 1993; Jolivet et al., 1998). Sedimentary units originally deposited on the Adriatic lithosphere were progressively accreted to the Apennines wedge (Patacca et al., 1992).

The Early-Middle Pleistocene alluvial and transitional deposits filling the Adriatic foredeep have been regionally tilted to the NE (Kruse and Royden, 1994) and incised by a parallel drainage system. The central and southern Apennines (south of 43˚N) currently show little evidence of active compression and the tectonic regime is instead dominated by active NE-SW crustal extension. Compressive focal mechanisms and shortening are still active in the external parts of the northern Apennines (Frepoli and Amato, 1997) where intermediate earthquakes, down to depths of 90 km, may suggest a still active subduction (Selvaggi and Amato, 1992). Historical and recent seismicity, active faulting (Valensise et al., this volume), and geodesy (Hunstad et al., 2003) indicate that 3–5 mm/yr of extension may be accommodated across the central-southern Apennines in a 30–50 km wide belt. This belt of active extension closely follows the regional topographic culmination close to the drainage divide. The relative positions of drainage divide, highest elevations and the eastern front of normal faulting vary along the Apennines (Figure 1). In the northern Apennines, the striking correspondence of these three elements may suggest that the topography is controlled by an eastward-migrating wave of normal faulting, lowering the west side of the orogenic belt (Mazzanti and Trevisan, 1978). In the central and southern Apennines, the drainage divide is localized along the crest of the regional topographic culmination, while the highest peaks are probably controlled by selective erosion in the southern (Amato et al., 1995) or advancement of the extensional front in the central Apennines (D'Agostino et al., 2001).

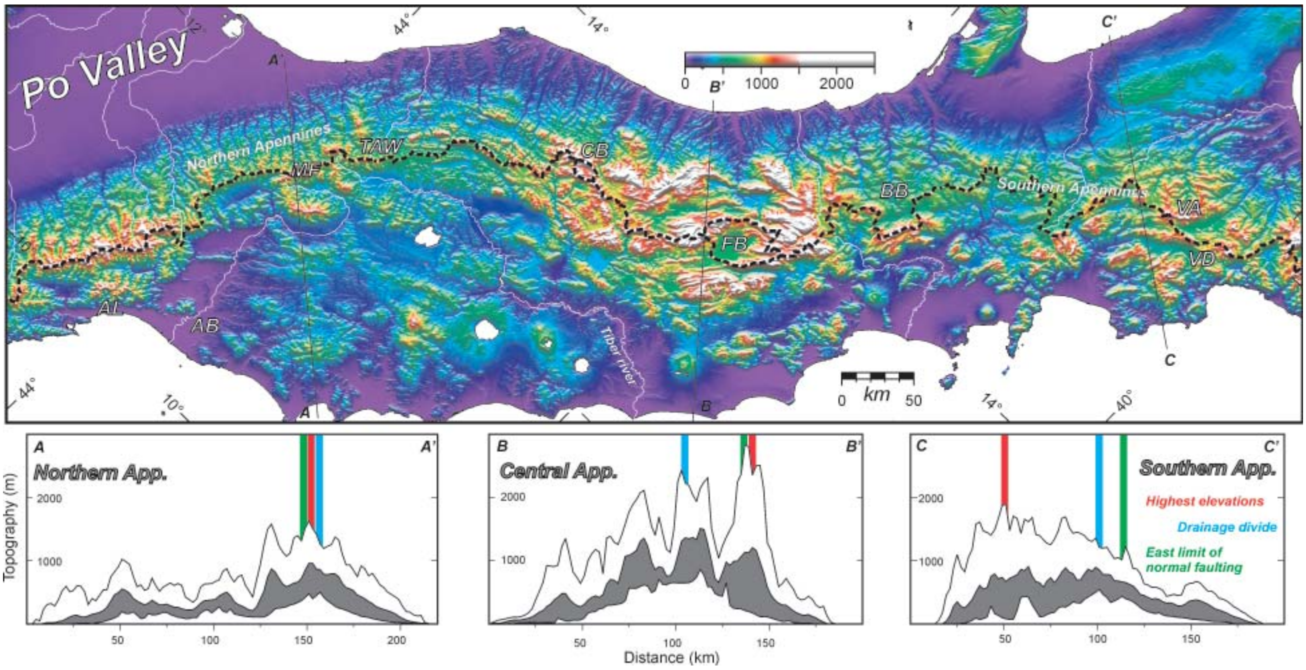


Figure 1 Shaded topography of the Italian peninsula (Oblique Mercator Projection) with swath profiles across northern, central and southern Apennines. Upper, median, and lower lines in the profiles correspond to maximum, average, and minimum elevation perpendicular to a 40-km swath.
Legend: AL, Apuane Alps; AB, Arno basin; MF, Monte Falterona; CB, Colfiorito basin; FB, Fucino basin; BB, Boiano basin; VA, Val D'Agri basin; VD, Vallo di Diano; TAW, Tyrrhenian-Adriatic watershed.

Thermochronological data and exhumation history

Low temperature thermochronology is ideally suited for reconstructing the thermal history of rocks in the uppermost part of the crust because it records time and rates of cooling related to rock exhumation. Measurements are mostly carried out on apatite fission-tracks (AFT), since they yield times and rates at which rocks cooled below approximately the 110˚C isotherm. Since the geothermal gradient is usually poorly constrained, exhumation rates are commonly affected by significant errors. Whenever the elevation-dependence method can be adopted, whereby several samples from different elevations along a transect are analysed, the geothermal gradient has not to be introduced in order to obtain the exhumation rate. The (U-Th)/He analyses on apatite, being based on a closure temperature of ca. 70˚C, allow a closer view of the latest stages of exhumation. Unlike the geological evolution of the Apennines, which is fairly well known, its geomorphic history has only recently aroused widespread scientific interest, also because of the availability of thermochronological data. A preliminary study on exhumation rates in the northern Apennines, based on apatite fission tracks, was published by Boettcher and McBride (1993). Since then, the regional knowledge of exhumation rates over different time spans has been increasing at a fast pace (Abbate et al., 1994; Balestrieri et al., 1996, 2003; Ventura et al., 2001; Zattin et al., 2002).

On a regional scale, AFT data indicate that higher exhumation rates are at present occurring at Mt. Falterona, over the drainage divide of the northern Apennines, consisting of Miocene foredeep deposits (Marnoso Arenacea Fm.). According to Zattin et al. (2002), the pre-exhumation configuration features a 4- to 5-km-thick cover (depending whether a geothermal gradient of 20˚C/km or 25˚C/km is assumed) of overlying Ligurian Units and Epiligurian Units, which was completely eroded in the last 5 Ma at a mean rate, then, of 0.8 to 1 mm/yr. A thickness of 3.8 km was calculated by Reutter et al. (1983) in the same Formation approximately at the same site and for the same chronological interval, by working out vitrinite reflectance data. A residual veneer of Ligurids, presently buried under the fluvial deposits both in Mugello and in the nearby Casentino basin indicates that the 5-km-thick Ligurids cover had not been completely unroofed when the basin became the site of flood plain sedimentation, that is around 2.0 Ma (Benvenuti, 1997).

AFT analyses, carried out west of the drainage divide, gave Pliocene-Quaternary mean exhumation rates ranging from 0.5 to 1.7 mm/yr across the Apuane Alps (Abbate et al., 1994) and Late Miocene-Quaternary mean denudation rates of 0.3–0.4 mm/yr in the Ligurian Apennines (Balestrieri et al., 1996). It should be pointed out that higher rates in both areas derive from cooling rates assuming a geothermal gradient of 30˚C/km. Lower rates (0.5 mm/yr and a minimum of 0.2 mm/yr, respectively) derive directly from the slope of the age-elevation profiles. A mean exhumation rate of 0.2–0.3 mm/yr in the Ligurian Apennines since Late Miocene and of 0.5 mm/a in the Apuane Alps since Early Pliocene can be held as reference values. Further recent investigations in the Apuane Alps area (Balestrieri et al., 2003) allow a more detailed time span of exhumation rates to be worked out. Between ca. 11 Ma and ca. 6 Ma, the cooling rate was between 10 and 16˚C/Ma which corresponds, assuming a geothermal gradient of 25˚C/km (Pasquale et al., 1997), to exhumation rates of 0.4–0.6 mm/yr. Between 6 Ma and 4 Ma, cooling rates increased to between 38 and 55˚C/Ma. equivalent to an exhumation rate of 1.3–1.8 mm/yr, assuming a geothermal gradient of 30˚C/km. The last part of the thermal path (4 Ma to present) is not well constrained due to lack of AHe data, but the average exhumation rate is between 0.6 and 0.9 mm/yr (assuming a geothermal gradient of 30˚C/km). The slowing down of the exhumation rate since Middle Pliocene is certainly related to the surface exposure of the highly resistant rocks belonging to the Tuscan Metamorphic Unit and Paleozoic Basement which began to occur at that time, as proven by the lithologic composition of the continental basin fed to the east by the Apuane Alps erosion (Calistri, 1974). Despite the lower exhumation rates, the present average altitude of the Apuane peaks (1500 to 1900 m) is very close to that of the Apennine divide. The local relief of the Apuane Alps reflects the greater resistance of the underlying rock types which contrast so strongly in form with the adjacent Apennine range that the Apuane Alps have been considered a separate range of a possibly different origin (Wezel, 1985).

