Univerza v Ljubljani
Naravoslovnotehniška fakulteta
Oddelek za geologijo

Vodič za predmet

SKLEPNE TERENSKE VAJE

2009 - Sicilija

17.05. – 26.05.2009

Avtorja in nosilca predmeta:

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Okvirni program ekskurzije na Sicilijo, 17.05.-26.05.2009

1. Dan: nedelja 17.05.2009
1. Vožnja Ljubljana-Napoli-Palermo
Spanje: trajekt

2 Dan: ponedeljek 18.05.2009
1. Palermo / Monreale
2. Mt. Kumeta - jurska stratigrafija
3. San Vito lo Capo - morske terase
Spanje: Trapani

3. Dan: torek 19.05.2009
1. Segesta / Selinunte - templji
2. Capo Bianco / Punta di Maiata - cikličnost v sedimentih
Spanje: Agrigento

4. Dan: sreda 20.05.2009 - italijanski vodič
1. Vulcanelli di Macalube - vulkanizem
2. Piazza Armerina - vila, drobirski tok
Spanje: Catania

5. Dan: četrtek 21.05.2009 - italijanski vodič
1. Etna - vulkanizem
Spanje: Taormina

6. Dan: petek 22.05.2009 - italijanski vodič
1. Gole di Alcantara - bazalti
2. Mt. Peloritani - metamorfni kompleks
Spanje: Lipari

7. Dan: sobota 23.05.2009 - italijanski vodič
1. Lipari - vulkanizem, obsidijan
Spanje: Lipari

8. Dan: nedelja 24.05.2009 - italijanski vodič
1. Lipari, Stromboli - vulkanizem
Spanje: Stromboli

9. Dan: ponedeljek 25.05.2009
1. Stromboli - vulkanizem
Spanje: trajekt

10. Dan: torek 26.05.2009
1. Vožnja Stromboli-Napoli-Ljubljana
Vrnitev v Ljubljano

Tyrrhenian Sea
Mediterranen Sea
Ionian Sea
Strait of Messina
SICILIA
TRAPANI
ERICE
PALERMO
TAORMINA
SIRACUSA
MESSINA
CATANIA
ENNA
CALTANISSETTA
AGRIGENTO
RAGUSA
REGGIO CALABRIA
2. dan
3. dan
4. dan
5. dan
6. dan
7., 8. dan
8., 9. dan
Golfo di Castellammare
Golfo di Carini
Golfo di Palermo
Golfo di Termini Imerese
Golfo di Patti
Golfo di Milazzo
Golfo di Catània
Golfo di Augusta
Golfo di Noto
Marsala
Mazara del Vallo
Castelvetrano
Selinunte
Menfi
Sciacca
Ribera
Eraclea Minoa
Porto Empedocle
Licata
Gela
Vittoria
Comiso
Modica
Scicli
Noto
Avola
Pachino
Portopalo di Capo Passero
Augusta
Lentini
Caltagirone
Piazza Armerina
Cefalù
Milazzo
Tindari
Patti
Bronte
Randazzo
Giardini di Naxos
Acireale
Favignana
Pelagie Islands
KEY
Airport
Ferry port
Railway station
Motorway (highway)
Main road
Minor road
Other road
0 kilometres 25
0 miles 15
ISOLE EOLIE
Alicudi
Filicudi
Salina
Panarea
Stromboli
Lipari
Vulcano
Milazzo
SICILIA

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09	Iveta	Ana
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12	Kokošin	Jure
13	Korat	Lidija
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15	Košir	Mateja
16	Kržišnik	Nina
17	Marinč	Mihael
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19	Perme	Jure
20	Rot	Miriam
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22	Strahinić	Marko
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25	*Ribičič*	*Mihael*
26	*Rožič*	*Boštjan*
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28	*Šmuc*	*Andrej*
29	*Udovč*	*Miran*
30	*Verbovšek*	*Timotej*
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by William Cavazza[1] and Forese Carlo Wezel[2]

The Mediterranean region—a geological primer

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

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

Introduction

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

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

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

Overview of present-day Mediterranean geological elements

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

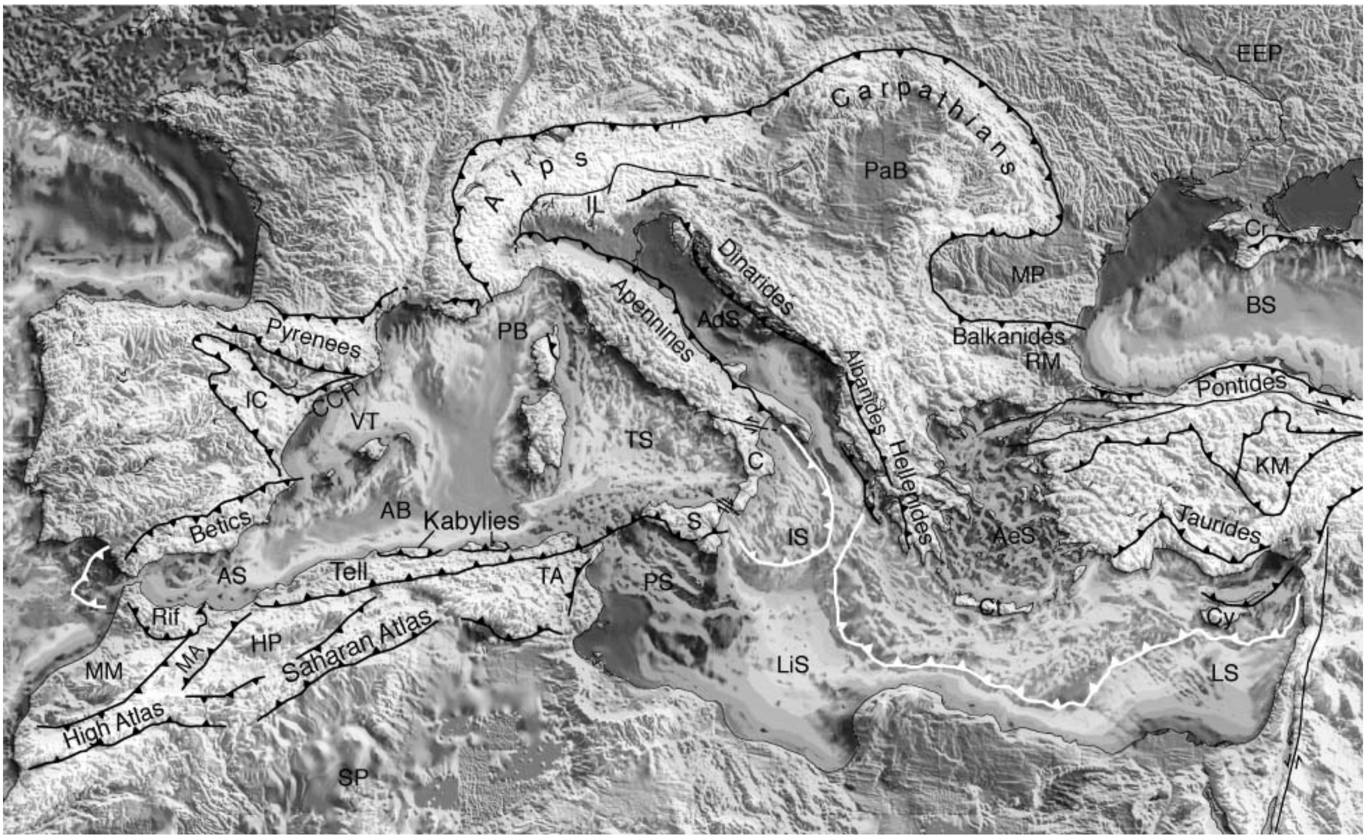

***Figure 1** Digital terrain model of the Mediterranean region with major, simplified geological structures. White thrust symbols indicate the outer deformation front along the Ionian and eastern Mediterraean subduction fronts. AB, Algerian basin; AS, Alboran Sea; AdS, Adriatic Sea; AeS, Aegean Sea;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

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

The modern marine basins of the Mediterranean Sea (Figure 1) are variably floored by (i) remnants of the Tethyan oceanic domains (Ionian and Libyan seas, E Mediterranean), (ii) Neogene oceanic crust (Algero-Provençal basin and Tyrrhenian Sea), (iii) extended continental lithosphere (Alboran Sea, Valencia Trough, Aegean Sea), and (iv) thick continental lithosphere (Adriatic Sea). (i) In the **Ionian-Libyan Sea** and the **eastern Mediterranean** geophysical data (low heat-flow values and thick lithospheric mantle) and palinspastic reconstructions point to the presence of old (Permian?) oceanic crust underneath a thick pile of Mesozoic and Cenozoic sediments which hampers direct sampling and dating; these two oceanic domains are currently being subducted beneath the Calabria-Peloritani terrane of southernmost Italy (see Bonardi et al., 2001, for a review) and the Crete-Cyprus arcs, respectively. The more than 2,000 m deep **Black Sea** is partly floored by oceanic crust and probably represents the remnant of a complex Cretaceous-Eocene back-arc basin which developed on the upper plate of a north-dipping subduction zone (see following section). The western portion of the Black Sea opened in Cretaceous-Paleocene time whereas the East Black Sea basin has a Paleocene-Eocene age (see Robinson, 1997, for a review). (ii) The oceanic **Algero-Provençal basin** opened in the Burdigalian, as indicated by paleomagnetic data and by the transition from syn-rift to post-rift subsidence of its margins (Vially and Trémolières, 1996). Rifting in this area occurred as early as the early Oligocene and induced the development of a series of grabens in southern France and Sardinia both on-land and offshore. The deepest portion of the **Tyrrhenian Sea** is floored by Plio-Quaternary oceanic crust; along its western and eastern margins rift-related grabens contain sedimentary deposits as old as ?Serravallian-Tortonian, thus marking the age of the onset of extension in this region (e.g. Kastens et al., 1990; Mattei et al., 2002). (iii) The **Alboran Sea** is floored by thinned continental crust (down to a minimum of 15 km) and it is bounded to the north, west and south by the Betic-Rif orocline. The basement of the Alboran Sea consists of metamorphic rocks similar to those of the Internal Zones of the Rif-Betics (see below). During the Miocene, considerable extension in the Alboran domain and in the adjacent internal zones of the Betic-Rif occurred coevally with thrusting in the more external zones of these mountain belts. Such late-orogenic extension can be interpreted as the result of subduction roll-back toward the west whereby thickened continental crust extends rapidly as the subduction zone retreats (Lonergan and White, 1997; Gutscher et al., 2002). The **Valencia Trough** is floored by thinned continental crust covered by Mesozoic sedimentary deposits; this assemblage underwent extension starting from the late Chattian. Structurally related to the oceanic Provençal basin to the northeast, the Valencia Trough displays younger syn-rift deposits thus indicating a progressive southwestward rift propagation from southern France (Camargue, Gulf of Lions) (Roca, 2001). The **Aegean Sea** is located in the upper plate of the Hellenic subduction zone. Crustal-scale extension in this region has been accommodated by shallow dipping detachment faults. It has started at least in the early Miocene, and continues today in areas like the Corinth-Patras rift and the southern Rhodope Massif in western Turkey. Miocene extension was accompanied by exhumation of metamorphic rocks and by the intrusion of granitoid and monzonitic magmas at upper crustal levels. According to Jolivet (2001), the engine for Aegean extension is gravitational collapse of a thick crust, allowed by extensional boundary conditions provided by slab retreat; the rather recent tectonic "extrusion" of Anatolia added only a rigid component to the long lasting crustal collapse in the Aegean region. (iv) The **Adriatic Sea** is floored by 30–35 km thick continental crust whose upper portion is mostly made of a thick succession of Permian-Paleogene platform and basinal carbonates. The Adriatic Sea is fringed to the west and east by the flexural foredeep basins of the Apennines and Dinarides-Albanides, respectively, where several kilometers of synorogenic sediments were deposited during the Oligocene-Quaternary. The Mesozoic Adriatic domain has been considered a continental promontory of the African plate (e.g., Channel et al., 1979; Muttoni et al., 2001); such domain—also known as Adria—includes not only what is now the Adriatic Sea but also portions of the Southern Alps, Istria, Gargano and Apulia.

