Introduction: The face of Italy as source of inspiration in the geological sciences

So vast is Italy's collection of geological forms, structures and phenomena that they have attracted the attention of naturalists since ancient times. Indeed, such have been the discoveries and contributions made in Italy that Charles Lyell was motivated to write in his *Principles of Geology* that the Italians "preceded the naturalists of other countries in their investigations into the ancient history of the earth". He thought that they still maintained a decided pre-eminence.

With its position in the centre of the Mediterranean, Italy has always been a preferred destination for geologists coming from different parts of the world, attracted by the extraordinary range of geological situations present in the country. Experience with Italian outcrops and landscapes has been an important element in the development of scientific concepts in general.

The varied geology of Italy results from a long history that has produced a unique geodynamic setting, distinguished by the presence of bivergent Alpidic fold belts with two thrust systems directed respectively towards the European and Adria-African forelands, and their associated centres of igneous activity. The opposing thrust fronts of the two mountain chains, the Alps and Apennines, are characterized by a series of orogenic arcs whose convexity faces their respective forelands and whose concave faces are occupied by internal collapse basins (e.g., the Tyrrhenian Sea).

Great precursors. Two thousand years ago, the Latin poet Ovid (43 BC–AD 17) in his great hexametric poem *Metamorphoses* noted the importance of the continual mutation of biological forms, of the earth's surface geography and of its geology. In his verses (XV, 262–265) he sketched out a preliminary model of sedimentary cycles:

"Vidi ego, quod fuerat quondam solidissima tellus, / esse fretum, vidi factas ex aequore terras, / et procul a pelago conchae iacuere marinae, / et vetus inventa est in montibus ancora summis."

("I have myself seen what was once most solid ground disappear into the sea, and have heard of land risen out of the sea; marine shells lay far from the sea, and an ancient anchor rested on top of a mountain.")

The Florentine poet Giovanni Boccaccio (1313–1375) wrote in 1340 that fossil shells in the Tuscany hills had formerly lived in the sea that once covered the land. However, the first person to recognize the organic nature of fossils was the Tuscan monk Ristoro d'Arezzo (1239–1282) whose encyclopaedic manuscript *La composizione del mondo* (The composition of the world) appeared in 1282. This great naturalist also understood that flowing waters erode the valleys, transport loose materials and deposit them in depressions that then become alluvial plains.

The precursor of the modern science of geology, and of modern methods of scientific inquiry, can fairly be said to have been the incomparable Tuscan genius and polymath Leonardo da Vinci (1452–1519) who collected many penetrating and perspicacious observations that enabled him to acquire a coherent, synthetic vision of geological phenomena, which he expressed in his notes, diagrams, doodles, designs and objective drawings of natural phenomena, which were originally written for his own benefit and not for immediate publication, as well as in his paintings. Unfortunately, the lack of publication of his works in any coherent form prevented his extraordinary contribution to geology from becoming known to his contemporaries and immediate successors. Indeed, some of the Leonardian codexes have become familiar only in recent times. He was the first to recognize that terrestrial landscapes are modelled by fluvial erosion and was well aware that the fossil shells found on the highest Apennine peaks are the remains of living organisms from the sea, which once inundated large parts of northern Italy that were later uplifted to form the mountains. He was also the first to demonstrate the ability of present-day causes to explain the origins of geological phenomena and to coin the term "folding" with respect to rocks structure. Could the enigmatic smile of the Monna Lisa, backed by a landscape sculpted by erosion, reflect the painter's recognition of the importance of stream channels in the formation of landscapes? In another celebrated painting, *The Virgin and Infant Jesus with Saint Anne*, which hangs in the Louvre, Leonardo depicted with extraordinary precision two turbidite sequences (see G.B. Vai).

The cooling of the Earth, the oldest hypothesis of terrestrial dynamics, was proposed towards the end of the 16th century by Giordano Bruno (1548–1600), a natural philosopher who was burned at the stake in Rome by the Inquisition, perhaps also for having insisted that the Earth orbits around the sun. He also had the temerity to associate the seasons with the Earth's rotation (1584).

It is thus not surprising that in a manuscript of 1603 the first person to use the term "geology" in a scientific sense was the Bolognese naturalist Ulisse Aldrovandi (1522–1605).

The Danish naturalist Niels Steensen (1638–1686), whose name was Latinized to Nicholaus Steno and Italianized to Niccolò Stenone, was appointed physician to the Grand-Duke of Tuscany, Ferdinand II. In but a few years, 1666–1669, this brilliant man brought his researches into the origins of the Tuscan landscape to fruition with the publication of his *Prodromus* (1669), literally "Introduction to a Dissertation", a concise work of 76 pages and a masterpiece of inductive logic. In this great milestone in the history of geology, he set down the foundations of modern geological theory on the basis of his field observations. He enunciated three basic stratigraphical principles—superposition, original horizontality and original lateral continuity of beds—which enabled a relative sequence of geological events to be worked out. In 1988, Steno was beatified by the Catholic Church for his work as an apostolate missionary and Bishop in northern Germany, Denmark and Norway during the years after his great geological discoveries. He lies buried in the church of San Lorenzo in Florence.

In 1760, the term "geology" was used for the first time in its modern sense by Giovanni Arduino (1714–1795) of Verona, an inspired fieldworker and one of the leading pioneers of modern geology. In 1759, after a long period of study in Tuscany, Vicenza and Verona, he set down the fundamental tenets of the modern chronostratigraphical scale. A full thirty years ahead of A.G. Werner in Saxony, he was the first to subdivide Earth history into four "general and successive orders" characterized by terrains that he called *Primary, Secondary, Tertiary and Quaternary*, as the geological eras are known today. He considered each of these four main layers or orders to be composed of a large quantity of minor layers.

Another great forerunner, who anticipated the geodynamic theories propounded by James Hutton in 1785, was the Venetian Abbot Anton Lazzaro Moro (1687–1764). In 1740, he argued that physiographic relief results from expansional movements in the interior of the Earth.

If one can argue that the field of stratigraphy was largely invented in Italy, much the same can be said of the study of natural hazards, which are of course particularly active throughout the Italian peninsula. Large explosive eruptions are called *plinian* in recognition of the first surviving description of an eruption, that of Vesuvius in AD 79, as observed by Pliny the Younger. In two letters written to Tacitus sixteen years after the event, he described the death of his uncle Pliny the Elder, the famous naturalist and author of the 37-volume *Historia Naturalis*, who like many inhabitants of Pompeii and Herculaneum was mown down by pyroclastic flows.