In summary, thermochronological data show that the long-term exhumation of the northern Apennines has occurred at an average rate of 0.7 mm/yr since 11 Ma. This value is remarkably close to the average erosion rate evaluated from the sediment volume deposited in the Adriatic foredeep (Bartolini et al., 1996). Local accelerations of exhumation rates coincide with the onset of normal faulting and intramontane basin deposition through the creation of local relief and accommodation space for sediments (Balestrieri et al., 2003). Acceleration of exhumation rates in the Mt Falterona area took place about 3 Ma later than the onset of similar exhumation in the Apuane Alps. This acceleration and migration of the locus of highest erosion rate is consistent with an eastward migrating pulse of normal faulting and enhanced local relief creation.

Evolution of the drainage network and the "watershed" basins

The present-day shape of the Italian peninsula is closely linked with the recent formation of a long-wavelength topography responsible for raised Pleistocene marine sediments and terraces on both flanks of the range (Bordoni and Valensise, 1998; D'Agostino et al., 2001). This morphological evolution is associated with a major change of the depositional environments in the intramontane basins: Upper Pliocene to Middle Pleistocene fluvial-lacustrine environments changed to Middle to Upper Pleistocene fluvial-alluvial sequences in a regionally correlated phase of basin fill incision and drainage integration (Bartolini and Pranzini, 1981; D'Agostino et al., 2001). The Early Pleistocene is characterized by lacustrine environments in most of the intramontane depressions, recorded by widespread lake beds that are revealed within the incised basin fills (Bosi and Messina, 1991; Bartolini, 2003). The Early-Middle Pleistocene lacustrine deposits are generally overlain by units that are transitional from lacustrine and low-gradient fluvial environments to coarser deposits representative of alluvial fans (Miccadei et al., 1998). This transition is frequently marked by erosion and incision of the lake beds so that the Middle Pleistocene deposits are often entrenched and unconformably overlie the fluvial-lacustrine units. After the Middle Pleistocene, deposition of lacustrine sediments in the intramontane basins was drastically reduced and continued only in basins that maintained internal drainage. Over the Quaternary as a whole, but with slightly different timing, this evolution is typical of the intramontane basins of the northern (Argnani et al., 1997) and southern Apennines (Capaldi et al., 1988). The only surviving closed (Fucino and Colfiorito) and semi-closed (Boiano, Vallo di Diano, Val d'Agri basins) basins are those on the Apennine watershed, most distant from the marine base level, where continued normal faulting is still able to provide local subsidence that defeats their capture by the regional drainage network. These basins generally contain aggradational sequences up to 1000 m thick, made up of poorly exposed fluvial and lacustrine Pleistocene sediments drilled by boreholes or imaged by seismic reflection lines (Cavinato et al., 2002). These basins are bounded by normal fault systems, displaying evidence of Late Quaternary-Holocene faulting events, and their historical seismicity includes M>6 seismic events along basin-bounding faults (Valensise et al., this volume). In contrast to the fluvially-dissected landscapes and incised continental Pleistocene sequences that are common farther from the drainage divide, these "watershed" basins are characterized by flat and weakly incised lacustrine depositional surfaces, alluvial plain and fan surfaces that suggest an incomplete integration with the regional drainage network. Historical drainage operations, aimed at increasing agricultural land, artificially lowered basin outlets or excavated artificial galleries, causing the disappearance of seasonal or perennial lakes such as the shallow but large Fucino lake, finally drained in 1875 AD.

Detailed geological and geophysical investigations in the Fucino basin document (Figure 2) the persistent Plio-Quaternary lacustrine environments and the asymmetrical geometry of the basin

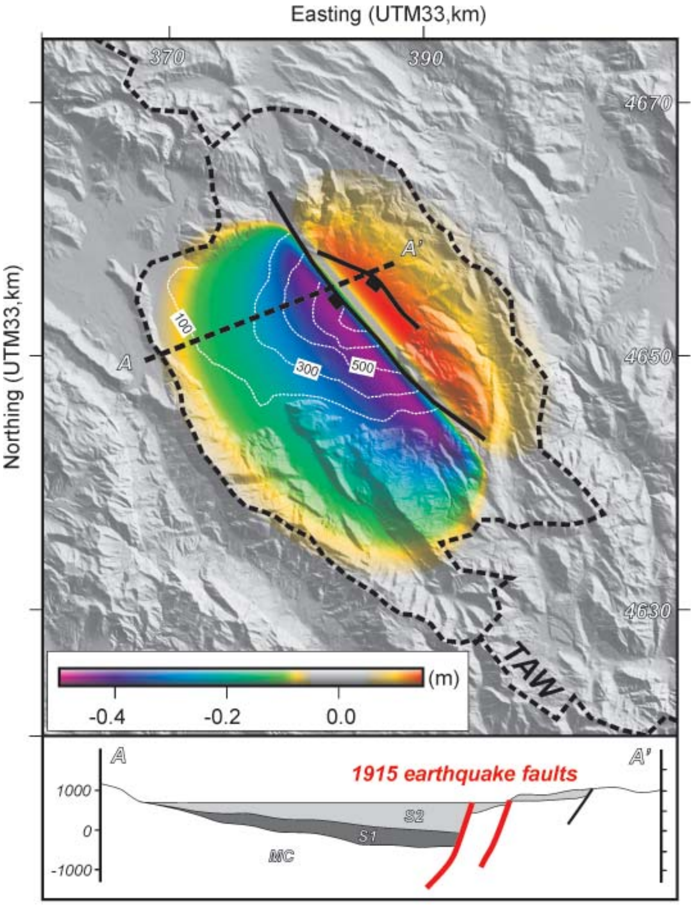


***Figure 2** Fucino basin schematic map. The color scale corresponds to the vertical coseismic deformation of the 1915 Ms 6.9 earthquake (fault parameters from Ward and Valensise, 1989). Main Quaternary basin-bounding faults are shown as black lines. Thin dashed white lines indicate isochron contours (in ms TWT) of the Plio-Quaternary basin fill (from Cavinato et al., 2002). TAW, Tyrrhenian-Adriatic watershed. Also shown is the schematic cross-section across the basin (redrawn from Cavinato et al., 2002); MC, Meso-Cenozoic rocks; S1, Upper Pliocene-Early Pleistocene; S2, Early Pleistocene-Holocene.*

fill thickening toward the master NW trending fault systems that ruptured during the 1915 Ms = 6.9 Fucino earthquake. The geometry of the basin fill shows that the fault segment which ruptured during that earthquake may have accumulated a vertical offset larger than 1000 m by repeated 1915-like seismic events. The remarkable correspondence between the distribution of the 1915 coseismic vertical deformation and the extent of internal drainage provides a striking example of the struggle of intramontane basin faults to preserve internal drainage in the "watershed" basins against integration by regional drainage network by repeated seismic events that "dam" externally flowing rivers or by providing local subsidence not compensated by basin aggradation.

In most places today the effects of downcutting by the regional network dominate over the local subsidence caused by slip on the active faults. For this reason, except in some of the higher and more distant parts of the regional network, the more subtle geomorphological effects of fault activity are usually obscured.

The evolution from an internally-drained system to a through-going river network was ultimately related to the development of a long-wavelength topographic bulge and regional uplift. D'Agostino et al. (2001) used the spectral ratio (admittance) of free-air gravity anomalies to topography to show that long-wavelength topography is supported by mantle dynamics, which is consistent with geological evidence showing the absence of significant Quaternary crustal thickening or underplating. They argued that the Quaternary evolution of the Italian peninsula has been controlled by the formation of a NW-SE long-wavelength (> 150 km) topographic bulge. In most places today the effects of downcutting by the regional network dominate over the local subsidence caused by slip on the active faults. For this reason, except in some of the higher and more distant parts of the regional network, the more subtle geomorphological effects of fault growth and interaction are usually obscured. The surviving closed or semi-closed basins along the divide allow us to see the interaction between the regional downcutting and normal faulting. The preservation of these basins on the watershed is related to two factors:

i) They are both far from the base levels of the streams that are trying to capture them. The wave of regressive erosion triggered by the base level fall has effected these internal regions later than those basins closer to the coasts, such as the Tiber, Sulmona, and Campo Imperatore basins. Once captured, the increased discharge through these well-integrated basins makes it likely that fluvial incision rates at their outlets will always dominate over local vertical motions related to faulting.
ii) Normal faulting is most active along the crest of the topographic bulge. Only in this location can faulting efficiently delay the capture of closed basins by causing subsidence that preserves internal drainage.

The Apennines provide an ideal place for observing this type of interaction because of their short distance from the marine base level, so that effects of base-level variations are rapidly propagated upstream, and because the active faulting is localized on top of the long-wavelength bulge.