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

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

The **Alps** are the product of continental collision along the former south-dipping subduction zone between the Adriatic continental domain of the African plate to the south and the southern continental margin of the European-Iberian plate to the north. The lithosphere is thicker (ca. 200 km) in the western Alps, while it is in the order of 140 km along the central and eastern Alps (see Dal Piaz et al., this issue, and contributions in Pfiffner et al., 1996, and Moores and Fairbridge, 1997, for an introduction to the Alps). This supports the notion that collisional coupling was stronger to the west. In fact, the eastern Alps are largely made up of tectonic units derived from Apulia, the Austroalpine nappes, while the western Alps are exclusively made up by more external, and tectonically lower units of the European margin, the Briançonnais terrane and the intervening oceanic units (see Piccardo, this issue). The western Alps include outcrops of blueschists and coesite-bearing, eclogite-facies rocks formed at pressures of up to 30 kbars at depths which may have reached 100 km (see Compagnoni, this issue). Such rocks have yielded radiometric ages as old as 130 Ma, although widespread Eocene metamorphic ages constrain—along with other structural and stratigraphic data—the timing of the collision.

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

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

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

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

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

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

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

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

Most syntheses of the geology of the Mediterranean region have focused on the orogenic belts and have largely disregarded the large marginal intraplate rift/wrench basins located along the adjacent cratons of Africa-Arabia and Europe, ranging in age from Paleozoic to Cenozoic. Peritethyan extensional basins are instead key elements for understanding the complex evolution of this area, as their sedimentary and structural records document in detail the transfer of extensional and compressional stress from plate boundaries into intraplate domains (see contributions in Roure, 1994, and Ziegler et al., 2001). The development of the peritethyan rift/wrench basins and passive margins can be variably related to the opening of the Tethyan system of oceanic basins and the Atlantic and Indian oceans (see following section). Some of these basins are still preserved whereas others were structurally inverted during the development of the Alpine-Mediterranean system of orogenic belts or were ultimately incorporated into it. Examples of inversion include the **Iberian Chain** and **Catalonian Coast Range** (Figure 1) which formed during the Paleogene phases of the Pyrenean orogeny through inversion of a long-lived Mesozoic rift system which developed in discrete pulses during the break-up of Pangea, the opening of the Alpine Tethys and the north Atlantic Ocean (Salas et al., 2001). The Mesozoic rift basins of the **High Atlas** of Morocco and Algeria underwent a first mild phase of inversion during the Senonian followed by more intense deformation during the late Eocene. Frizon de Lamotte et al. (2000) have interpreted the latter, main inversion phase as the result of far-field stress transfer from the north during initiation of northward subduction along the southern margin of Iberia and contemporaneous development of the Rif-Tell accretionary prism. Increased coupling between the prism and the African continental margin induced a third phase of inversion in the Quaternary.

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

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

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

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

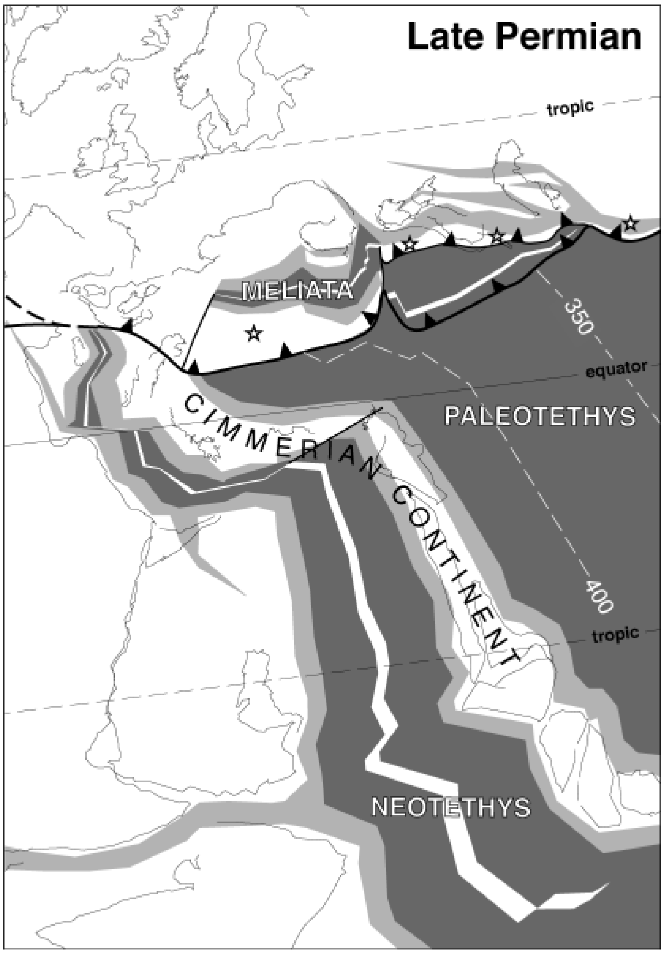

Figure 2 Paleogeographic reconstruction of the western Tethyan area during the late Permian (from Stampfli et al., 2001b, with minor modifications). Stars indicate magmatic activity.

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

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

The picture is further complicated by the Valais-Pyrenean rift zone which started to develop in the early Jurassic as an eastward extension of the central Atlantic, detaching Iberia from Europe (Figure 3, Aptian), and closed by late Eocene time to form the Alps-Carpathians orogenic system (Figure 3, Eocene-Oligocene boundary) (Stampfli et al., 2002). Mid-Jurassic opening of the Ligurian-Piedmont-south Penninic ocean resulted in the development of a new set of passive margins which were traditionally considered for a long time as segments of the northern margin of a single "Tethyan Ocean" stretching from the Caribbean to the Far East. It is somehow a paradox that the Alps—which for almost a century served as an orogenic model for the entire Tethyan region—are actually related to neither paleotethyan nor neotethyan evolution and instead have their origin in the Atlantic Ocean to the west.

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

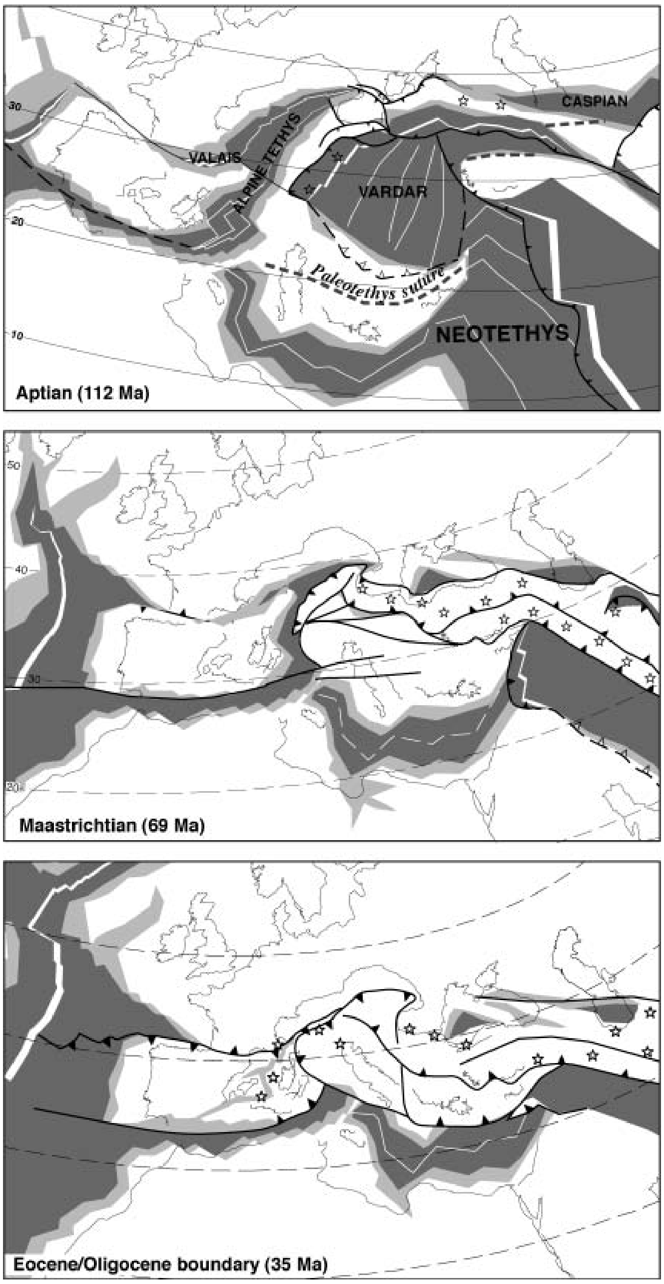

Figure 3 Paleogeographic reconstructions of the western Tethyan area during the Aptian, Maastrichtian and at the Eocene/Oligocene boundary. Note the progressive narrowing and suturing of the oceanic domains comprised between the Eurasian and Iberia continental blocks to the north and the Africa/Arabia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

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

In spite of prolonged indentation along the Alpine front, the Neogene of the Mediterranean region is characteristically dominated by widespread extensional tectonism. A number of small continental microterranes (Kabylies, Balearic Islands, Sardinia-Corsica, Calabria) rifted off the European-Iberian continental margin and drifted toward the south or southeast, leaving in their wake areas of thinned continental crust (e.g. Valencia Trough) or small oceanic basins (Algerian, Provençal and Tyrrhenian basins) (Figure 4). The E Mediterranean is similarly characterized by widespread Neogene extensional tectonism, as indicated by thinning of continental crust along low-angle detachment faults in the Aegean Sea and the 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).