Much later, the Milanese seismologist and vulcanologist Giuseppe Mercalli (1850–1914) systematically described earthquake damage, using what came to be known as the Mercalli scale, with its ten categories of increasing intensity. It is also worth remembering the work of the pioneer Macedonio Melloni (1798–1854), who demonstrated in 1856 that rocks become permanently magnetized when they cool down. To cap it all, the world's first vulcanological observatory was founded in 1841 on the flanks of Mount Vesuvius by order of the Bourbon King of Naples, Ferdinando II .

Lastly, let us not forget the work of the violinist and geologist of Parma, Roberto Mantovani (1854–1933), who was the Italian consul on the island of Réunion. In 1889 and 1909, he espoused ideas on continental displacement that were remarkably similar to those later propounded by Alfred Wegener. He believed in an expanding Earth (see G. Scalera).

This special issue comprises a selection of nineteen short, synthetic articles on Italian geology. It is merely a sample from a much larger selection that we are offering to the participants who come to Italy in August 2004 for the 32 IGC. Enjoy the articles and see you in Florence!

Forese Carlo Wezel
Editor of this special issue dedicated to
aspects of the geology of Italy, site of
the 32nd International Geological Congress

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

The Mediterranean region—a geological primer

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

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

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

Introduction

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

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

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

Overview of present-day Mediterranean geological elements

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

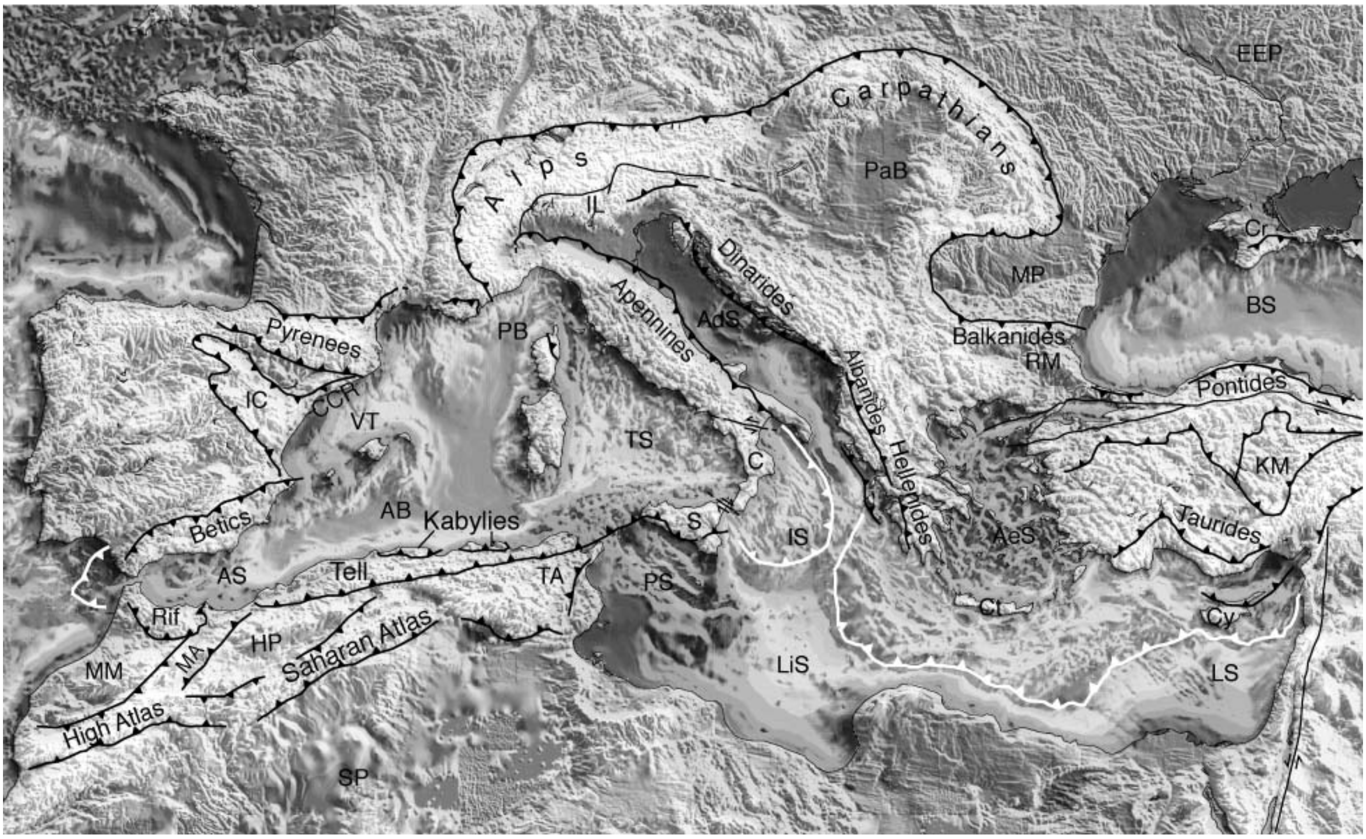

Figure 1 Digital terrain model of the Mediterranean region with major, simplified geological structures. White thrust symbols indicate the outer deformation front along the Ionian and eastern Mediterraean subduction fronts. AB, Algerian basin; AS, Alboran Sea; AdS, Adriatic Sea; AeS, Aegean Sea; BS,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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