Discussion

The consistency between various geological, geomorphologic and geophysical data sets suggests that a still active extensional-compressional migrating belt may control the topography in the northern Apennines. Seismological and geodetic evidence, together with tomographic images of a mantle low-velocity zone (Lucente et al., 1999), suggest that crustal deformation may be driven by the passive sinking of the Adriatic lithosphere. This driving mechanism is evidenced in the geomorphology through the eastward migration of significant exhumation rates and the close correspondence of highest elevations, drainage divide and eastern front of normal faulting.

Geological reconstructions provide rates of eastward Neogene migration of volcanism and crustal deformation of about 1–2 cm/yr (Jolivet et al., 1998). The lack of evidence for such high values of present-day rates of deformation may suggest that the whole geodynamic processes is currently slowing down.

In the central-southern Apennines, the internally-drained Pleistocene drainage system evolves to a through-going river network in association with the whole arching of the Italian peninsula contemporaneously with a significant slowing down of foredeep subsidence and compressional deformation in the Adriatic side. Uplift of marine deposits and incising drainage is symmetrical to both sides of the range. Active extension is now focused along the top of the regional topography and watershed where continuing normal faulting and basin subsidence “shield” intermontane basins form integration in the regional drainage network. This evolutionary pattern seems to be scarcely consistent with the continuing migration of the extensional-compressional belt, and is instead more likely related to a late-stage phase of the Adriatic lithosphere subduction.

References

Abbate, E., Balestrieri, M.L., Bigazzi, G, Norelli, P. and Quercioli, C., 1994, Fission track datings and recent rapid denudation in Northern Apennines, Italy: Memorie della Società Geologica Italiana, v. 48, pp. 579-585.

Alvarez, W., 1999, Drainage on evolving fold-thrust belts: a study of transverse canyons in the Apennines, Basin Research: v. 11. no. 3, pp. 267-284.

Amato, A., Cinque, A. and Santangelo, N., 1995, Il controllo della struttura e della tettonica plio-quaternaria sull’evoluzione del reticolo idrografico dell’Appennino meridionale: Studi Geologici Camerti, vol. spec., 2, pp. 23-30.

Argnani, A., Bernini, M., Di Dio, G.M., Papani, G. and Rogledi, S., 1997, Stratigraphic record of crustal-scale tectonics in the Quaternary of the Northern Apennines (Italy): Il Quaternario, v. 10, no. 2, pp. 595-602.

Balestrieri, M.L., Abbate, E. and Bigazzi, G., 1996, Insights on the thermal evolution of the Ligurian Apennines (Italy) through fission-track analysis: Journal of the Geological Society of London, v. 153, pp. 419-425.

Balestrieri, M.L., Bernet, M., Brandon, M. T., Picotti, V., Reiners, P. and Zattin M., 2003, Pliocene and Pleistocene exhumation and uplift of two key areas of the Northern Apennines: Quaternary International, pp. 67-73.

Bartolini, C., 2003, When did the Northern Apennine become a mountain chain?: Quaternary International, v. 101-102, pp. 75-80.

Bartolini, C., Caputo, R. and Pieri, M. 1996, Pliocene–Quaternary sedimentation in the Northern Apennine Foredeep and related denudation: Geological Magazine, v. 133, pp. 255-273.

Bartolini, C. and Pranzini, G., 1981, Plio-Quaternary evolution of the Arno basin drainage: Zeitschrift für Geomorphologie N.F., Suppl. Bd., v. 40, pp. 77-91.

Benvenuti, M., 1997, Physical stratigraphy of the fluvio-lacustrine Mugello Basin (Plio-Pleistocene, Northern Apennines): Giornale di Geologia, s. 3, v. 59, pp. 91-111.

Boettcher, S.S. and McBride, E.F., 1993, Thermal histories of piggy-back and foreland basins in the northern Apennines, Italy, derived from apatite fission track thermochronology: EOS Transactions, v. 74, no. 43, p. 547.

Bordoni, P. and Valensise, G., 1998, Deformation of the 125ka marine terrace in Italy: Tectonic implications, in Coastal Tectonics, eds. Stewart, I.S. and Vita-Finzi, C., Geological Society, London, Special Publications, 146, pp. 71-110.

Bosi, C. and Messina, P., 1991, Ipotesi di correlazione fra successioni morfo-litostratigrafiche plio-pleistoceniche nell’Appennino laziale-abruzzese: Studi Geologici Camerti, vol. spec. CROP 11, pp. 257-263.

Calistri, M., 1974, Studi di geomorfologia e neotettonica: Il Pliocene fluvio-lacustre della conca di Barga: Memorie della Società Geologica Italiana, v. 13, pp. 1-21.

Capaldi, G., Cinque, A. and Romano, P., 1988, Ricostruzione di sequenze morfoevolutive nei Picentini Meridionali (Campania, Appennino Meridionale), Geografia Fisica e Dinamica Quaternaria: v. 1, pp. 207-222.

Carminati, E., Giunchi, C., Argnani, A., Sabadini, R. and Fernandez, M., 1999, Plio-Quaternary vertical motion of the Northern Apennines: Insights from dynamic modelling: Tectonics, v. 18, no. 4, pp. 703-718.

Cavinato, G.P., Carusi, C., Dall’Asta, M., Miccadei, E., and Piacentini, T., 2002, Sedimentary and tectonic evolution of Plio-Pleistocene alluvial and lacustrine deposits of Fucino Basin (central Italy): Sedimentary Geology, v. 148, pp. 29-59.

Cavinato, G.P., and DeCelles, P.G., 1999, Extensional basins in the tectonically bimodal central Apennines fold-thrust belt, Italy: Response to corner flow above a subducting slab in retrograde motion: Geology, v. 27, pp. 955-958.

D’Agostino, N., Jackson, J., Dramis, and F., Funiciello, R., 2001, Interactions between mantle upwelling, drainage evolution and active normal faulting: an example from the central Apennines (Italy): Geophysical Journal International, v. 147, pp. 475-497.

Elter, P., Giglia, G., Tongiorgi, M. and Trevisan, L., 1975, Tensional and compressional areas in recent (Tortonian to Present) evolution of the Northern Apennines: Bollettino di Geofisica Teorica e Applicata, v. 17, pp. 3-18.

Frepoli, A., and Amato, A., 1997, Contemporaneous extension and compression in the northern Apennines from earthquake fault plane solutions: Geophysical Journal International, v. 129, pp. 368-388.

Hunstad, I., Selvaggi, G., D'Agostino, N., England, P., Clarke, P. and Pierozzi, M., 2003, Geodetic strain in peninsular Italy between 1875 and 2001: Geophysical Research Letters, v. 30, no. 4, pp. 1181-1184.

Jolivet, L., Faccenna, C., Brunet, C., Cadet, J. P. Funiciello, R., Mattei, M., Rossetti, F., Storti, F., D’Agostino, N., Goffè, B., Parra, T. and Theye, T., 1998, Midcrustal shear zones in postorogenic extension: the Northern Tyrrhenian Sea case: Journal of Geophysical Research, v. 103, pp. 12123-12160.

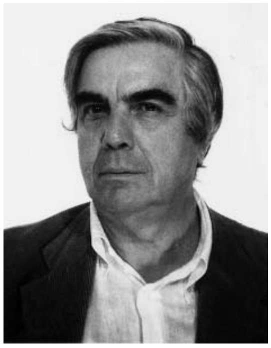
Kruse, S. and Royden, L.H., 1994, Bending and unbending of an elastic lithosphere: The Cenozoic history of the Apennine and Dinaride foredeep basins: Tectonics, v. 13, pp. 278-302.

Lucente, F.P., Chiarabba, C., Cimini, G.B. and Giardini, D., 1999, Tomographic constraints on the geodynamic evolution of the Italian region: Journal of Geophysical Research, v. 104, pp. 20307-20327.