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).

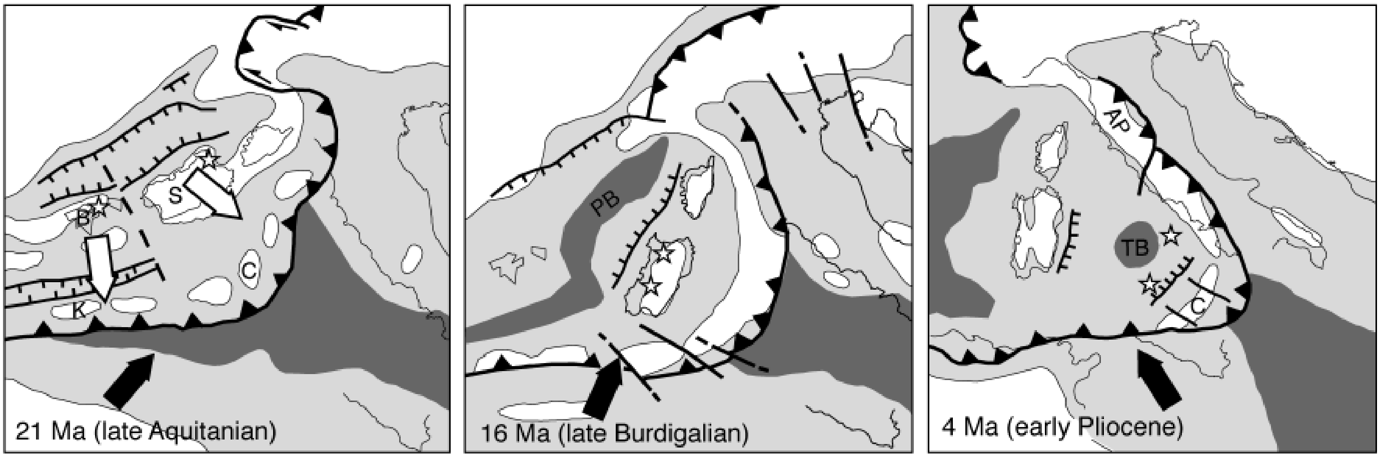

Figure 4 Schematic maps showing the paleotectonic evolution of the W Mediterranean during Neogene time (modified after Bonardi et al., 2001, and Roca, 2001). Only active tectonic elements are shown. White, exposed land; light gray, epicontinental sea; darker gray, oceanic crust. Black arrows indicate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

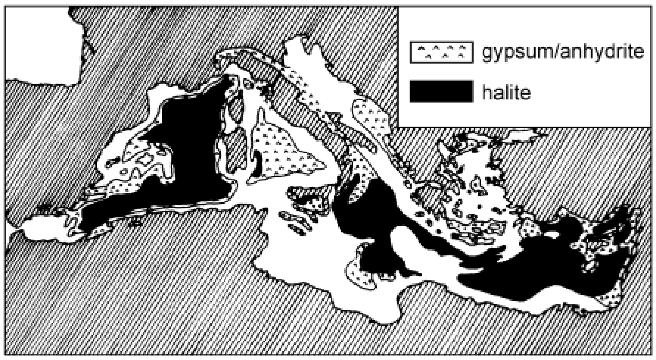

Figure 5 Areal extent of the Messinian evaporites in the Mediterranean region. Modified after Rouchy (1980).

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

The well-exposed Messinian outcrops of central Sicily provide one of the thickest and most complete occurrences of this stage and have been instrumental in the development of current thinking on the Mediterranean evaporites (Figure 6). Hereafter we provide a short description of the stratigraphy of this area as an example of the complexities of the Messinian stratigraphy. At the periphery of the basin the Lower Evaporites—i.e. the Messinian succession below the intra-Messinian unconformity—consist only of two relatively thin units (Figure 6): the Tripoli Formation (laminated diatomites) and the Calcare di Base (evaporitic limestone). In the deepest portions of the basin, the Lower Evaporites are much thicker and comprise, from bottom to top, the Tripoli Fm, the Lower Gypsum Fm (LGF), and the Halite Fm (HF). The LGF is composed of up to 300 m of selenite gypsum with random orientation, indicating that gypsum from the periphery was reworked, deposited in deeper water, and recrystallized; its upper parts consists of gypsum turbidites. The HF is made of up to 800 m of halite with intercalations of potash/magnesium salt beds; this unit was deposited in deep depressions, fed also by clastic resedimentation and slumping. Related to intra-Messinian tectonics, slumping began when the gypsum turbidites of the LGF were deposited and reached its acme at the end of the sedimentaion of the HF. Subaerial erosion occurred in the marginal zones of the basins at the same time as the strata of salts filled up the deep, subsiding depressions. As the potash beds were covered by halite and anhydrite, there are indications of freshening of the brine during the late stages of salt deposition. It appears that these cannot be easily explained by Hsü's (1972) hypothesis of a "deep, dry basin".

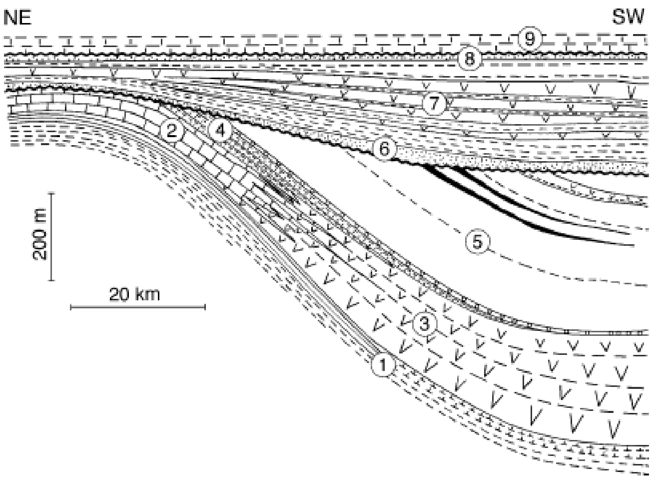

Figure 6 Schematic stratigraphic cross section of the Messinian of Sicily (modified after Decima and Wezel, 1973). 1) Pre-evaporitic clay, marl and diatomite (Tripoli Fm.); 2) evaporitic limestone (Calcare di Base); 3) lower gypsum beds (Gessi di Cattolica); 4) gypsum turbidites; 5) halite and potash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

In Sicily the Lower Evaporites close with the HF, whereas at other Italian sites they terminate with a flysch-like, marly-arenaceous deposit (for example, in the Marche Region), which indicates rapid filling of subsiding troughs. Terrigenous sedimentation was accompanied by cinerite deposition. Taken together, these events suggest that the salts are relatively deep marine syn-diastrophic deposits which correspond to a significant phase of marine regression. In Sicily the salts have been affected by intense tectonic compression with diapiric folds (Decima and Wezel, 1973). The Lower Evaporites were thus deposited during widespread regression which created barriers and subdivided the Tortonian depositional area, with the emersion of vast tracts of land, such as the Central Alboran Sea and the northern Tyrrhenian Sea. At the peak of the lowstand a subaerial erosional surface developed and resulted in the widespread *intra-Messinian inter-regional discontinuity*, which corresponds to a sequence boundary separating the Lower and Upper Evaporite deposits.

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

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

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

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

Structural setting of the Apennine-Maghrebian thrust belt

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

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

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

Introduction

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

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

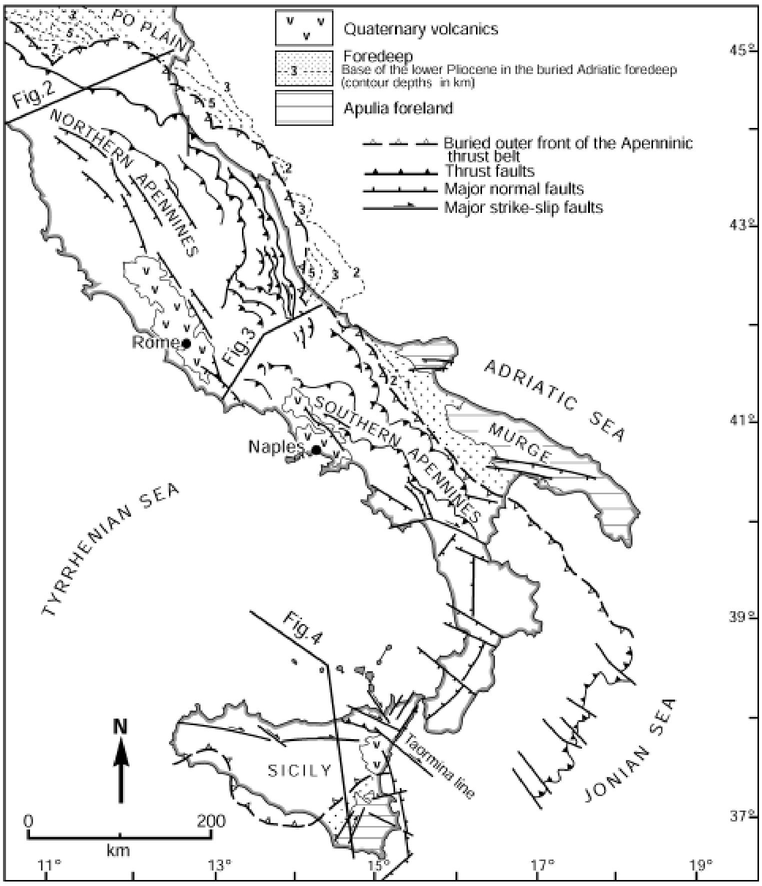

Figure 1 General structural map of the Apennine-Maghrebian chain. Fig.2, Fig.3, Fig.4: locations of the cross-sections of Figures 2–4. (After Ghisetti and Vezzani, 1999, modified)

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

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

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

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

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

The Kabylo-Calabride terranes

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

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

Internal domain

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

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

Liguride units

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

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

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

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

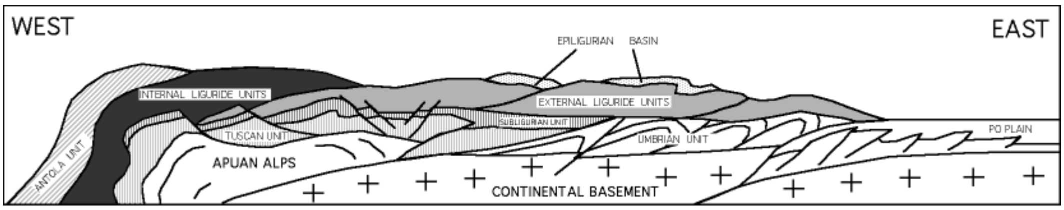

***Figure 2** Schematic cross-section showing the geometric relations among the major structural units of the northern Apennine. For location, see Figure 1. (P. Elter)*