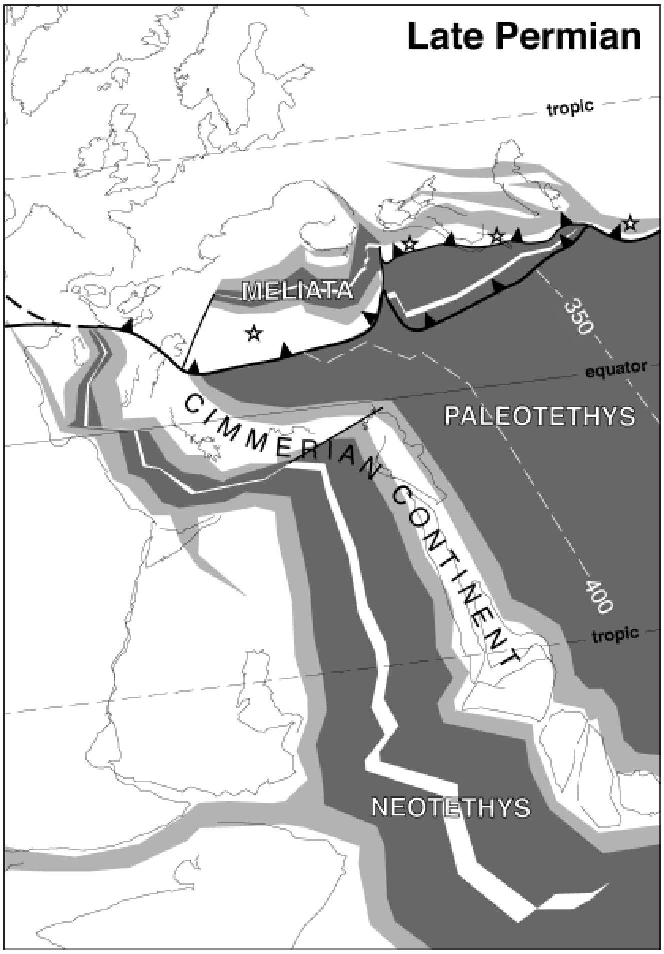

Figure 2 Paleogeographic reconstruction of the western Tethyan area during the late Permian (from Stampfli et al., 2001b, with minor modifications). Stars indicate magmatic activity.

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

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

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

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

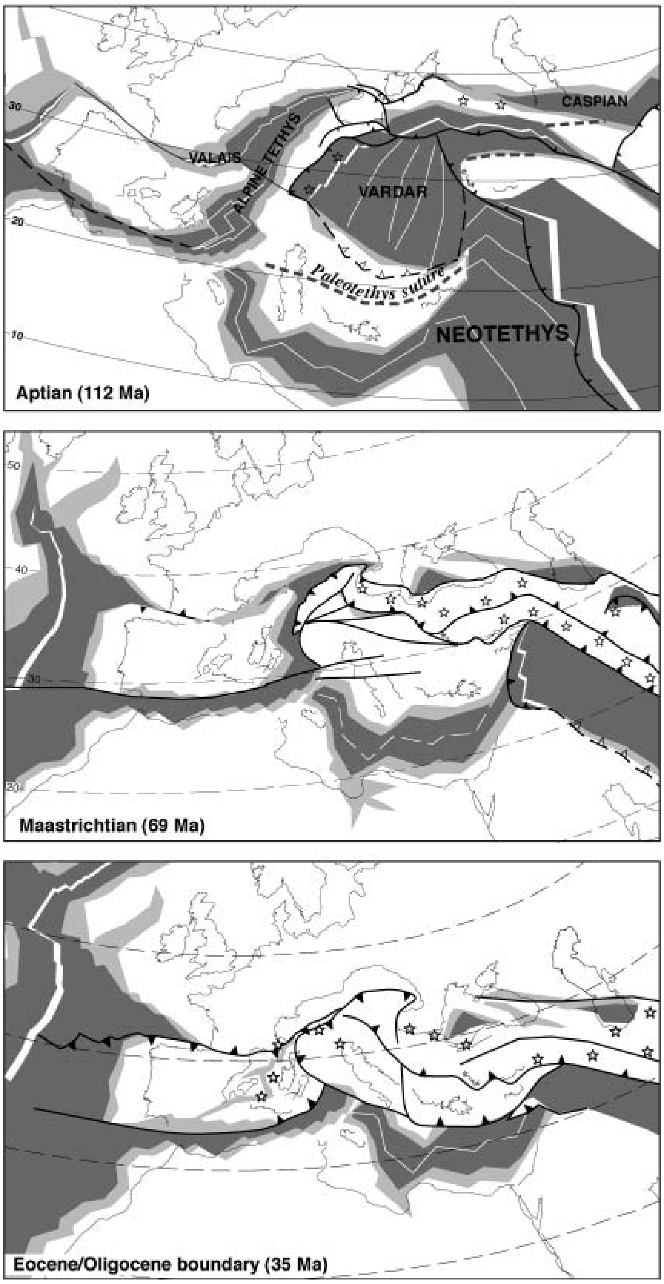

Figure 3 Paleogeographic reconstructions of the western Tethyan area during the Aptian, Maastrichtian and at the Eocene/Oligocene boundary. Note the progressive narrowing and suturing of the oceanic domains comprised between the Eurasian and Iberia continental blocks to the north and the Africa/Arabia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

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

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

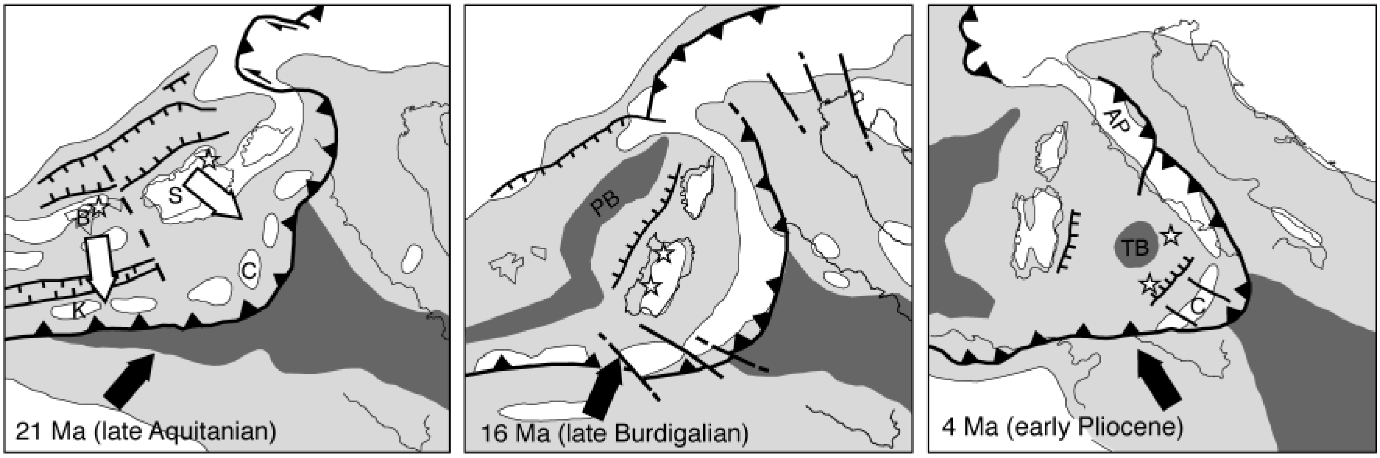

Figure 4 Schematic maps showing the paleotectonic evolution of the W Mediterranean during Neogene time (modified after Bonardi et al., 2001, and Roca, 2001). Only active tectonic elements are shown. White, exposed land; light gray, epicontinental sea; darker gray, oceanic crust. Black arrows indicate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