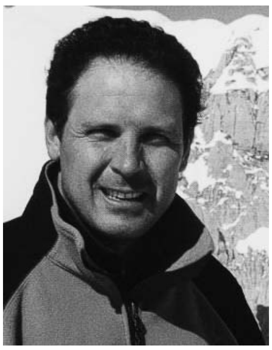
Malinverno, A., and Ryan, W.B.F., 1986, Extension in the Tyrrhenian Sea and shortening in the Apennines as results of arc migration driven by sinking of the lithosphere: Tectonics, v. 5, pp. 227-245.
Mazzanti, R. and Trevisan, L., 1978, Evoluzione della rete idrografica nell'Appennino centro-settentrionale: Geografia Fisica e Dinamica Quaternaria, v. 1, pp. 55-62.
Miccadei, E., Barbieri, R. and Cavinato, G.P., 1998, La geologia quaternaria della conca di Sulmona (Abruzzo, Italia centrale): Geologica Romana, v. 34, pp. 59-86.
Pasquale, V., Verdoya, M., Chiozzi, P. and Ranalli, G., 1997, Rheology and seismotectonic regime in the northern central Mediterranean: Tectonophysics, v. 270, pp. 167-195.
Patacca, E., Sartori, R., and Scandone, P., 1992, Tyrrhenian basin and Apenninic arcs: Kinematics relations since late Tortonian times: Memorie della Società Geologica Italiana, v. 45, pp. 425-451.
Reutter, K., Teichmuller, M., Teichmuller, R. and Zanzucchi, G., 1983, The coalification pattern in the Northern Apennines and its paleo geothermic and tectonic significance: Geologische Rundschau, v. 72, no. 3, pp. 861-894.
Selvaggi, G. and Amato, A., 1992, Intermediate-depth earthquakes in the Northern Apennines (Italy): evidence for a still active subduction?: Geophysical Research Letters, v. 19, pp. 2127-2130.
Serri, G., Innocenti, F. and Manetti, P., 1993, Geochemical and petrological evidence of the subduction of delaminated Adriatic continental lithosphere in the genesis of the Neogene–Quaternary magmatism of central Italy: Tectonophysics, v. 223, pp. 117-147.
Valensise, G, Amato, A., Montone, P. and Pantosti, D., Earthquakes in Italy: past, present and future: Episodes, this volume.
Ventura, B., Pini, G.A., and Zuffa, G.G., 2001, Thermal history and exhumation of the Northern Apennines (Italy): evidence from combined apatite fission track and vitrinite reflectance data from foreland basins sediments: Basin Research, v. 13, pp. 1-19.
Ward, S.N., and Valensise, G.R., 1989, Fault parameters and slip distribution of the 1915 Avezzano, Italy, earthquake derived from geodetic observations: Bulletin of the Seismological Society of America, v. 79, pp. 690-710.
Wezel, F.C., 1985, Structural features and basin tectonics of the Tyrrhenian sea, in Stanley D.J. and F.C. Wezel, eds, Geological evolution of the Mediterranean basin, Springer-Verlag, New York, pp. 186-192.
Zattin M., Picotti V., and Zuffa G. G., 2002, Fission-track reconstruction of the front of the Northern Apennine thrust wedge and overlying Ligurian Unit: American Journal of Science, v. 302, pp. 346-379.
Zeitler P.K., Meltzer, A. S., Koons, P.O., Craw, D., Hallet, B., Chamberlain, C. P., Kidd, W. S. F., Park, S. K., Seeber, L., Bishop, M. and Shroeder, J., 2001, Erosion, Himalayan geodynamics, and the geomorphology of metamorphism: GSA Today, v. 11, no. 1, pp. 4-9.

***Carlo Bartolini**, professor of Physical Geography and Geomorphology at the University of Florence, Italy, has worked on marine geology and coastal geomorphology, neotectonics and morphotectonics. His present research is focused on exhumation/denudation rates over the Northern Apennine and on the impact exerted on the drainage evolution by the differential uplift affecting the chain. He acted as secretary of INQUA Neotectonics Commission from 1987 to 1990 and president from 1991 to 1999.*

***Francesco Dramis** is professor of Geomorphology at the "Roma Tre" University, Italy. He has carried out research in several sectors of Geomorphology and Quaternary Geology, including Morphotectonics, Slope Geomorphology and Periglacial Geomorphology. He is currently working on the morphotectonic evolution of the intra-Apennine depressions, the relationships between active tectonics and large scale gravitational phenomena, and the geomorphic-sedimentary effects of Late Pleistocene-Holocene climate changes and tectonics in East Africa and the Mediterranean. He acted as Coordinator of the INQUA Working Group "Mountain Building" from 1996 to 2003.*



***Nicola D'Agostino** is a Research Scientist at the Italian National Institute of Geophysics and Volcanology (INGV) from 2002. He earned both his Degree in Geological Sciences in 1992 and PhD in 1998 from the University of Rome. His research activity includes studies on the Plio-Quaternary extensional tectonics of the Apennines, the application of spectral analyses of gravity and topography for the characterization of the long-term strength of the lithosphere, and the Quaternary geomorphology of the Apennines. His current interest is mainly focused on the application of geodetic (GPS) and geomorphologic analyses to the study of active deformation of the Italian area.*



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)

1 Enel GreenPower S.p.a. Via Andrea Pisano, 120 - Pisa (Italy); *batini.fausto@enel.it*
2 Department of Earth Sciences, University of Siena - Via Laterina, 8 - Siena (Italy); *brogiandrea@unisi.it*
3 Department of Earth Sciences, University of Siena - Via Laterina, 8 - Siena (Italy); *lazzarotto@unisi.it*
4 Department of Geology and Geophysics, University of Bari - Via Orabona 4 - Bari (Italy); *d.liotta@geo.uniba.it*
5 Department of Earth Sciences, University of Florence - Via G. La Pira, 4 - Firenze (Italy); *pandeli@geo.unifi.it*
6 CNR-Institute of Earth Sciences and Earth Resources-Section of Florence, Via G. La Pira, 4 - Firenze (Italy).

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

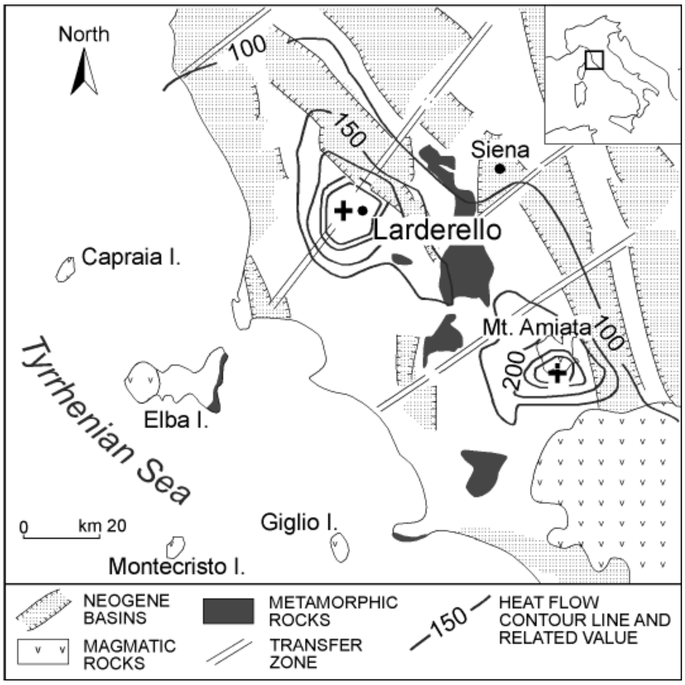


Figure 1 *Structural sketch map of southern Tuscany with the regional heat flow contour lines (equidistance: 50 mW/m²). The Larderello-Travale and Mt. Amiata geothermal fields are located in areas where heat flow reaches 1 W/m² and 0.6 W/m², respectively. (after Baldi et al., 1995)*

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).

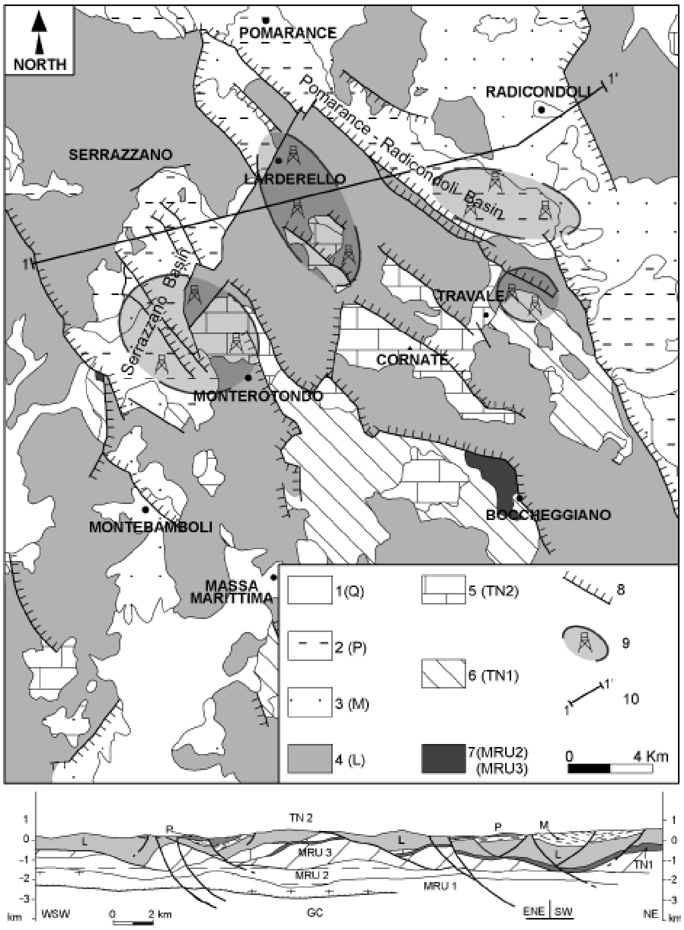


Figure 3 Geological sketch map of the Larderello-Travale area. Key: Neogene and Quaternary deposits: 1—Quaternary continental sediments; 2—Pliocene marine sediments; 3—Miocene continental and marine sediments; 4—Ligurian units l.s. (Jurassic- Eocene); 5—Tuscan Nappe: Late Triassic-Early Miocene sedimentary 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