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

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

Sub-Liguride units and Sicilide units

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

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

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

External domain

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

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

Carbonate platform successions

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

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

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

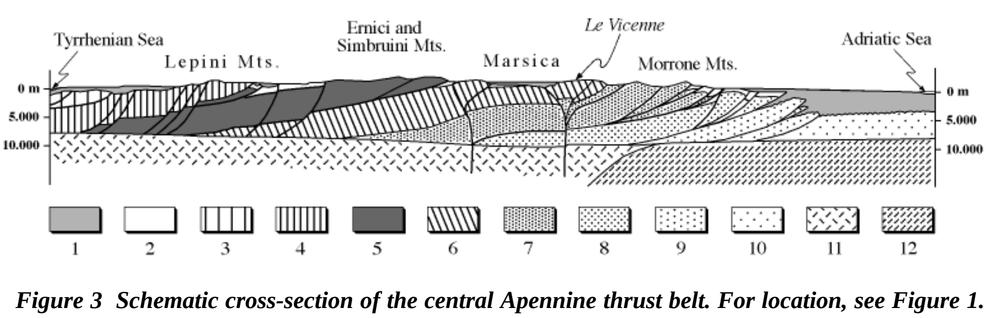

***Figure 3** Schematic cross-section of the central Apennine thrust belt. For location, see Figure 1. 1. marine and continental post-orogenic sedimentary cover (Tyrrhenian side) and intermountain basins (Late Messinian-Quaternary); marine syn- and post-orogenic deposits (Adriatic side: early Pliocene-Quaternary);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

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

Pelagic Basin successions

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

Kinematic reconstruction

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

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

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

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

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

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

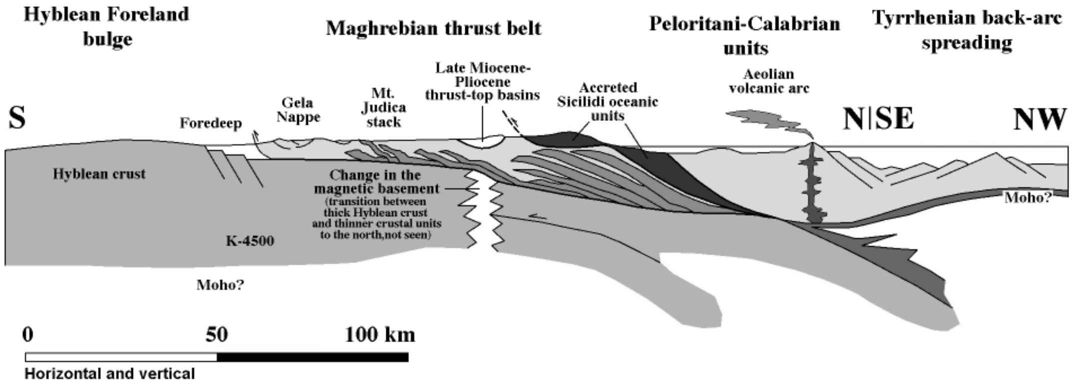

***Figure 4** Simplified section across the eastern Sicily, from the Hyblean foreland to the southern Tyrrhenian back-arc basin; K indicates the magnetic susceptibility. For location, see Figure 1. (M. Grasso)*

Open issues and discussion

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

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

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

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

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

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

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

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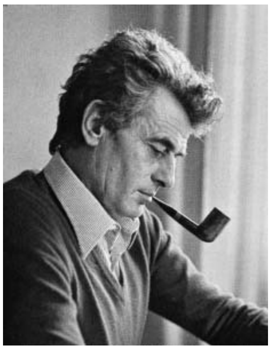

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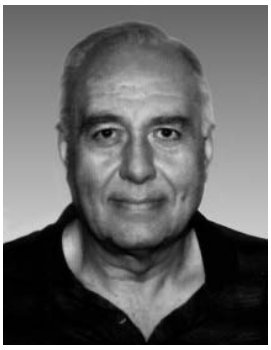

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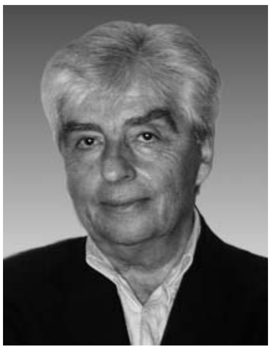

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Italian active volcanoes

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The eruptive histories, styles of activity and general modes of operation of the main active Italian volcanoes, Etna, Vulcano, Stromboli, Vesuvio, Campi Flegrei and Ischia, are described in a short summary.

Introduction

The arrangement of the Mediterranean area essentially results from the subduction of the African plate below the Eurasian one. This induced since the Late Cretaceous, the progressive closure of the Tethys ocean basin, whose remnant is the Mediterranean Sea. In addition to Africa and Eurasia, other microplates are involved in Mediterranean tectonics, although no unanimous consensus exists on their number and geometry. Within this intercontinental interplate system, where compressional and extensional events show close occurrence in time and space, areas with different margin characteristics (stable, unstable convergent, unstable divergent) closely coexist. As a response to such a situation, the Central Mediterranean area, and namely Italy, has been the site of vigorous volcanic activity since Oligocene. Currently at least four areas can be considered active: 1. the Campanian Plain and its offshore area, hosting Campi Flegrei, Ischia and Vesuvio (latest eruption in 1944); 2. the Aeolian archipelago and its extension into the Tyrrhenian Abyssal Plain, persistently active at Stromboli and with historical eruptions at Vulcano and Lipari; 3. Mount Etna, persistently active; 4. Sicily Channel, where sporadic submarine eruptions occurred in 1831 (Ferdinandea/Graham), 1891 (Foerstner, offshore of Pantelleria) and (possibly) 1911 (Mt. Pinne).

This paper summarizes the eruptive histories of Etna, Vulcano, Stromboli, Vesuvio, Campi Flegrei and Ischia, outlining their styles of activity and their general mode of operation. Magmas composition and evolution are not discussed (see Peccerillo, this volume). The involvement of the scientific leaders of current conspicuous research projects on each volcano warrants both an update and completeness of the report. Hopefully this short review, even though rather inhomogeneous, should represent a reliable point of reference for both research and popular scientific articles.

Etna (R.C.)

Mount Etna, the largest active volcano in Europe, reaching 3,350 m a.s.l. and covering a surface of about 1,260 km^2, started its activity at about 600 ka, after the end of Upper Pliocene to Pleistocene subaqueous and subaerial eruptions at the northwestern edge of the Iblean Plateau. Mt. Etna is a stratovolcano, consisting of edifices centered on distinct eruptive axes, of whom the most recent ones may be still recognized. Etnean magmas rise up from a mantle source into the crust, subject to strong extensional stresses, and the volcano is actually placed at the intersection of two major regional fault systems, trending respectively NNW-SSE (Iblean-Maltese) and NNE-SSW, along the coastline between Taormina and Messina. Data on mineral phases, fluid and melt inclusions and some seismological evidence suggest a 20 to 15 km deep plexus of magma-filled fractures and of shallower and smaller chambers, where magma resides, differentiates and eventually gives rise to the activity of the various centers which have followed each other in time (Figure 1). The oldest (600 ka) volcanics are submarine tholeiitic to transitional basalts, erupted on the sea floor (ca. 500 m depth) of a wide gulf, extending between the northern mountain chain and the Iblean Plateau to the south. They outcrop now as pillow-lavas, hyaloclastites and sills along the Ionian coast. After a strong regional uplift, subaerial tholeiites were erupted (ca. 300 ka): they currently outcrop only in the southwestern sector, but probably covered much wider areas, now buried under later volcanics. Similar lavas and Na-alkaline products are found on the sea floor, offshore of the Ionian coast. Central volcanoes, fed by Na-alkaline magmas, started to develop (200 ka) above the earliest volcanic levels. Most of these volcanoes are strongly dismantled by erosion and widely covered by younger volcanic levels; their products chiefly crop out along fault scarps or uplifted cliffs (E. flank). Within their volcanic sequence, volcaniclastic levels (pyroclastic fall and flow deposits and lahars mostly resulting from debris-flow) are interbedded with lava flows, as evidence of effusive to highly explosive Subplinian to Plinian activity.

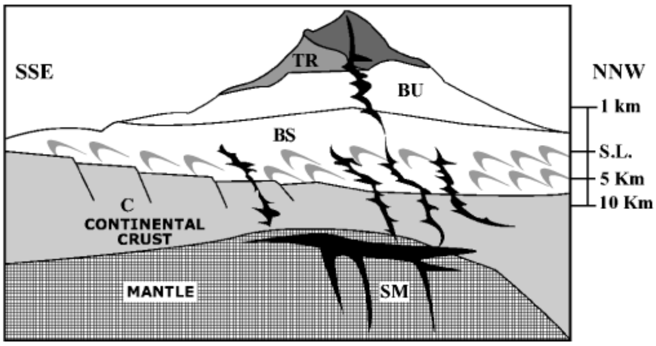

Figure 1 Sketch NNW-SSE section of Mount Etna. Note the different scale of elevations above and below sea level (lm). MB: Mongibello Unit; TR: Trifoglietto Unit; UB: Basal volcanic Units (older than 80 ka); BS: sedimentary Basement Successions; SM: main magma reservoirs.

The most recent Mongibello activity (< 30 ka) was characterized by recurrent, significant explosive activity until a few thousands of years ago. Paroxysmal eruptions gave origin to calderas, still recognizable, although largely filled by younger products ("Ellittico caldera", 15 ka, ca. 4.5 km across; "Piano caldera", 122 B.C). During the last centuries explosive activity of Mongibello was quite mild, semipersistent at the summit vents to sporadic from lateral vents. Intensity of summit vent phenomena is very variable (quiet steam emission to strombolian explosions and lava fountaining), infrequently combined with small lava effusions, lasting few hours up to several months or even years. At present there are several vents in the top region (the Chasm, Western, Northeastern, and Southeastern vents), each behaving independently and suggesting a complex feeding system for their activity.