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

The Messinian salinity crisis

The complex Neogene geologic context of the Mediterranean region, characterized by the advanced stage of collisional coupling between the Eurasian and the African plates, is further complicated by an important episode of evaporitic deposition during Messinian (late Miocene) time. Such evaporites and—to a lesser extent—the associated post-evaporitic siliciclastics have been the focus of much attention and debate; this section summarizes some salient geological data collected at sea and on land in order to interpret the boundary conditions leading to their deposition. The literature available on this subject is abundant; only a few references are reported here.

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

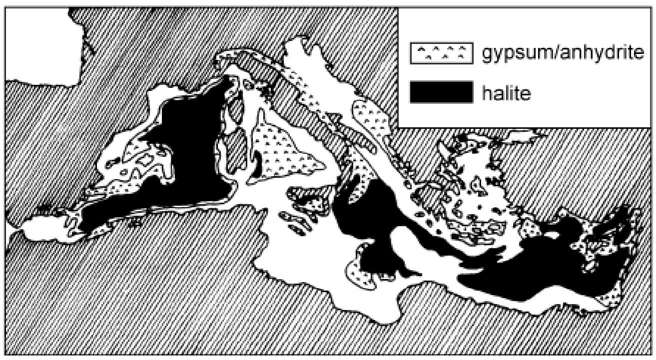

Figure 5 Areal extent of the Messinian evaporites in the Mediterranean region. Modified after Rouchy (1980).

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

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

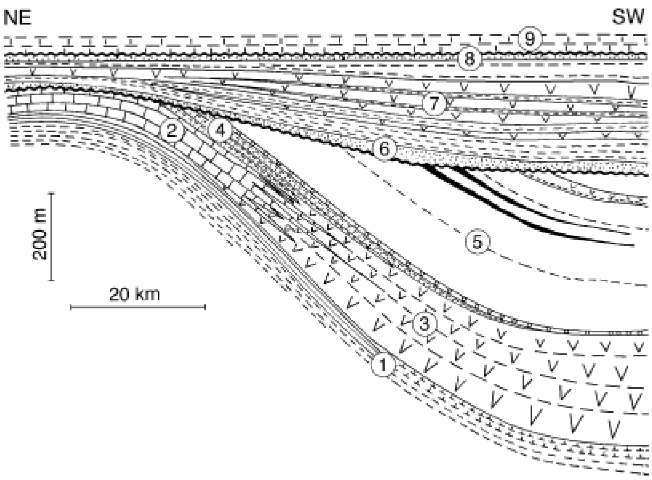

Figure 6 Schematic stratigraphic cross section of the Messinian of Sicily (modified after Decima and Wezel, 1973). 1) Pre-evaporitic clay, marl and diatomite (Tripoli Fm.); 2) evaporitic limestone (Calcare di Base); 3) lower gypsum beds (Gessi di Cattolica); 4) gypsum turbidites; 5) halite and potash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

intra-Messinian unconformity—consist only of two relatively thin units (Figure 6): the Tripoli Formation (laminated diatomites) and the Calcare di Base (evaporitic limestone). In the deepest portions of the basin, the Lower Evaporites are much thicker and comprise, from bottom to top, the Tripoli Fm, the Lower Gypsum Fm (LGF), and the Halite Fm (HF). The LGF is composed of up to 300 m of selenite gypsum with random orientation, indicating that gypsum from the periphery was reworked, deposited in deeper water, and recrystallized; its upper parts consists of gypsum turbidites. The HF is made of up to 800 m of halite with intercalations of potash/magnesium salt beds; this unit was deposited in deep depressions, fed also by clastic resedimentation and slumping. Related to intra-Messinian tectonics, slumping began when the gypsum turbidites of the LGF were deposited and reached its acme at the end of the 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".

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

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

by Giuliano F. Panza[1,2], Antonella Pontevivo[1], Giordano Chimera[1], Reneta Raykova[1], and Abdelkrim Aoudia[1,2]

The lithosphere-asthenosphere: Italy and surroundings

1 Department of Earth Sciences, University of Trieste, Via Weiss 4, 34127 - Trieste, Italy. *panza@dst.units.it*
2 The Abdus Salam International Centre for Theoretical Physics, SAND Group, Trieste, Italy.

The velocity-depth distribution of the lithosphere-asthenosphere in the Italian region and surroundings is imaged, with a lateral resolution of about 100 km, by surface wave velocity tomography and non-linear inversion. Maps of the Moho depth, of the thickness of the lithosphere and of the shear-wave velocities, down to depths of 200 km and more, are constructed. A mantle wedge, identified in the uppermost mantle along the Apennines and the Calabrian Arc, underlies the principal recent volcanoes, and partial melting can be relevant in this part of the uppermost mantle. In Calabria, a lithospheric doubling is seen, in connection with the subduction of the Ionian lithosphere. The asthenosphere is shallow in the Southern Tyrrhenian Sea. High velocity bodies, cutting the asthenosphere, outline the Adria-Ionian subduction in the Tyrrhenian Sea and the deep-reaching lithospheric root in the Western Alps. Less deep lithospheric roots are seen in the Central Apennines. The lithosphere-asthenosphere properties delineate a differentiation between the northern and the southern sectors of the Adriatic Sea, likely attesting the fragmentation of Adria.

Introduction

The first definition of the gross features of the lithosphere-asthenosphere system in Italy and surroundings dates back to Panza et al. (1980) and it is chiefly based on the analysis of Rayleigh wave dispersion. More recent models are based both on surface waves (e.g. Marquering and Snieder, 1996; Martinez et al., 1997, 2000, 2001; Ritzwoller and Levshin, 1998; Yanovskaya et al., 1998, 2000; Pasyanos et al., 2001; Karagianni et al., 2002; Pontevivo and Panza, 2002) and body waves tomography (e.g. Gobarenko, 1990; Spakman, 1990; Babuska and Plomerova, 1990; Alessandrini et al., 1995, 1997; Papazachos et al., 1995; Papazachos and Kiratzi, 1996; Cimini and De Gori, 1997; Parolai et al., 1997; Piromallo and Morelli, 1997; Bijwaard et al., 1998; Lucente et al., 1999). Based on the existing information derived both from refraction and reflection experiments, and body-wave and surface-wave tomography, a compilation of the compressional velocity (V_p), shear velocity (V_s), and density (ρ) distribution in space is due to Du et al. (1998).