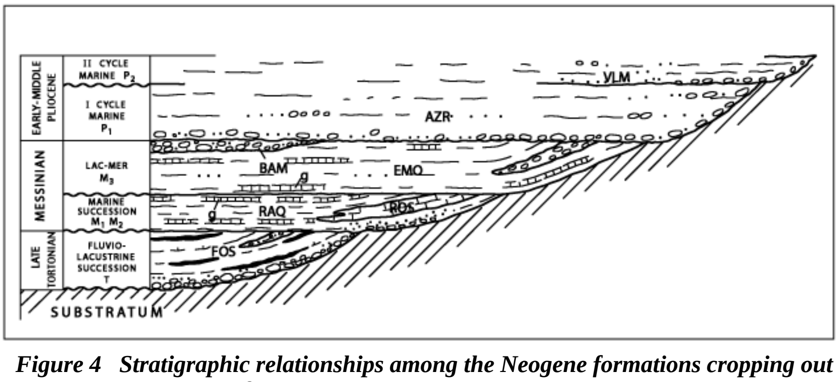


Figure 4 Stratigraphic relationships among the Neogene formations cropping out in the surroundings of the Larderello-Travale geothermal area; RAQ: Raquese Stream Fm. (Early Messinian), ROS: Rosignano Limestone (Early Messinian), EMO: Clays and gypsum of Era Morta River (Late Turolian), BAM: Montebamboli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

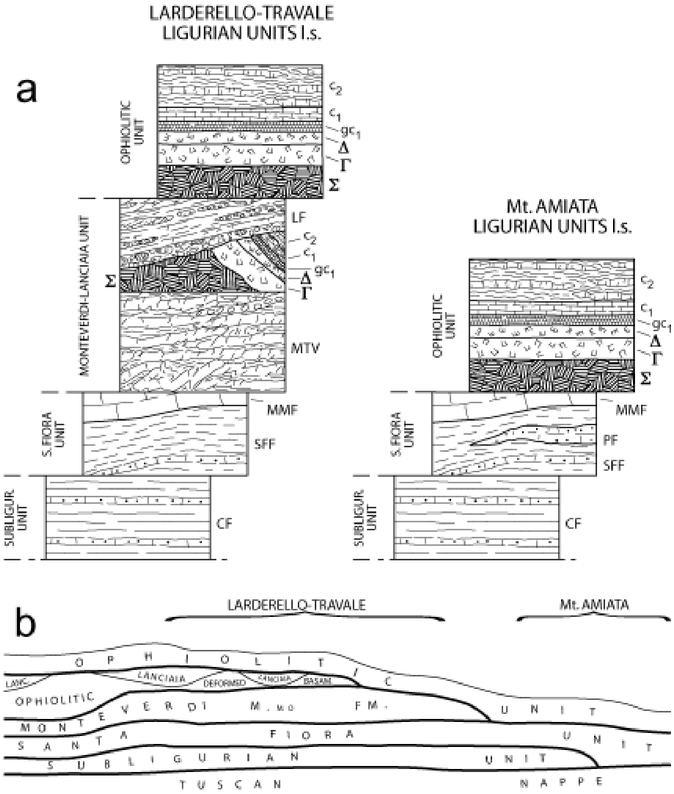


Figure 5 a) Structural and stratigraphic relationships between the Ligurian and Subligurian Units in the Larderello-Travale and Mt. Amiata geothermal areas. Ophiolitic Unit: Δ, Γ, ∑: Ophiolites (serpentinites, gabbros, basalts—Middle-Late Jurassic), gc_1: Mt. Alpe cherts (Late Jurassic), c_1: Calpionella 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.

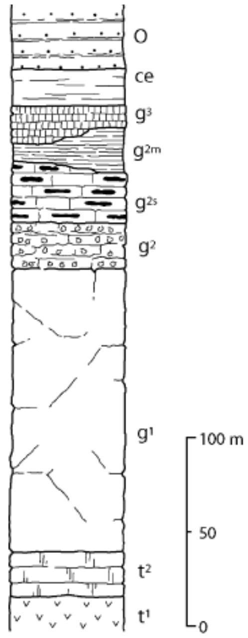


Figure 6 Stratigraphic succession of the Tuscan Nappe.
Symbols: O: Macigno Fm. (Late Oligocene-Early Miocene); ce: Scaglia toscana Fm. (Cretaceous-Oligocene); c^1: Maiolica Fm. (Early Cretaceous); g^3: Diaspri Fm. (Malm); g^{2m}: Marne a Posidonia (Dogger); g^{2s}: Calcare Selcifero Fm. (Middle-Late 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

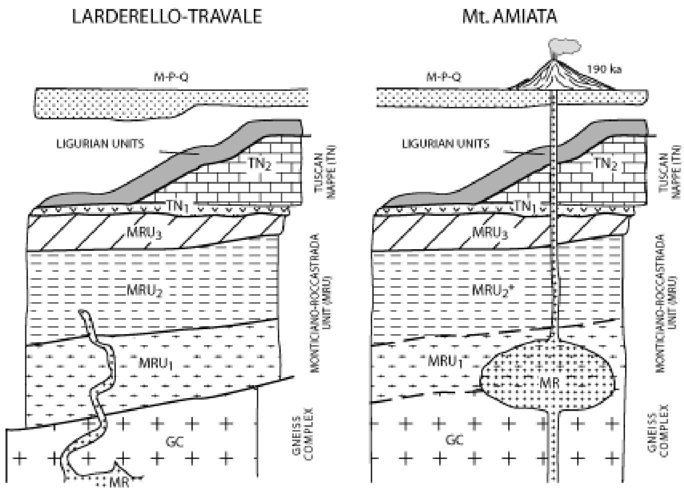


Figure 7 Tectonostratigraphic units in the Larderello-Travale and Mt. Amiata areas. Q-P-M: Quaternary, Pliocene and Miocene sediments; MR—Magmatic rocks; Tuscan Nappe (TN): TN_2—Early Miocene-Rhaetian sequence; TN_1—Late Triassic evaporite (Burano Fm.); Monticiano-Roccastrada Unit (MRU): MRU_3—Triassic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

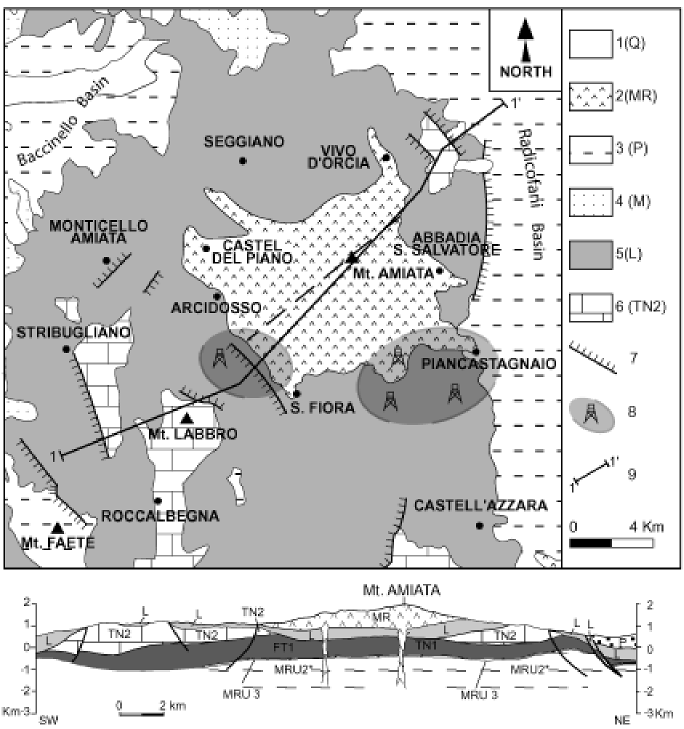


Figure 8 Geological sketch map of the Mt. Amiata area. Keys: 1—Quaternary continental sediments; 2—Magmatic rocks; 3—Pliocene marine sediments; 4—Miocene continental, brackish and marine sediments; 5—Ligurian Units l.s. (Jurassic-Eocene); 6—Tuscan Nappe (Late Trias-Early Miocene); 7—normal faults; 8—Main 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).*

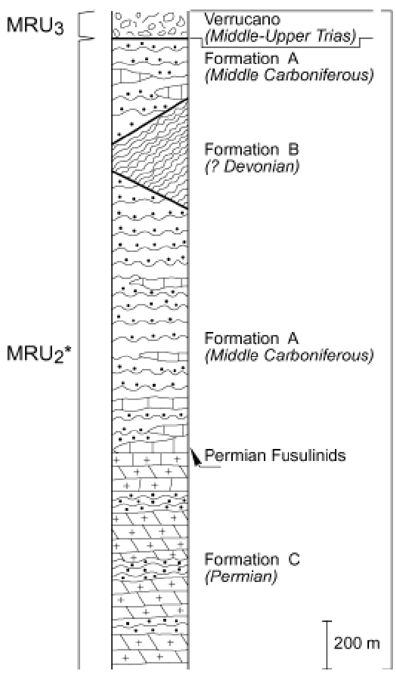


***Figure 9** Relationship among the Triassic and Palaeozoic formations belonging to Monticiano-Roccastrada Unit encountered by geothermal wells in the Mt. Amiata area (after Elter & Pandeli, 1991).*