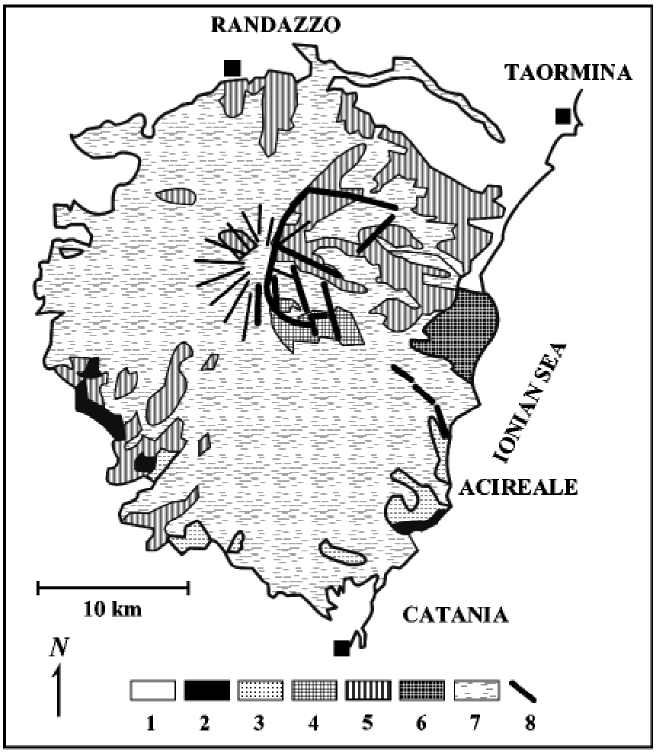

Figure 2 Sketch map showing the distribution of the main units of the Etnean volcano. 1) Sedimentary basal levels; 2) Tholeiites; 3) Ancient Na-basalts and hawaiites; 4) Trifoglietto Unit (mugearites); 5) Ellittico volcano (Mongibello Unit; hawaiites to trachytes); 6) Detrital alluvial fan from the Valle del Bove; 7) Recent Mongibello (hawaiites and mugearites); 8) Edge of the Valle del Bove and main tectonic lineaments.

Peripheral vents can open also at low elevations (down to 300 m a.s.l.), even outside the edge of the volcanic cover (Gravina di Catania, Mojo Alcantara). They mostly pour out lava flows, with tephra originating modest spatter ramparts to large cinder cones, either isolated or associated along the feeding fractures. The Summer 2001 and Winter 2002 eruptions are to be ascribed to this last type of fairly explosive events. This activity lasts from a few days to several months, exceptionally for years (Table 1); flow volumes and shapes depend on eruption duration and rate and also on flank topography. In the last 350 years, around 70 eruptions occurred, irregularly distributed in time and space. A deep horse-shoe shaped valley (Valle del Bove) carves the eastern flank of the volcano; it might have been formed by caldera collapses of ancient edifices, sliding of the seawards unbutressed volcanic mass, rapid erosion of steep flanks. Large amounts of detrital materials derived from the Valle del Bove are forming an alluvial fan in the vicinity of Giarre-Riposto (Figure 2). Recent lavas are mostly aa; less commonly pahoehoe, or have their surface covered with irregular slabs variously embricated or piled on top of each other. In these flows complicated tube systems may form, along which the thermally isolated melt can flow over great distances, feeding lava fronts as far as 10 km or more from the vents. Almost 60% of the Etnean region has been covered by at least one lava flow since the 13th century, including even some densely populated sectors at low elevations. Even if the recent activity is moderately hazardous for human life, it seriously threatens all human activities in this densely populated area, due to complete destruction in the lava flooded surface, remaining barren for centuries.

Table 1 Some data on Etna historic eruptions

Year	Flank	Duration (days)	vent m a.s.l.	fronts m a.s.l.	Lengh (km)	Area (km^2)	Volume (10^6 m^3)
1634-38	South	1224	2050	450	9.5	12.6	105
1669	South	122	825	0	16	37.5	977
1792-93	South	370	1950	600	6.5	8	80
1892	South	173	1913	970	7	10	111
1911	North	13	2310	550	7.5	6.3	65
1928	East	18	1900	25	8	5 4	0
1950-51	East	372	2530	800	10	10.5	168
1971	South	32	2965	2500	3.5	3.4	35
1971	East	36	1820	780	6.8	4.1	40
1979	East	4	2850	870	6.5	7.5	75
1981	North	7	1883	600	7.5	6	30
1983	South	131	2410	1020	7	6	70
1991-93	East	473	2420	730	8	7	250
1634-38	South	1224	2050	450	9.5	12.6	105

Vulcano (L.L.)

The Island of Vulcano (22 km^2) represents the top of a much larger structure, which has the base at a depth of 1000 m bsl. Its evolution results from six main stages of volcanic activity related to the formation of different structures: Primordial Vulcano, Caldera del Piano, Lentia Complex, Caldera di La Fossa, La Fossa, Vulcanello (Figure 3). The Primordial Vulcano is a truncated composite cone formed between 120 and 100 ka. Most products have basaltic to shoshonitic composition. Renewal of activity occurred at different times on its flanks, between 40 and 20 ka. The summit of the cone collapsed between 98 and 78 ka, leading to the formation of the Caldera del Piano, which was progressively infilled by lava flows probably outpoured along the ring faults. Several pyroclastic units erupted from local vents were deposited within the caldera between 77 and 21 ka. After 50 ka the volcano-tectonic activity shifted towards NW, culminating with the formation of the southern sector of La Fossa Caldera. Between 24 and 13 ka intermediate to rhyolitic magmas outpoured in the northern part of the island forming a series of lava domes that formed the Lentia Complex. A large part of this structure collapsed about 15 ka ago, leading to the formation of the western sector of La Fossa Caldera, probably related to the largest explosive eruption occurring at Vulcano and which formed the Tufi di Grotte dei Rossi deposits.

The composite tuff cone of La Fossa began its activity about 6 ka within the caldera. The last 1888–90 eruption, which is the reference type of the vulcanian-type eruptions, produced a small blanket of coarse ash and the famous bread crust bombs. A significant part of the products fell back into the vent, leading to a 120 m accumulation of tephra inside the crater. Intermediate to evolved compositions dominate among La Fossa rocks. Minor shoshonitic and latitic products occur. Vulcanello, whose latest eruption was in the 16th century, is the youngest (< 2 ka) structure of the island: it consists of a multiple, mostly shoshonitic, lava-flow platform and three ENE-SSE nested tuff cones.

At present fumarolic activity occurs both on the flanks and in the pericrateric area of La Fossa cone with a maximum 2002–2003 temperature of about 400°C. It is worth noting that in 1992–1993 a temperature of about 700°C was reached.

The volcanic risk at La Fossa is mainly related to the high concentration of the population, reaching 15,000 during the summer at Vulcano Porto village, at the foot of the cone. The prediction of the kind of future events has been derived through detailed stratigraphic studies. The volcanic history consists of five "Eruptive Successions" separated by erosional unconformities. Each Succession includes (Figure 4) a series of Eruptive Units (up to a total of 15), distinguished by the lithological features and dispersal pattern of products, related to dry surge, wet surge, pyroclastic fall, lava flows and lahars. The erupted volume decreases with time, the most voluminous being the Punte Nere Succession, forming the main structure of La Fossa cone. The quiescence period between Successions is variable with a maximum rest of about 800 years between 1st and 2nd Successions. The present 113 year rest represents a true quiescence period. The length of quiescence between eruptions does

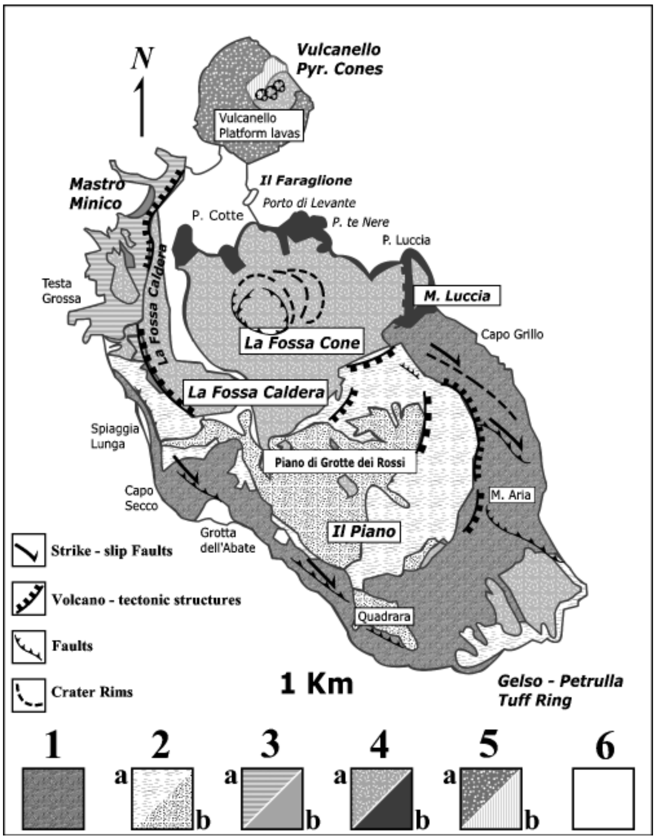

Figure 3 Sketch map of Vulcano Is. 1. Primordial Vulcano; 2. Piano Caldera infilling products; 3. Mastro Minico–Lentia Complex: 3a. rhyolitic lavas, 3b. latitic and trachytic lavas; 4. La Fossa Caldera and La Fossa cone products: 4a. pyroclastics, 4b. lava flows; Vulcanello Cones: 5a. leucite-bearing potassic lavas, 5b. trachytic lava flow; 6. Beach, alluvial and detritial deposits.

not influence the eruptive style and magnitude of eruption; this is probably related to the fact that the magmatic processes at La Fossa occur in an open system where the magma continuously degasses, therefore preventing magmatic gas accumulation. The dynamics of the eruption is mostly related to magma-water interaction.

The probable scenario of a future eruption could be the following: 1) tectonic event brings the hot primitive magma from the deeper reservoir into contact with the more evolved shallower one; 2) the hydrothermal system is triggered and the upper volcanic system fractured, inducing craterization processes; 3) the effective contact of the uprising magma with water produces a hydromagmatic eruption generating dry and wet surges whose deposits form a tuff cone. During the eruption minor magmatic processes can occur, forming lapilli and bomb deposits (vulcanian phase) and, if the eruption rate is sufficient, lava flow emission can close the eruption. In this scenario the most dangerous phenomena are the surge events that represent the most distructive potential. If we take into account the 15 Volcanic Units emplaced during the whole history of La Fossa cone, the probability that the eruption will start with base surge events is 64%, 14% for strombolian and effusive activities and 7% for a vulcanian eruption. Using the dry surge eruptions occurring after the Punte Nere Succession as a reference, in 50% of the cases the surge clouds overrode the morphologic barrier of Caldera della Fossa, in 75% they reached the walls of the caldera and Vulcanello. In 100% of the cases they affected the area close to the foot of the cone where there is the village of Vulcano Porto. On the basis of a sedimentological model based on the dry surge deposits from Palizzi Succession, in the area of Porto a dynamic pressure of about 4 kPa is expected.