We show here features of the lithosphere-asthenosphere system that characterize Italy and surroundings, with a multiscale lateral resolution, as obtained from the simultaneous inversion of regionalized surface wave tomography (e.g. Pontevivo and Panza, 2002; Panza and Pontevivo, 2002; Chimera et al. 2003) and refraction and reflection seismology data (e.g. Aljinovic and Blaskovic, 1987; Bally et al., 1986; Blundell et al., 1992; Catalano et al., 1996, 2001; Cernobori et al., 1996; Cristofolini et al., 1985; De Voggd et al., 1992; Doglioni et al., 2001; Ferrucci et al., 1991; Finetti et al., 2001; Gentile et al., 2000; Improta et al., 2000; Kissling and Spakman, 1996; Morelli, 1998; Mostaanpour, 1984; Pepe et al., 2000; Pialli et al., 1995, 1998; Scarascia and Cassinis, 1997).

Data and method

The data and methods used to obtain the tomographic maps are described by Pontevivo and Panza (2002), Panza et al. (2003a), Chimera et al. (2003), Levshin et al. (1972, 1992), Ditmar and Yanovskaya (1987) and Yanovskaya and Ditmar (1990). The tomographic maps can be discretized with a proper grid and for each cell of the grid the cellular average group or phase velocity curve is computed. The cellular dispersion curves can be grouped according to their shape and average value (e.g. Panza et al., 2003b) to define regional properties. The lateral resolving power common to most of the available surface-wave tomography (Pontevivo and Panza, 2002) is of about 200 km, but if some parameters of the uppermost part of the crust are fixed on the base of *a priori* independent geological and geophysical information, the lateral resolving power of the cellular mean dispersion curves can be improved and this justifies the choice to perform the inversion for cells of 1°×1° (Panza and Pontevivo, 2002; Panza et al., 2003a). If dispersion relations are available for periods as low as 1 sec, local studies can be performed at the scale of a few tens of km.

Due to the complexity of the area we prefer non-linear inversion, since it is independent from the initial model. Through the non-linear inversion, known as the hedgehog method (Valyus et al., 1969; Valyus, 1972; Knopoff, 1972), of the group and phase velocity curves at regional, cellular and local scale, average multiscale lithospheric models that reach a depth of about 250 km are obtained. As *a priori* information, we use the existing literature. In the inversion, the unknown Earth model is replaced by a set of parameters and the definition of the structure is reduced to the determination of the numerical values of these parameters. In the elastic approximation, the structure is modelled as a stack of N homogeneous isotropic layers, each one defined by four parameters: V_p, V_s, ρ and thickness. Each parameter can be fixed (during the inversion the parameter is held constant accordingly to independent geophysical evidences—the *a priori* information), independent (the variable parameters that can be well resolved by the data) or dependent (the parameter has a fixed relationship with an independent parameter). For each cell, a set of solutions, which are consistent with the observations and with the resolving power of the data (Knopoff and Panza, 1977; Panza 1981), is obtained.

Retrieval of multiscale structural models

In Figure 1, three examples of models of the crust and of the upper mantle are presented. In each frame, the inverted dispersion data, the set of solutions (thin lines) V_s versus depth, the explored part of the parameters space (grey area), and the chosen solution (bold line) are shown. It could be attractive to consider as solution a median of all solutions, but this is formally not correct. At the base of our choice

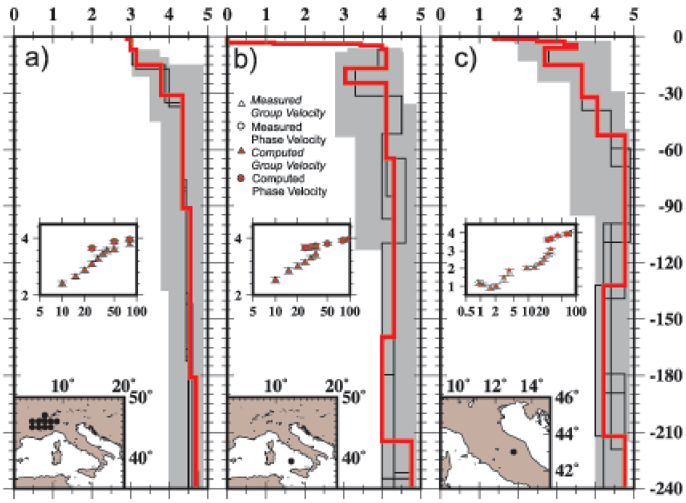

Figure 1 Models (V_S) of crust and upper mantle for: (a) Western Alps, (b) 1°×1° cell containing Vavilov Seamount, (c) part of UMD; as marked by dots. In each frame, measured (with error bars) and computed dispersion relations are given; thin lines: set of solutions; grey area: explored part of the parameters space; bold line: chosen solution.

of the representative solution there is a tenet of modern science known as Occam's razor: it is vain to do with more what can be done with fewer (Russel, 1946). To reduce the effects of the projection of possible systematic errors into the inverted model, the root mean square (r.m.s.) of the chosen solution is as close as possible to the average r.m.s. computed from all the solutions.

The set of solutions in Figure 1a corresponds to the Western Alps region defined by Panza et al. (2003b). Due to the complexity of the Alpine domain, at crustal level, the model is formally correct but has no straightforward geological significance. On the other side at mantle level, the slight increase of V_s from about 4.35 km/sec, just below the Moho, to about 4.7 km/sec, at depths larger than 180 km, is consistent with the presence of lithospheric roots, as first indicated by Panza and Mueller (1979).

In Figure 1b, the solutions correspond to the 1°×1° cell in the central area of the Southern Tyrrhenian Sea, which contains the Vavilov Seamount (Panza and Pontevivo, 2002). The Moho is very shallow (about 7 km deep) and the lid thickness is less than 10 km, with V_s about 4.1 km/sec. Below this lid, there is a very well developed low velocity layer, centred at a depth of about 20 km, with V_s about 3.0 km/sec and thickness of about 8 km. This value of V_s is consistent, accordingly with Bottinga and Steinmetz (1979), with about 10% of partial melting. The V_s just below this very low velocity layer, about 4.1 km/sec, defines the uppermost asthenosphere. In the asthenosphere, V_s varies between 4.1 km/sec and 4.3 km/sec.