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,

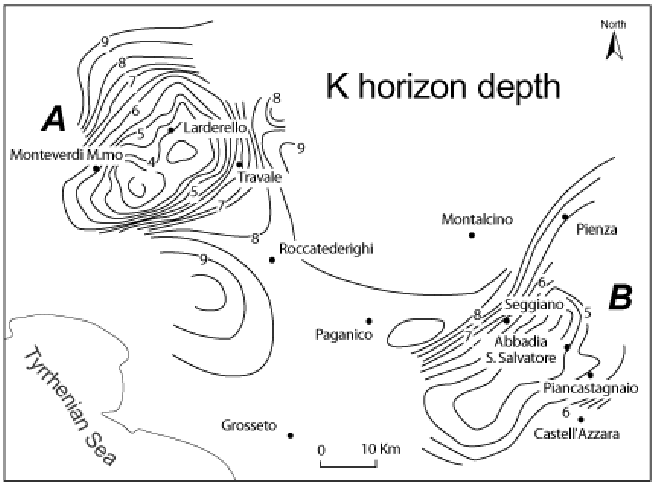


Figure 10 Contour lines in kilometres (equidistance: 0.5 km) of the K-horizon depth. A and B show respectively the Larderello-Travale and Mt. Amiata geothermal areas (after Cameli et al., 1998).

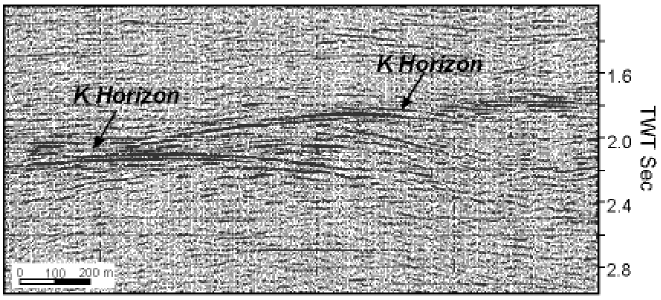


Figure 11 The K-horizon in the Larderello-Travale geothermal area. Vertical axis: TWT seconds.

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).

References

Baldi P., Bellani, S., Ceccarelli, A., Fiordelisi, A., Squarci, P. and Taffi, L., 1995, Geothermal anomalies and structural features of southern Tuscany - World Geothermal Congress Proceedings - Florence, Italy, May 1995, 1287-1291.

Baldi, P., Bertini, G., Cameli, G.M., Decandia, F.A., Dini, I., Lazzarotto, A. and Liotta D., 1994, La tettonica distensiva post-collisionale nell'Area Geotermica di Larderello (Toscana meridionale), Studi Geol. Camerti, Vol. spec., 1994/1, 183-193.

Barbier, E., 2002, Geothermal energy and current status: an overwiew, Renewable and suistainable energy, Reviews, 6, 3-65.

Batini, F., Burgassi, P. D., Cameli, G. M., Nicolich, R. and Squarci P., 1978, Contribution to the study of the deep lithospheric profiles: deep reflecting horizons in Larderello-Travale geothermal field, Mem. Soc. Geol. It., 19, 477-484.

Batini, F., Bertini, G., Gianelli, G., Pandeli, E. and Puxeddu M., 1985, Deep structure of the Larderello field: contribution from recent geophysical and geological data, Mem. Soc. Geol. Ital., 25 (1983), 219-235.

Bertini, G., Cameli, G.M., Costantini, A., Decandia, F.A., Dini, I., Elter, F.M., Lazzarotto, A., Liotta, D., Pandeli, E. and Sandrelli, F., 1994, Structural features of southern Tuscany along the Monti di Campiglia-Rapolano Terme cross-section, Mem. Soc. Geol. It., 48, 51-59.

Bertini, G., Cappetti, G., Dini, I. and Lovari, F., 1995, Deep drilling results and updating of geothermal knowledge of the Monte Amiata area, Proceedings of the World Geothermal Congress, Florence (Italy) 18-31 May 1995 - International Geothermal Association (1995), 2, 1283-1286.

Bertini, G., Gianelli, G. and Battaglia, A., 1996, Risultati ed interpretazione delle datazioni radiometriche (metodo $^{230}Th/^{234}U$) dei campioni di minerali idrotermali presenti nelle rocce attraversate dai sondaggi geotermici (Larderello e Monteverdi) e negli affioramenti di rocce mineralizzate (Sassa e Canneto-Malentrata), ENEL-CNR-CISE internal report, Pisa 1996, 18 pp.

Brogi, A., Lazzarotto, A., Liotta, D. and Ranalli G., 2003, Extensional shear zone as imaged by reflection seismic lines: the Larderello geothermal field (central Italy), Tectonophysics, 363, 127-139.

Brunét, C., Monié, P., Jolivet, L. and Cadet, J.P., 2000, Migration of compression and extension in the Tyrrhenian Sea, insights from $^{40}Ar/^{39}Ar$ ages on micas along a transect from Corsica to Tuscany, Tectonophyics, 321, 127-155.

Calcagnile, G., and Panza, G.F., 1981, The main characteristics of the lithosphere-astenosphere system in Italy and surrounding regions, Pure Appl. Geophys., 119, 865-879.

Cameli, G.M., Dini, I. and Liotta, D., 1993, Upper crustal structure of the Larderello geothermal field as a feature of post collisional extensional tectonics (Southern Tuscany, Italy), Tectonophysics, 224, 413-423.

Cameli, G.M., Dini, I. and Liotta, D., 1998, Brittle/ductile boundary from seismic reflection lines of southern Tuscany (Northern Apennines, Italy), Mem. Soc. Geol. It., 52, 153-163.

Cappetti, G., Passaleva, G., Sabatelli, F., 2000, Italy country update report 1995-1999, Proc. World Geother. Congr. 2000, Kyushu-Tohoku, May 28-June 10, 109-116.

Carmignani L., Decandia F.A., Fantozzi P.L., Lazzarotto A., Liotta D. and Meccheri M., 1994, Tertiary extensional tectonics in Tuscany (Northern Apennines, Italy), Tectonophysics, 238, 295-315.

Del Moro, A., Puxeddu, M., Radicati Di Brozolo, F., and Villa, I. M., 1982, Rb-Sr & K-Ar ages on minerals at temperatures of 300-400˚C from deep wells in the Larderello geothermal field (Italy), Contr. Mineral. Petrol., 81, 340-349.

Elter, F.M. and Pandeli, E., 1990, Alpine and Hercynian Orogenic phases in the Basement rocks of the Northern Apennines (Larderello Geothermal field, Southern Tuscany, Italy), Ecl. Geol. Elv., 83(2), 241-264.

Elter, F.M. and Pandeli, E., 1991, Structural features of the metamorphic Palaeozoic - Triassic sequences in deep geothermal drillings in the Monte Amiata area (SE Tuscany, Italy), Boll. Soc. Geol. Ital., 110, 511-522.

Ferrari, L., Conticelli, S., Burlamacchi, L. and Manetti, P., 1996, Volcanological evolution of the Monte Amiata, Southern Tuscany: New geological and petrochemical data, Acta Vulcanologica, 8, 41-56.

Gianelli, G., 1994, Ipotesi di un modello di crosta superiore per le aree geotermiche toscane, Studi Geologici Camerti, Vol. Spec. 1994/1, CROP18, 195-200.

Gianelli, G., and Laurenzi, M.A., 2001, Age and cooling rate of the geothermal system of Larderello, Geoth. Resources Counsil Transaction, 25, 731-735.

Gianelli, G., Manzella, A. and Puxeddu, M., 1997, Crustal models of the geothermal areas of southern Tuscany, Tectonophysics, 281, 221-239.

Gianelli, G., Puxeddu, M., Batini, F., Bertini, G., Dini, I., Pandeli, E. and Nicolich, R., 1988, Geological model of a young volcano-plutonic system: the geothermal region of Monte Amiata (Tuscany, Italy), Geothermics, 17, 719-734.

Kligfield, R., Hunziker, J., Dallmeyer, R.D. and Schamel, S., 1986, Dating of deformation phase using K-Ar and ^{40}Ar- ^{39}Ar techniques: Results from the Northern Apennines, Jour. of Structural Geol., 8, 781-798.

Liotta, D. and Ranalli, G., 1999, Correlation between seismic reflectivity and rheology in extended lithosphere: southern Tuscany, Inner Northern Apennines, Italy, Tectonophysics, 315, 109-122.

Manzella, A., Ruggieri, G., Gianelli, G. and Puxeddu, M., 1998, Plutonic-geothermal system of southern Tuscany: a review of the crustal models, Mem. Soc. Geol. It., 52, 283-294.

Minissale, A., 1991, The Larderello geothermal field: a review, Earth-Sciences Reviews, 31, 133-151.