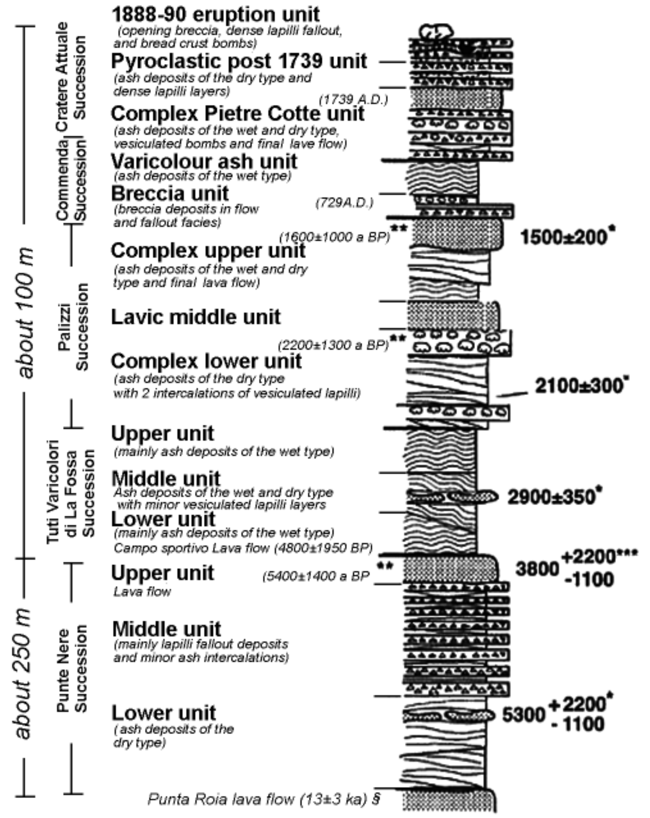

Figure 4 Composite columnar section of La Fossa cone.

Stromboli (M.R.)

Stromboli is the northernmost island of the Aelolian archipelago located about 60 km off the coast of Calabria. The island is about 12.2 km^2 in area and consists of an almost regular cone with steep slopes which rises from a depth of 1500–2000 m reaching an elevation of 924 m in the peak of Vancori. About 1.5 km off the northeast of the main island there is a small rock, Strombolicchio, which represents the eroded neck of an older central volcano. The main structural feature of the island is a large horseshoe-shaped depression (Sciara del Fuoco) bounded by cliffs several hundreds meters high which occupy the NW flank of the cone. The active craters are located at an elevation of about 700 m within the Sciara del Fuoco.

All the rocks of Stromboli are volcanic in origin and probably formed within the last 100 ka. They consist of subaerial lava flows, pyroclastic materials and subordinate volcaniclastic sediments. Dykes and sills are exposed on the eroded slopes of the volcano. Products of Stromboli have been referred to cycles which are separated by important structural events. The products of the old activity of the main island consists of lavas and pyroclastics with calc-alkaline affinity and age of about 100 ka. The lava plug of Strombolicchio is also calc-alkaline with an even older 200 ka. The post-100 ka evolution of the volcano has been subdivided into four periods (Figure 5): Paleostromboli (100–35 ka), Vancori (?25–13 ka), Neostromboli (13–5 ka), recent Stromboli (5.0 ka-present). Products of Neostromboli consist of K-rich basalts often bearing small phenocrysts of leucite (Lc-shoshonites). Lavas and pyroclastics of the Vancori period are evolved andesites of the high potassium calc-alkaline suite.

The morphostructural evolution of Stromboli has been dominated by caldera collapses and gravitational failure of its flanks. Caldera structures have been inferred for the period older than 13 ka. Since then Stromboli has produced at least three major collapses of the northwest flank of the cone. The older one which affected the Vancori edifice had an estimated volume of 2–3 km^3. The following

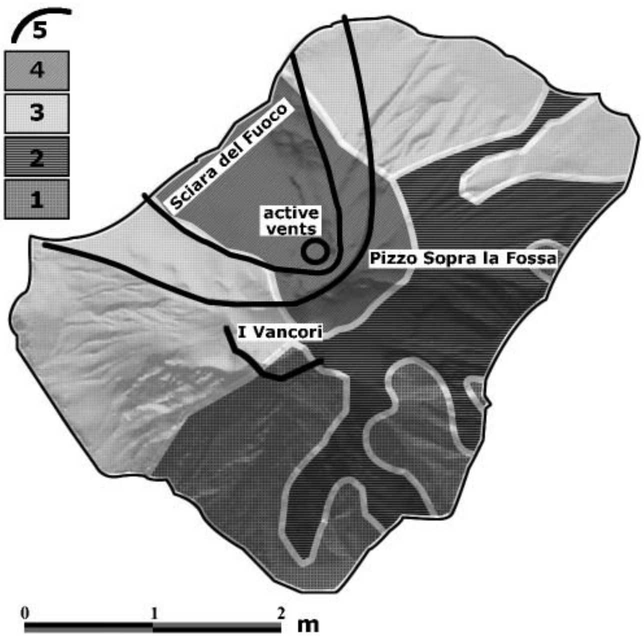

Figure 5 Sketch of Stromboli Island showing the areas covered by products of different periods of activity: 1. Paleo-stromboli (100–35 ka); 2. Vancori (?25–13 ka); 3. Neostromboli (13–5 ka); 4. recent Stromboli (5.0 ka–present); 5. flank collapses

collapse episode (volume 1–1.5 km^3) beheaded a 900 m high cone whose summit was situated above the present crater area. After this event a cone of about 1,000 m high was formed (Pizzo Sopra la Fossa). The Pizzo Sopra la Fossa edifice was in turn largely dismantled by another gravitational failure (The Sciara del Fuoco collapse) during which a volume of about 1.08 km^3 of volcanic material slid into the sea.

The typical activity of Stromboli consists of intermittent mild explosions ejecting scoriaceous bombs, lapilli and ash from vents where glowing lava stands at high level in an open conduit. The explosions occur when large bubbles of compressed gas bursts at the surface of the magma column resulting in the formation of a jet of hot gas and incandescent lava fragments. The explosions last a few seconds and take place at regular intervals, the most common time interval being 10–20 min. Although the volcanic activity of Stromboli has been known since the Classic Age, historical sources older than 1000 A.D. are not accurate enough to assess if the activity was exactly the same as we see today. Explosive activity is associated with continuous streaming of gas with an estimated output of 6,000–12,000 t/day and consisting mainly of H_20 (3,200–6,300 t/day), CO_2 (2,900–5,800 t/day), SO_2 (400-800 t/day) and minor HCl and HF. The routine activity of the volcano is periodically interrupted by lava flows within the Sciara del Fuoco and more violent explosions. Mid scale explosions consist of short-lived, cannon-like blasts that eject meter-sized spatter and blocks within a distance of several hundreds of meters from the craters. On average two or three explosions per year occurred over the past 100 years. Less frequently much more violent explosions occur. They produce showers of incandescent scoriae and spatter within a distance of several kilometers from the craters sometimes affecting the two villages on the coast (Stromboli and Ginostra).

Volcanic hazards of Stromboli are fallout of heavy pyroclastics (blocks and bombs) launched by cannon-like explosions, hot avalanches and tsunami. Fallout of ballistics represent the main hazard to people who climb the volcano to observe the persistent activity, at times (April 2003) reaching inhabited areas. As in December 2002 related to effusive fracturing, moderate tsunamis could be generated by subaqueous landlides of part of the Sciara del Fuoco slope. They severely treated the main village of Stromboli situated on low land along the northeastern coast of the island.

Somma-Vesuvio (R.S)

The Somma-Vesuvio volcanic complex consists of an older volcano dissected by a summit caldera, Mt. Somma, and a recent cone, Vesuvio, that grew within the caldera after the AD 79 “Pompeii” eruption. The volcano is relatively young: the Somma stratocone most probably postdates the 39-ka old Campanian Ignimbrite eruption of Campi Flegrei and stopped activity at about 20 ka. It consists of a pile of thin lava flows interbedded with spatter and cinder deposits (K-basalt-trachybasalt to K-tephrite-phonotephrite). About 18 ka ago the style of activity changed: after a long quiescence a large Plinian eruption (“Pomici di Base”) ejected K-trachytic to K-latitic magmas. In the following 16 ka three other Plinian eruptions occurred, each preceded by long rest periods: “Mercato” (8 ka, K-phonolite), “Avellino” (3.4 ka), and “Pompeii” (AD 79), both K-phonolite-tephriphonolite, while effusive activity was limited to a few lava flows from lateral vents (approx. 17 ka). In between the four major Plinian events, 8–10 minor explosive eruptions occurred, Subplinian to Vulcanian in style. After AD 79, the recent cone began to form. It grew discontinuously during periods of persistent Strombolian and effusive activity occurred in the Ist–IIIrd, V–VIIIth (after AD 472 “Pollena” eruption), X–XIIth centuries and in 1631–1944. A dozen explosive eruptions alternated with the open conduit activity, each preceded by long (100–300 years) rest. The largest eruptions of this last period of activity occurred in AD 472 and 1631 and were Subplinian in size and dynamics.

Mt. Somma is a nested, polyphased caldera related to the emptying of shallow reservoirs during large eruptions. Four caldera-forming events have been recognized (Figure 6), occurred during Plinian eruptions. The structural collapses constantly accompanied the final phreatomagmatic phases common to all four events. The

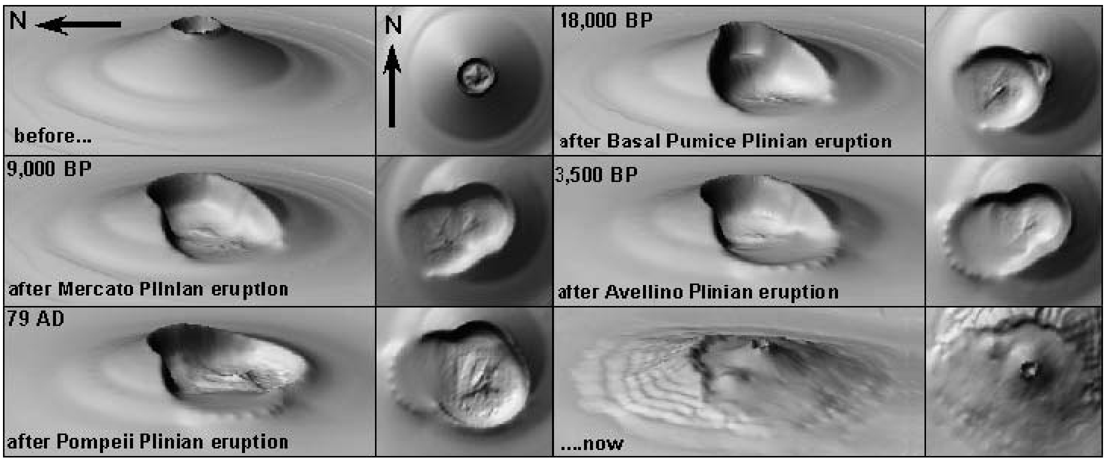

Figure 6 Schematic reconstruction of the morphological evolution of SV caldera.