In Figure 1c, the chosen structure corresponding to the Umbria-Marche geological Domain (UMD) is characterized by a layered crust, about 32 km thick, with a relatively high velocity upper and lower crust (V_s about 3.20–3.65 km/sec) separated by a low-velocity transition zone (V_s about 2.75 km/sec) about 10 km thick. The Moho is followed by a relatively low velocity layer (V_s of about 4.0 km/sec), about 20 km thick. Below this layer, a lithospheric root, with V_s about 4.75 km/sec, reaches the depth of about 130 km, which is the top of the asthenosphere, with V_s about 4.2 km/sec and about 70 km thick. The outlined V_s sequence versus depth in the uppermost mantle is consistent with the concept of mantle wedge, decoupling the crust from the underlying lithosphere. Therefore, we define mantle wedge the low velocity zone (V_s less than about 4.2 km/sec) in the uppermost mantle that overlies the high velocity lid (V_s greater than about 4.5 km/sec).

Selected cross sections

Examples of sections, crossing key areas, are given in Figure 2 (Panza and Pontevivo, 2002; Panza et al., 2003a; Chimera et al., 2003). In Figure 2b, two vertical sections trending NE-SW from the Tyrrhenian Sea across the Southern Apennines to the Dinarides are plotted. In the same sections, the shallow and intermediate-depth seismicity, with the depth error bars as given by ISC and falling in a stripe about 100 km wide and centred on the profiles, is shown.

The northernmost section AA' crosses Vavilov seamount in the Tyrrhenian Sea, Apennines, middle Adriatic Sea and Dinarides. Starting from A', the most evident feature is the presence of a high velocity lid, with V_s about 4.8 km/sec. This lid reaches the maximum depth of about 155 km in the zone that goes from the western side of the Apennines to the Tyrrhenian coast. More to the southwest, this thick high velocity lid is missing. The section BB', less than about 100 km southeast of AA', crosses the Tyrrhenian Sea, the Vesuvio and Phlegraean Fields zone, the Gargano region, the Adriatic Sea and the Dinarides. Along BB', the high velocity body with $4.6 \leq V_s \leq 4.8$ km/sec reaches depths of about 110 km under the Dinarides, about 170 km under the Adriatic Sea and about 150 km under the western side of the Apennines. More to the southwest, below Vesuvio and Phlegraean Fields, the high velocity body extends to depths not less than 250 km.

In Figure 2c, a balanced cross section from the Tyrrhenian to the Ionian Seas, along CC', is plotted down to 500 km. Our data do not resolve deeper than about 250 km, therefore, below this depth, the subducting Ionian lithosphere is outlined on the basis of the

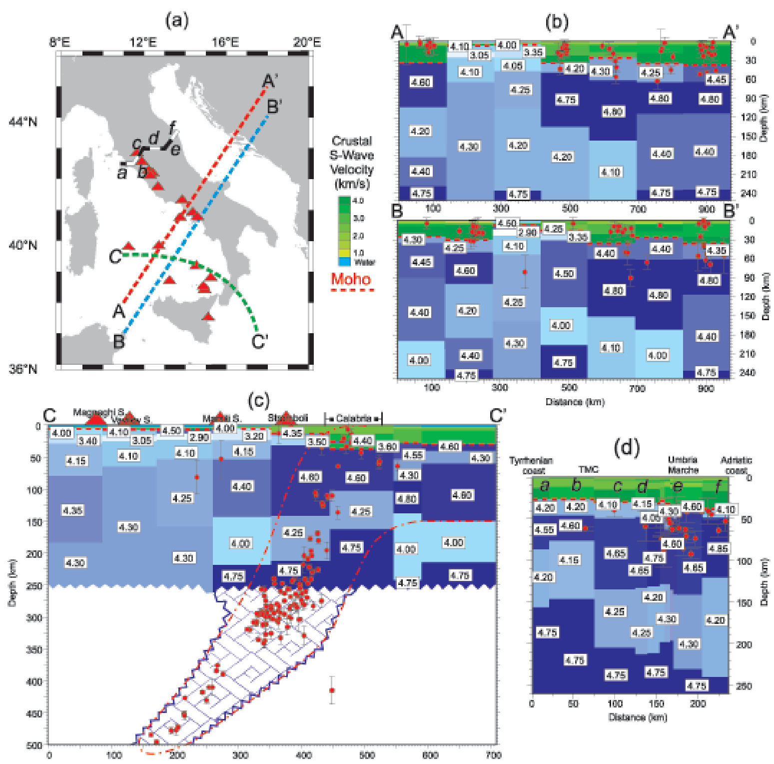

Figure 2 (a) Position of sections and of all recent volcanoes (triangles) (Amiata-Vulsini, Cimino-Vico-Sabatini, Albani, Roccamonfina, Phlegraean Fields-Vesuvio, Vulture, Ischia, Stromboli, Vulcano-Lipari, Etna, Ustica, Marsili, Magnaghi, Vavilov); (b) Tyrrhenian Sea-Southern Apennines-Dinarides: the almost continuous high velocity body seen along BB' is not visible along AA'; (c) Tyrrhenian-Ionian Sea: outlined Ionian slab; shallow, intermediate-depth and deep earthquakes that fall into a band, about 100 km wide, along sections in (b) and (c), with the depth error bars being shown; (d) lithosphere-asthenosphere system from the Tyrrhenian (a) to the Adriatic coast (f) and related intermediate-depth seismicity; the mantle wedge supports the lithospheric delamination beneath Central Italy.

hypocenters distribution of the intermediate-depth and deep seismicity. In correspondence to the shallow-mantle magma sources of the volcanic bodies Magnaghi-Vavilov and Marsili, low V_s layers (very shallow asthenosphere) below the very soft thin lid are detected. A very low velocity layer (mantle wedge) below a thin uppermost lid in the Stromboli area and a lithospheric doubling beneath Calabria are seen. In the southernmost part of CC', the crustal thickness is about 30 km and the lithospheric upper mantle is characterized by a layering where a relatively low velocity body (V_s about 4.3 km/sec) lies between two fast ones. At depths greater than about 150 km, a very well developed low velocity (V_s about 4.0 km/sec) asthenospheric layer is present. Crossing Calabria, the low velocity asthenospheric layer is absent and the relatively low velocity body (V_s about 4.25 km/sec) in the lithospheric mantle becomes deeper and thicker going towards west.