Musumeci, G., Bocini, L. and Corsi, R., 2002, Alpine tectonothermal evolution of the Tuscan Metamorphic Complex in the Larderello geothermal field (Northern Apennines, Italy). Jour. of the Geol. Soc., London, 159, 443-456.

Pandeli, E., Gianelli, G., Puxeddu, M. and Elter, F.M., 1994, The Paleozoic basement of the Northern Apennines: stratigraphy, tectono-metamorphic evolution and Alpine hydrothermal processes, Mem. Soc. Geol. It., 48, 627-654.

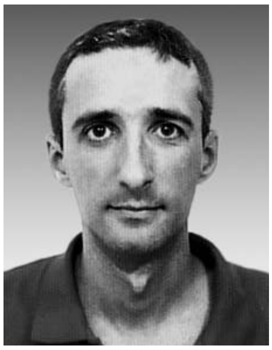
Serri, G., Innocenti, F. and Manetti, P., 1993, Geochemical and petrological evidence of the subduction of delaminated Adriatic continental lithosphere in the genesis of the Neogene-Quaternary magmatism of Central Italy, Tectonophysics, 223, 117-147.

Van Bergen, M.J., 1983, Polyphase metamorphic sedimentary xenoliths from Mt. Amiata volcanics (Central Italy); evidence for a partially disrupted contact aureole. Geol. Rundsch., 72, 637-662.

Villa, I. and Puxeddu, M., 1994, Geochronology of the Larderello geothermal field: new data and the 'closure temperature' issue, Contrib. Mineral. Petrol., 315, 415-426.

Fausto Batini *is manager of geothermal exploration of the Enel Green Power Italian electric company. He has published numerous research papers on the Italian geothermal fields.*

Domenico Liotta *is Associate Professor of Structural Geology at the Geology and Geophysics Department of Bari University, Italy. He is author of several papers on the evolution and structure of the Larderello crust.*



Andrea Brogi *is Post-Doctoral fellowship at the Department of Earth Science of Siena University, Italy. His research interest is focused on tectonic evolution of the Northern Apennines and, in particular, on the extensional tectonics affecting the Tyrrhenian Sea and Tuscany.*

Enrico Pandeli *is Associate Professor of Stratigraphy and Sedimentary Geology at the Earth Science Departmente of Florence University, Italy. His scientific interests mainly regard the stratigraphy, petrography and structural setting of the metamorphic and sedimentary tectonic Units of the Northern Apennines.*

Antonio Lazzarotto *is Full Professor of Geology at the Department of Earth Science of Siena University, Italy. He is author of stratigraphic papers, geological surveys and geological syntheses about the Northern Apennines. He was Leader of several national and multidisciplinary projects (Progetto Energetica, CROP, New Edition of the Geological Map of Italy).*

by Angelo Peccerillo

Plio-Quaternary magmatism in Italy

Dipartimento di Scienze della Terra, University of Perugia, Piazza Università, 06100 Perugia, Italy. E-mail: *pecceang@unipg.it*

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

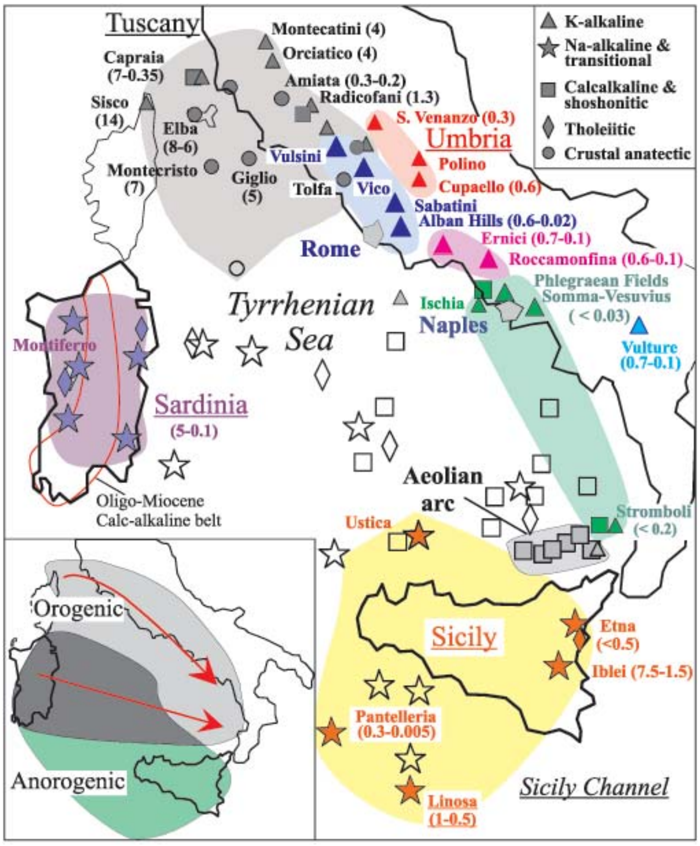


***Figure 1** Distribution of Recent magmatism in Italy. Open symbols indicate seamounts. Ages (in Ma) are given in parentheses. Different colours denote various magmatic provinces. Inset: schematic distribution of orogenic and anorogenic volcanism: red arrows indicate migration of orogenic magmatism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

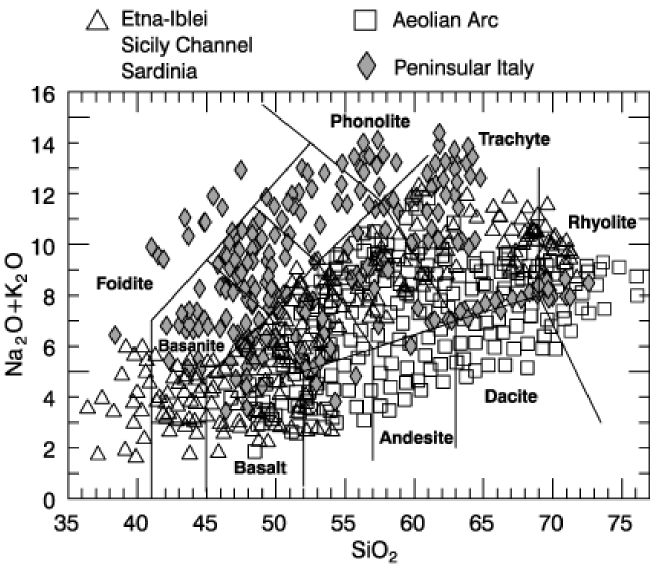


Figure 2 Total alkali vs. silica classification diagram for Italian Plio-Quaternary magmatic rocks. For source of data see Peccerillo (2002).

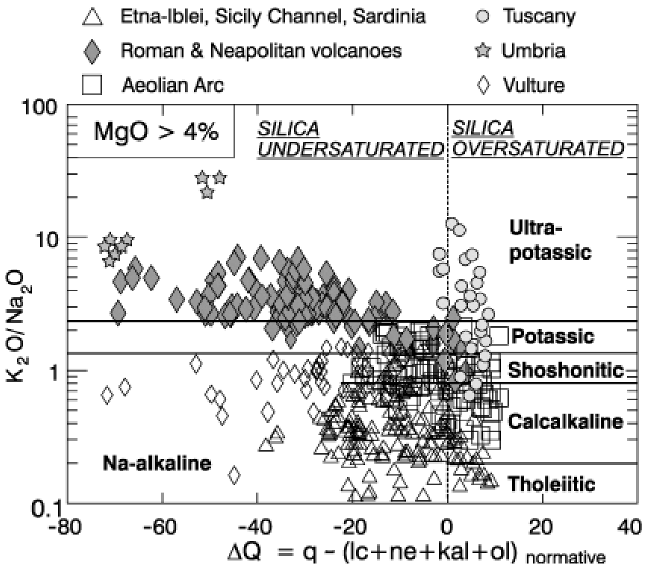


Figure 3 ΔQ vs. K_2O/Na_2O classification diagram for Plio-Quaternary mafic (MgO > 4%) volcanic rocks from Italy. ΔQ is the algebraic sum of normative quartz (q), minus leucite (lc), nepheline (ne), kalsilite (kal) and olivine (ol). Silica oversaturated rocks have $\Delta Q > 0$, whereas silica undersaturated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