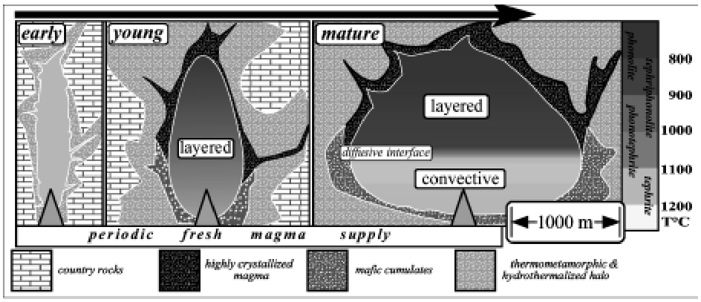

Figure 7 Highly speculative sketch of the evolution of Vesuvio magma chambers growing following the periodic arrival of periodic tephritic inputs (modified after Cioni et al., 1997).

geometry of the caldera did not suffer significant modifications related to the interplinian volcanic activity until the post AD 79 "reconstructing" period, whose products covered and mostly obliterated the seaward lower rim of the caldera. The eruptive history of Vesuvio reflects a plumbing system characterized by the constant presence of shallow magma chambers and alternating periods of open and obstructed conduit conditions. The chambers were supplied by deep, mafic magma batches investigated through melt inclusions in high-T crystals (mostly olivine and diopside). These revealed the not truly primitive nature of the melts entering the chamber, as well as a change from K-basalt to K-tephrite occurred between the Avellino and Pompeii eruptions. When the conduit is open, the reservoir is continuously tapped through persistent Strombolian activity. The periodic arrival of fresh magma in the full plumbing system results in either quiet lava effusions and moderate growth of the magma chamber (<3 km depth) or in explosive-effusive polyphased eruptions whose dynamics induce the complete emptying of the reservoir. After short quiescent periods (reflecting the system recharge), Strombolian conditions are restored, initiating a new cycle. When the conduit is obstructed the magma chamber grows until, after quiescent intervals of variable length, a explosive eruption is initiated. The increasing volume (Figure 7) is accompanied by changes in the aspect ratio of the chamber as well as in the compositional layering: (1) initial stage, high aspect ratio chamber, moderate volume (0.01–0.1 km^3), almost homogeneous mafic melt; (2) young stage, medium aspect ratio chamber, medium volume (0.1–0.5 km^3), almost continuous gradient from mildly evolved to felsic melt; (3) mature stage, low aspect ratio chamber, large volume (0.5–5 km^3), two fold layering with stepwise gradient separating a lower, convective, mildly evolved portion from an upper, statically stratified, felsic portion.

The present 60-year long quiescence departs from the pattern of open conduit conditions and after 1944 eruption the conduit remained obstructed. The volume of magma entered the chamber since then coud be in the order of 2×108 m^3. If totally ejected during a single explosive event, it could result into a Subplinian eruption, similar to the last of this kind (AD 1631). Such an eruption has been therefore taken as reference event for the presently Maximum Expected Eruption whose scenario, from field and historical data, was included in the Emergency Plan established by the Civil Defense.

Campi Flegrei (G.O.)

The Campi Flegrei caldera (CFc) is a nested and resurgent structure (Figure 8) resulting from two major collapses related to the Campanian Ignimbrite (CI; 39 ka) and Neapolitan Yellow Tuff (NYT; 15 ka) eruptions (Figure 8a). Rocks older than CI are only exposed along sea cliffs and high-angle scarps related to the CI caldera collapse. The oldest detected age on these rocks is of about 60 ka. The CI eruption, the largest of the Mediterranean area over the past 200 ka, extruded not less than 200 km^3 of trachytic to phonolitic-trachytic magma. The caldera collapse affected the area which at present includes the Campi Flegrei, Naples, the bay of Pozzuoli and part of the bay of Naples. A coincidence has been pointed out between the eruption and the bio-cultural modifications in Old World prehistory, including the Middle to Upper Palaeolithic cultural transition and the supposed change from Neanderthal to "modern" *Homo sapiens* anatomy, a subject still debated in the literature. The subsequent volcanism was concentrated within the CI caldera. The NYT phreatoplinian eruption, the second largest of the Campanian area, extruded about 40 km^3 of magma (alkali-trachyte to latite). The caldera related to this eruption was nested within the CI caldera, centered on the present Campi Flegrei. Volcanism of the past 15 ka has concentrated in three "Epochs of Activity" separated by quiescence (Figure 9). It has generated mostly explosive eruptions, variable in magnitude and generally characterised by alternating magmatic and phreatomagmatic explosions. During each Epoch, eruptions occurred at intervals of about 60 years, on average. During the I Epoch (12.0–9.5 ka), out of 34 explosive eruptions, only the Pomici Principali (10.3 ka) was a high-magnitude event. The II Epoch (8.6–8.2 ka) generated 6 low-magnitude explosive eruptions. The III

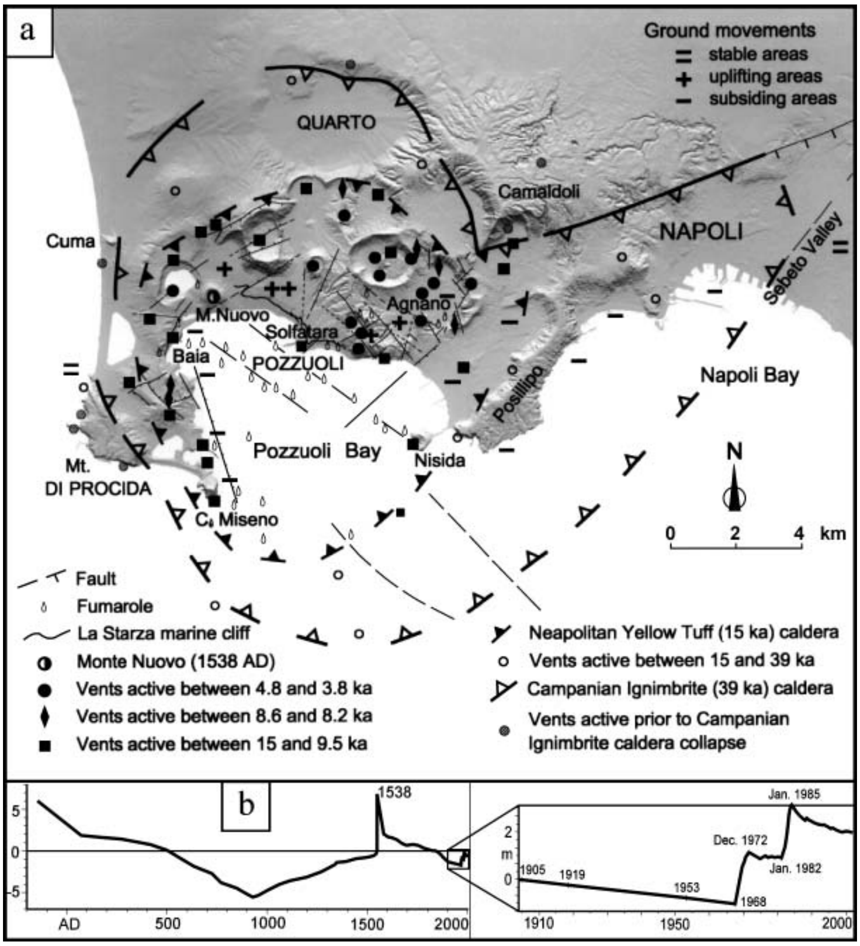

Figure 8 a) Structural map of the Campi Flegrei caldera; b) Vertical ground movements at Serapis Roman market in Pozzuoli.

Epoch (4.8 – 3.8 ka) produced 16 explosive and 4 effusive eruptions. During this Epoch the only high-magnitude event was the Agnano-Monte Spina eruption (4.1 ka). The first two periods of quiescence lasted 1.0 and 3.5 ka, respectively, while the last, begun at the end of the III Epoch, was interrupted in 1538 AD by the Monte Nuovo eruption, the last. Volcanism and quiescence are strictly related to formation and deformation of the caldera. During the I and II Epochs, magma erupted through the marginal faults of the NYT caldera. Between the II and III Epochs, a change in the stress regime occurred in the caldera. Before onset of the III Epoch, the La Starza block, which had been uplifted at variable resurgence rates with alternating periods of emersion and submersion, definitively emerged. During the III Epoch, magma was able to reach the surface almost only along the faults of the sector of the resurgent block under tensional stress regime. The whole CFc is subsiding, while the central part of the NYT caldera has been affected by resurgence since at least the second period of quiescence (Paleosol B in Figure 9). Resurgence occurs through a simple-shearing mechanism which has disjointed the NYT caldera floor in blocks (long-term deformation). The most uplifted block includes the La Starza marine terrace and has been displaced by about 90 m. Vertical ground movement has been well documented for the past 2000 years (Figure 8b). In the past 40 years, unrest episodes have affected the caldera in 1969–72, 1982–85, 1989, 1994 and 2000 and have generated uplifts of 170, 180, 7, 1, and 4 cm, respectively. The deformation has been interpreted as the result of ductile (expansion and deflation of the geothermal system) and brittle (fracturing of the magma chamber roof rocks) components, both generated by increase in pressure and temperature within the magma reservoir due to arrival of small magma batches, less evolved and hotter than the resident. The area deformed during the unrest episodes has a polygonal shape and its boundaries correspond to the structures bordering the resurgent block, suggesting that long-term deformation results from the summation of many short-term deformational events.

The magmatic system of the CFc includes a shallow, large-volume trachytic reservoir periodically refilled by new magma batches rising from a storage zone located between depths of 10 and 15 km. The shallow reservoir has been the site of differentiation processes. From 60 to 44 ka, the reservoir was growing due to input of new magma batches, while from 44 to 39 ka, it was an isotopically homogeneous, large-volume, zoned system, whose evolution culminated in the CI eruption. Arrival of new magma batches formed an apparently independent, large-volume reservoir which fed the NYT eruption. In the past 15 ka, three isotopically and geochemically distinct magmatic components were erupted as either homogeneous or mixed magma batches. One component is similar to the CI trachytic magma, the second is similar to the NYT latitic-alkalitrachytic magma, the third is a trachybasalt never erupted before. It has been hypothesized that the CI and NYT components represent residual portions of older, large-volume magma reservoirs which have fed eruptions since about 60 and 15 ka, respectively. The least-evolved component, erupted through vents located on a NE-SW regional fault system, likely represents the deeper seated magma tapped by regional faults. The persistent state of activity of the caldera and its intense urbanization make the volcanic risk very high. To mitigate such high risk, an emergency plan is in preparation by the Department of Civil Defense. Presently the area at highest risk, that is the one which could be affected by pyroclastic currents (Red Zone), has been defined and evacuation of the population (350,000 people) before the beginning of the eruption, is planned.