Figure 2d shows the lithosphere-asthenosphere system along a stripe from the Tyrrhenian to the Adriatic coasts (Chimera et al., 2003), particularly detailed in UMD (see zone *e* in Figure 2d). Beneath Central Italy high velocity bodies reach at least a depth of 130 km with a width of about 120 km. The crust exhibits clear V_s layering and lateral variation in thickness: less than 30 km below the Tuscan Metamorphic Complex (TMC) and about 35 km below UMD. The lid is thin (about 30 km) below the TMC, while it is about 70 km thick below UMD. Along the profile, particularly in the western part where it gets shallower, a developed mantle wedge separates the crust from the high velocity lid.

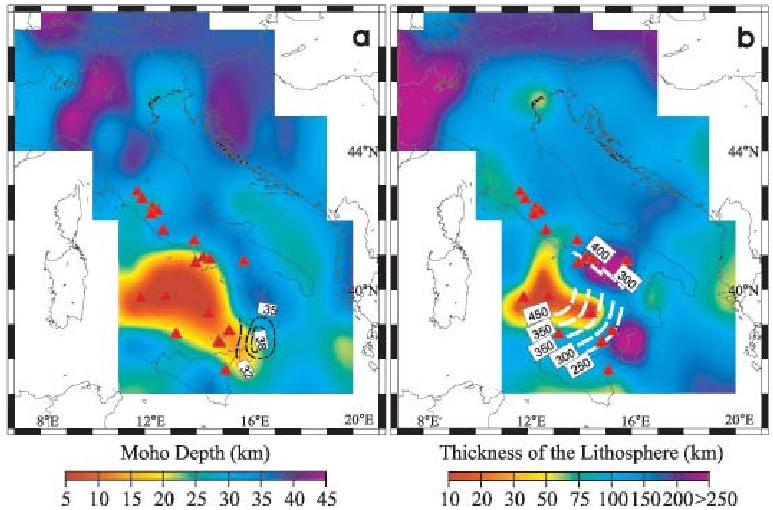

***Figure 3** (a) Moho depth with contouring of the deeper Moho where lithospheric doubling is unambiguously detected; (b) thickness of the lithosphere. Here and in Figure 4, the dashed lines schematise the subduction of the Ionian-Adria lithosphere, traced accordingly with ISC hypocenters distribution, and red triangles mark the recent volcanoes.*

Maps of the lithosphere-asthenosphere

The horizontal resolution of our maps is about 100 km and the vertical penetration reaches a depth of about 250 km. All the features shown at depth larger than 250 km are schematically based on the intermediate-depth and deep seismicity, as given by ISC, schematised by dashed segments in Figures 3b and 4b,c.

In Figures 3a,b, the Moho depth and the thickness of the lithosphere are shown, together with the recent volcanoes (red triangles). In Figure 3a, the contouring of the deeper Moho indicates where lithospheric doubling is unambiguously detected by our data. In the northernmost area of the map in Figure 3b, the lithospheric thickness is about 200 km, while in the Western Alps it is at least 250 km. The lithospheric thickness varies in the range of about 100–150 km along the Northern Apennines, around the Padan plain and in the Dinarides area, except in its westernmost part, where the lithosphere is only about 80 km thick. The Northern Adriatic Sea has a lithosphere thinner than the Central-Southern Adriatic Sea. In the southernmost Adriatic Sea and in the Otranto channel area, the lithosphere is less than about 100 km thick. In the Calabrian and Campanian areas the lithospheric thickness exceeds 250 km.

The two different dashing patterns in Figure 4a, where the V_s just below the Moho is shown, indicate the presence of the mantle

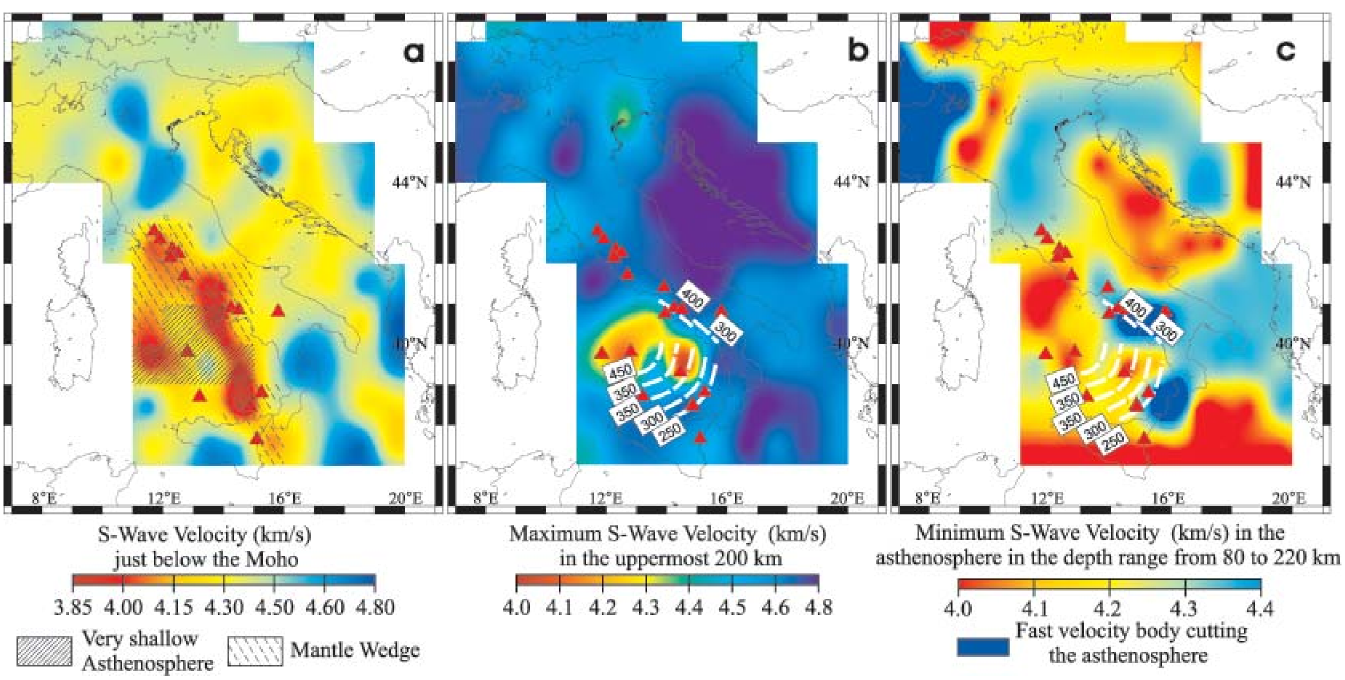

***Figure 4** (a) V_s just below the Moho (different dashed patterns outline where mantle wedge and a very shallow asthenosphere are detected); (b) maximum V_s in the uppermost 200 km; (c) minimum V_s in the asthenosphere, in the depth range from about 80 km to about 220 km. In (c) the dark blue areas indicate the fast velocity bodies cutting the asthenosphere.*