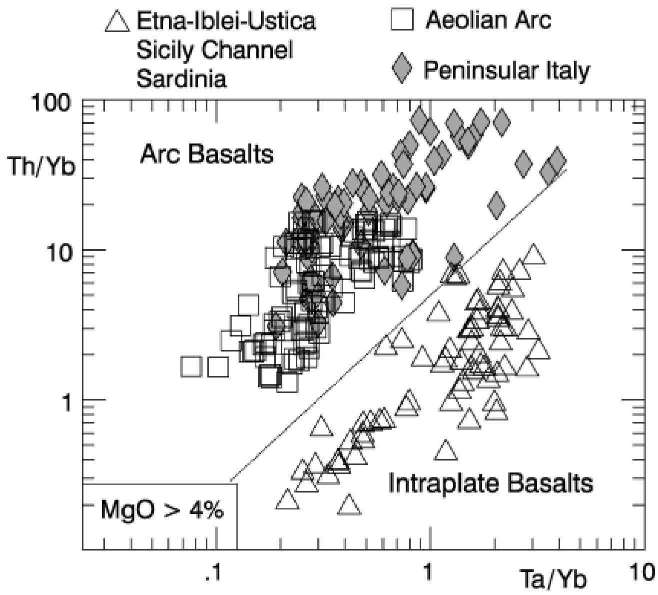
A large proportion of Italian Plio-Quaternary volcanic rocks have high-silica, low-MgO compositions. However, mafic rocks (MgO>3–4 wt%) deserve particular attention, since they are the closest relatives of primary mantle-derived magmas that were parental to erupted lavas, and can furnish the maximum 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

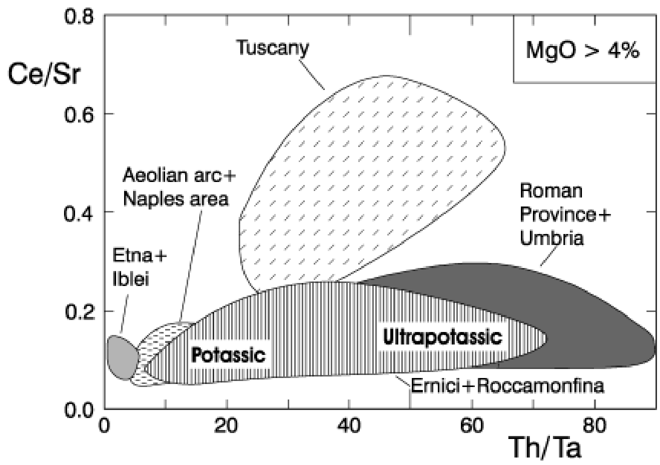


Figure 5 Variation of trace element ratios in Plio-Quaternary Italian mafic rocks. Note strong regional variation.

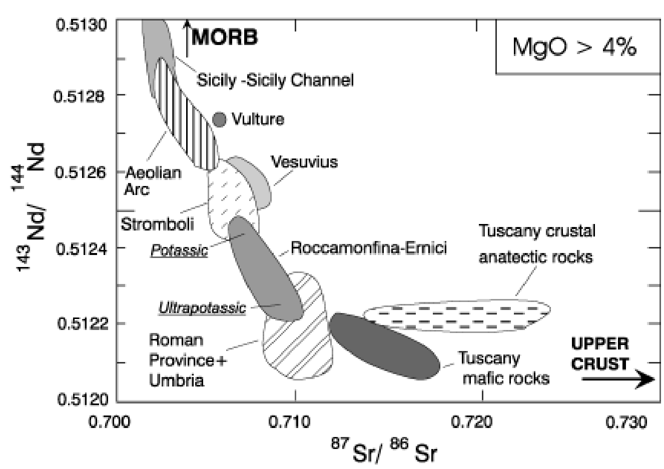


Figure 6 Sr vs. Nd isotope diagram for Plio-Quaternary mafic volcanic rocks from Italy. Note strong regional variation.

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

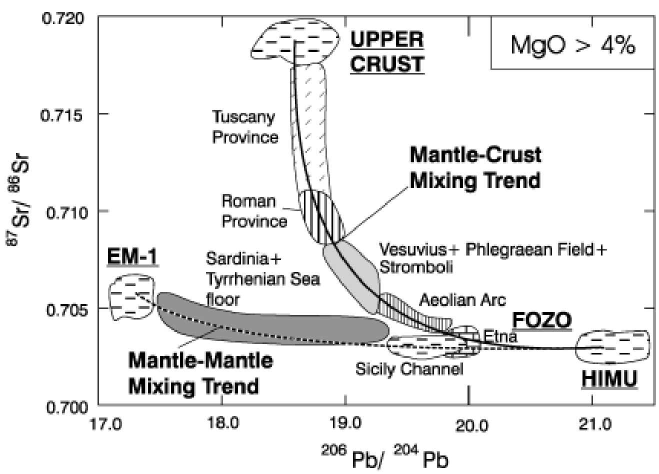


***Figure 7** $^{87}Sr/^{86}Sr$ vs. $^{206}Pb/^{204}Pb$ variations of Plio-Quaternary Italian mafic rocks. Central Italy orogenic magmatism falls along a mantle-crust mixing trend involving HIMU-FOZO and Upper Crust. Sicily, Sicily Channel, Sardinia and some Tyrrhenian Sea seamounts (anorogenic magmatism)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

Acknowledgements

The research on Italian volcanism has been financed by CNR, INGV and by the University of Perugia. The author thanks R.S. Harmon for the critical reading of the manuscript, and Dr. Carmelita Donati and Danilo Chiocchini, University of Perugia, for helping with construction of data base and drawing of figures.

References

Ayuso, R.A., De Vivo, B., Rolandi, G., Seal, R.R., and Paone, A., 1997, Geochemical and isotopic (Nd-Pb-Sr-O) variations bearing on the genesis of volcanic rocks from Vesuvius, Italy. Journal of Volcanology and Geothermal Research, v. 82, pp. 53-78.

Beccaluva, L., Macciotta, G., Morbidelli, L., Serri, G., and Traversa, G., 1989, Cainozoic tectono-magmatic evolution and inferred mantle sources in the Sardo-Tyrrhenian area. *in* Boriani, A., Bonafede, M., Piccardo, G. B., and Vai, G. B., Editors, Atti Convegni Lincei, v. 80, pp. 229-248.

Conticelli, S., D'Antonio, M., Pinarelli, L., and Civetta, L., 2001, Source contamination and mantle heterogeneity in the genesis of Italian potassic and ultrapotassic rocks: Sr-Nd-Pb isotope data from Roman Province and southern Tuscany. Mineralogy and Petrology, v. 74, pp. 189-222.

Doglioni, C., Harabaglia, P., Merlini, S., Mongelli, F., Peccerillo, A., and Piromallo, C., 1999, Orogens and slabs vs. their direction of subduction. Earth Science Review, v. 45, pp. 167-208.

Gasperini, D., Blichert-Toft, J., Bosch, D., Del Moro, A., Macera, P., and Albaréde, F., 2002, Upwelling of deep mantle material through a plate window: Evidence from geochemistry of Italian basaltic volcanics. Journal of Geophysical Research, v. 107, pp. 2367-2386.

Harmon, S. R., and Hoefs, J., 1995, Oxygen isotope heterogeneity of the mantle deduced from global ^{18}O systematics of basalts from different geotectonic settings. Contributions to Mineralogy and Petrology, v. 120, pp. 95-114.

Hoffmann, A.W., 1997, Mantle geochemistry: the message from oceanic volcanism. Nature, v. 385, pp. 219-229.

Lustrino, M., Melluso, L., and Morra, V., 2000, The role of lower continental crust and lithospheric mantle in the genesis of Plio-Quaternary volcanic rocks from Sardinia (Italy). Earth and Planetary Science Letters, v. 180, pp. 259-270.

Panza, G.F., and Mueller, S., 1979, The plate boundary between Eurasia and Africa in the Alpine area. Memorie della Società Geologica, v. 33, pp. 43-50.

Peccerillo, A., 1999, Multiple mantle metasomatism in central-southern Italy: geochemical effects, timing and geodynamic implications. Geology, v. 27, pp. 315-318.

Peccerillo, A., 2002, Quaternary magmatism in Central-Southern Italy: a new classification scheme for volcanic provinces and its geodynamic implications. Bollettino della Società Geologica Italiana, Volume Speciale n. 1 (2002), p. 113-127.

Santacroce, R., Cristofolini, R., La Volpe, L., Orsi, G., and Rosi, M., 2003, Italian Active Volcanoes. Episodes, this issue.

Spakman, W., Van Der Lee, S., and Van Der Hilst, R.D., 1993, Travel time tomography of the European-Mediterranean mantle down to 1400 km. Physics of the Earth and Planetary Interior, v. 79, pp. 3-74

Wezel, F.-C., 1985, Structural Features and Basin Tectonics of the Tyrrhenian Sea. *in* Stanley, D.J.and Wezel, F.-C., eds, Geological Evolution of the Mediterranean Basin, Springer-Verlag New York, pp. 153-194.

Wood, D.A., Joron, J.L., and Treuil M. ,1979, A reappraisal of the use of trace elements to classify and discriminate between magma series erupted in different tectonic settings. Earth and Planetary Science Letters, v. 45, pp. 326-336.

Zindler, A. and Hart, S.R., 1986, Chemical geodynamics. Annual Reviews of Earth and Planetary Science, v. 14, pp. 493-571.

Angelo Peccerillo *is full professor of Petrology at the University of Perugia, where he has been teaching igneous and metamorphic petrology, and volcanology. His main fields of interest are igneous petrology and trace element geochemistry. His research has been concentrated on ultrapotassic rocks, subduction-related magmatism and rift volcanism. He is author or co-author of some 130 scientific and of several popular and didactic publications, including three books.*