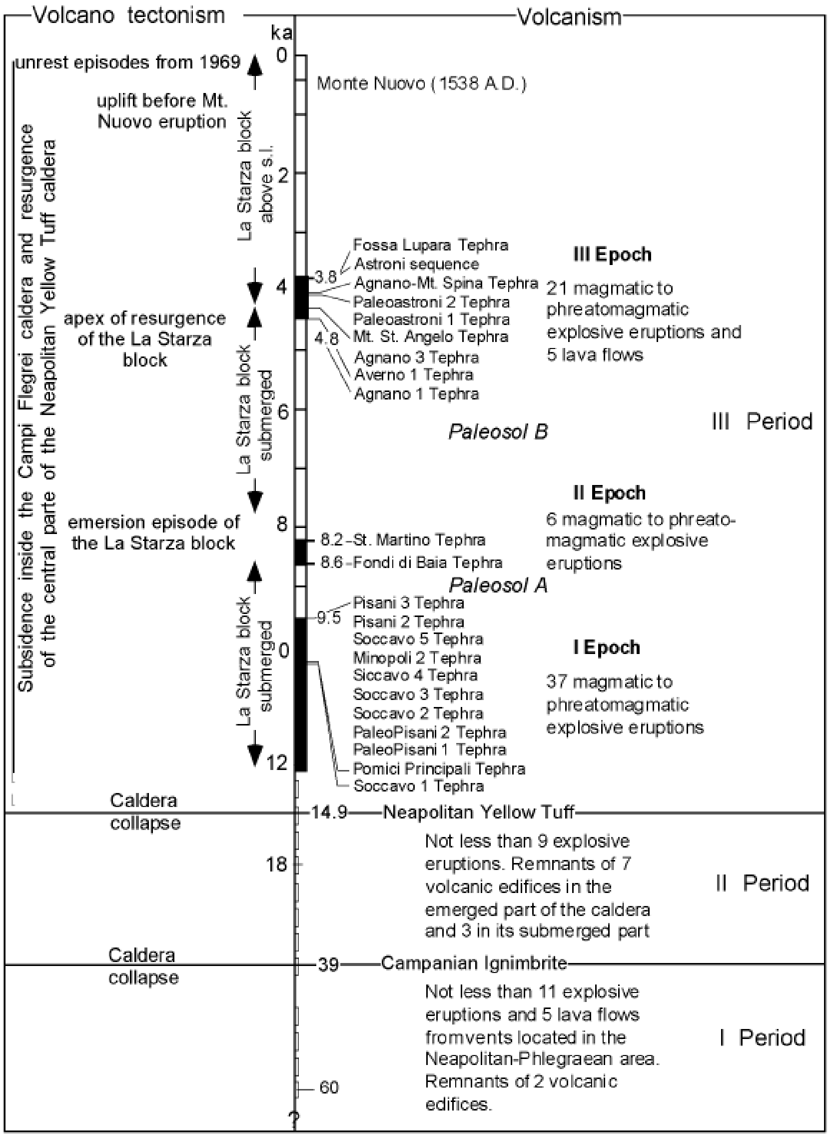

Figure 9 Chronogram of volcanic and deformational history of the Campi Flegrei caldera.

Ischia (G.O.)

The island of Ischia is a volcanic field (Figure 10) composed of volcanic rocks, landslide deposits and subordinate terrigenous sediments. The volcanic rocks range in composition from trachybasalt to latite, trachyte and phonolite; the most abundant are trachyte and alkalitrachyte. The volcano is located at the intersection of NE-SW and NW-SE regional fault systems. A caldera collapse accompanied the Mt. Epomeo Green Tuff (MEGT) eruption and was later deformed by a simple-shearing block resurgence, which has generated a net uplift of the Mt. Epomeo block by about 900 m over the past 30 ka. The beginning of volcanism on the island is not precisely known. The oldest exposed rocks, not dated, are the remnant of a complex volcanic edifice. The volcanism which followed the construction of this volcano (150–74 ka, Figure 11) produced mainly trachytic and phonolitic lava domes and subordinately alkalitrachytic pyroclastic deposits. A long quiescence followed, interrupted at about 55 ka by the caldera-forming MEGT eruption. The volcanological and magmatological history of Ischia in the past 55 ka has been subdivided in three periods of activity. The first period, initiated with the MEGT eruption, continued up to 33 ka with hydromagmatic and magmatic explosive eruptions. An early trachytic magma was followed by trachytic-to-alkalitrachytic magmas with increasing

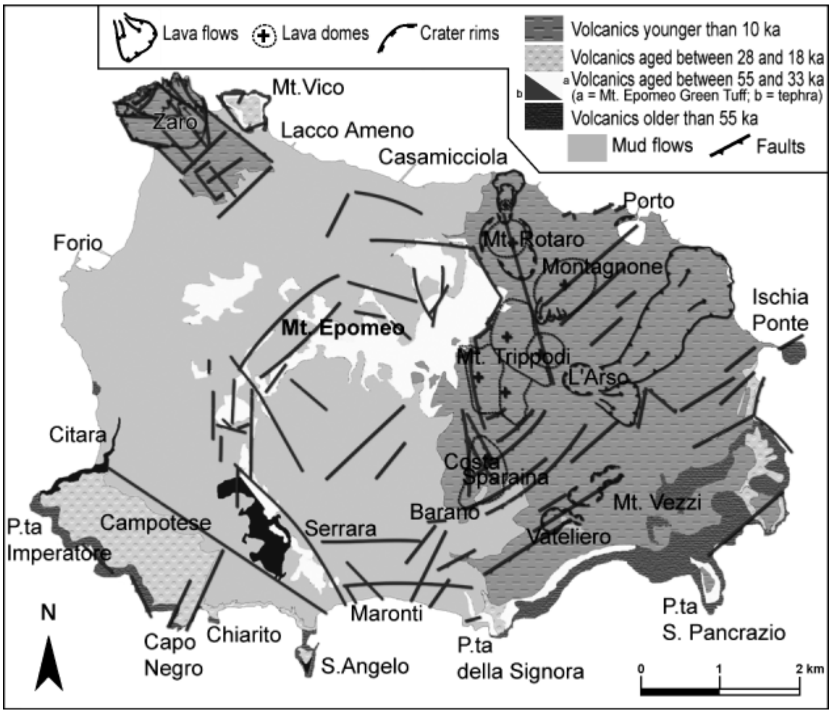

Figure 10 Geological sketch map of Ischia.

degree of differentiation and constant isotopic composition through time, suggesting that the magma chamber was differentiating mostly through fractional crystallisation processes. The trachybasaltic eruption of Grotta di Terra, at about 28 ka, marked the beginning of the second period, which continued until 18 ka with sporadic explosive (alkalitrachytic) and effusive (trachytic) eruptions. The erupted magmas varied through time from trachybasalt to alkalitrachyte with increasing incompatible elements and Sr isotope ratio, suggesting arrival of new magma, progressive differentiation and mixing with the resident alkalitrachytic magma. During the last period, which began at about 10 ka, volcanism was mostly concentrated within the eastern portion of the island, where normal faults were generated in response to the extensional stress regime induced by resurgence.

Lava effusions were followed by phreatomagmatic and magmatic explosive eruptions. Reactivation of regional faults, likely determined the reappraisal of volcanism also in the northwestern corner of the island, outside the resurgent area, at about 6.0 ka. After a period of quiescence, volcanism resumed again in the eastern portion of the island at about 5.5 ka (Figure 11). The following repose was interrupted at about 2.9 ka by a very intense phase of activity (35 eruptions) which ended in 1302 AD with the last eruption on the island. A decrease in Sr isotope ratio of the magma erupted at the beginning of this period suggests the arrival of a geochemically distinct magma into the system. Mostly trachytic and subordinately latitic magmas were erupted during this period. Compositional variations, and isotopic and mineralogical disequilibria suggest mixing among compositionally different magmas.

Although Ischia is still an active volcano and home for 50,000 people, no risk mitigation action have been planned as yet.

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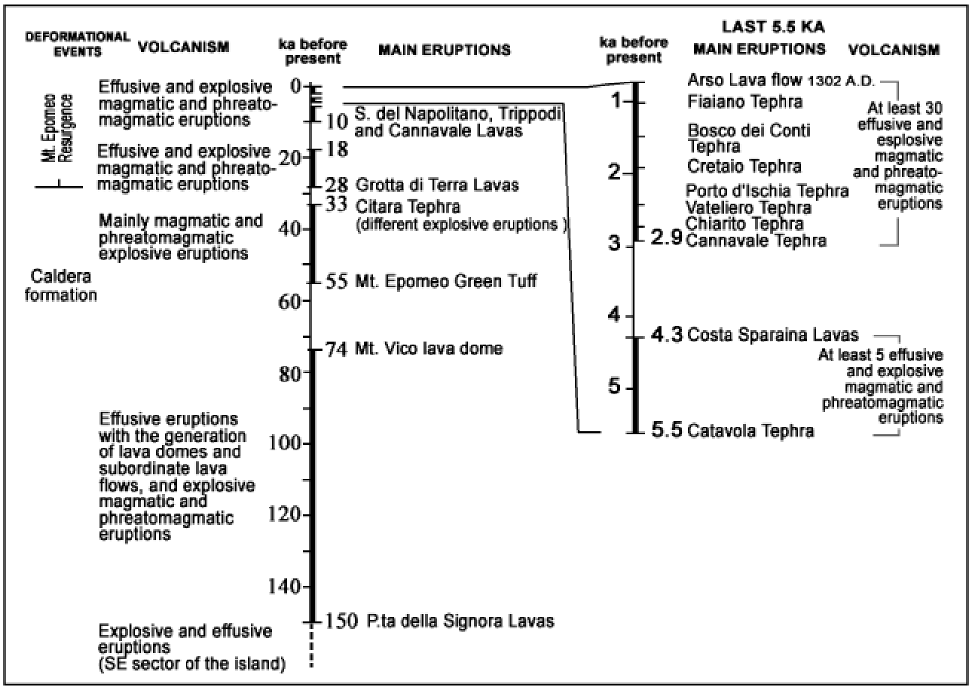

Figure 11 Chronogram of volcanic and deformational history of Ischia. (After de Vita et al., 2003)

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