wedge and of a very shallow asthenosphere, respectively. Large lateral variations (Figure 4b) characterize the maximum V_s in the uppermost 200 km. Peak values are found in the western Alps, central Po valley, Dinarides, Central Adria, Southern-Central Apennines, Northern Tyrrhenian and Ionian Seas. In correspondence to all the volcanoes, except the Tyrrhenian Seamounts, the maximum lithospheric V_s exceeds about 4.6 km/sec. Magnaghi-Vavilov and Marsili are separated by a region with relatively high V_s, but in correspondence to these volcanoes V_s is very low. The high-velocity bodies of the Ionian-Adria subducting slabs extend below the volcanoes of the Aeolian Arc and of the Campanian province. The minimum V_s in the asthenosphere, in the depth range from about 80 km to about 220 km, is shown in Figure 4c. The dark blue area in the northern part of the map corresponds to the fast velocity bodies present in the western Alps. East of this lithospheric body, the V_s in the asthenosphere is as low as in the Northern Adriatic. The properties of the asthenosphere in the Northern Adriatic Sea (V_s between 4.0–4.1 km/sec), are different from those of the Southern Adriatic Sea and around the Otranto channel (V_s larger than about 4.3 km/sec). Low asthenospheric V_s is seen in Sicily and in the Tyrrhenian and Ionian Sea. The dark blue areas around the Tyrrhenian Sea indicate the fast velocity Ionian-Adria slabs that cut the asthenosphere, and that can be traced at depth larger than 250 km from the distribution of subcrustal seismicity.

Discussion

In BB' (Figure 2b) the high-velocity body extending to depths not less than 250 km can be related to the westward subduction of the Adriatic lithosphere towards the Tyrrhenian Basin. This feature is in agreement with the results of De Gori et al. (2001). Along a section very close to AA' (Figure 2b) they find a weak velocity perturbation in the mantle beneath the mountain belt with a small high velocity anomaly dipping southwestward. This feature can be correlated with the layer with V_s about 4.75 km/sec, whose top is at about 220 km, in AA'. The rising of the bottom of the asthenosphere could be caused by remnants of high velocity bodies probably detached (Wortel and Spakman, 2002 and references therein) from the lithospheric roots, through thermo-mechanical processes. The remarkable difference between the two sections of Figure 2b, confirmed by completely independent data, indicates that the subducted lithosphere has a very complex morphology.

In CC' (Figure 2c), the body with V_s about 4.4 km/sec near the center, above the slab, is probably due to thermal effects induced by the mechanical interaction between the Ionian lithosphere and the hot Tyrrhenian upper mantle. In the center of the section, the layer with V_s around 4.0 km/sec, extending from about 140 km to 220 km depth, can be explained by dehydration processes and melting along the down-going slab (e.g. Goes et al., 2000 and references therein). The layering along the easternmost half of CC', in the Ionian area, seems to be consistent with the subduction of serpentinized and attenuated continental lithosphere, formed in response to the Jurassic extensional phase. During the tensional phase, the relatively low velocity (V_s in the range 4.25–4.30 km/sec) layer could be formed as a result of the serpentinization of peridotites. Such process produces V_s retardations of a few percent (Christensen, 1966). The presence of a low velocity layer of chemical and not of thermal origin is consistent with the low heat flow in the Ionian Sea (Della Vedova et al., 1991). This layer, when subducted, gets thicker, consistently with the dehydration of serpentine, which is responsible for the weakening of the neighboring material. The seismicity is distributed along the slab and it seems to decrease, but it is not absent, in correspondence with the serpentinized layer. In the studied part of the Ionian Sea, the lithosphere is attenuated continental, thermally relaxed after the Jurassic extensional phase, while in the Southern Tyrrhenian Sea it is very young oceanic.

Beneath Central Italy (see Figure 2d) there is clear evidence of lithospheric roots surmounted by a well-developed mantle wedge. Young magmatism at the surface and high heat flow in the TMC region suggest that, in agreement with petrological and geochemical data (Peccerillo et al., 2001), this layer may represent a partially molten mantle. In Tuscany, the mantle wedge is underlined by a thin lithosphere and an up-risen asthenosphere roof, in agreement with the heat flow data (Della Vedova et al., 1991). Along the same vertical section, the rising of the bottom of the asthenosphere may have the same origin discussed for section AA' (Figure 2b). The subcrustal earthquakes (ISC) cluster in the shallower part of the thick Adriatic lid and in the eastern part of the lithospheric root, consistently with a slab-like geometry, while the part of the lithospheric root and thin lid to the west seems to be almost free of seismic activity. The absence of deep seismicity and the non-in-depth continuity of the fast velocity body below Central Apennines (Figure 2d) and Southern Apennines (Figure 2b) clearly highlight a major difference when compared to the structure and related deep seismicity of the Calabrian arc, where there is sound evidence of a continuous slab.

The complex crustal structure, where shearing and thrusting involve the whole crust and the upper mantle, causing the occurrence of more than one Moho, as described by Nicolich and Dal Piaz (1990), is confirmed by our data and the map in Figure 3a reproduces several other features identified by the same authors. Near the Otranto channel the Moho is in the range of 25–30 km, i.e. shallower with respect to the results of Nicolich and Dal Piaz (1990), but well in agreement with the Moho depth proposed by Herak and Herak (1995).

In Figure 3b, the Western Alps lithosphere at least 250 km thick is consistent with the presence of the lithospheric root (Panza and Mueller, 1979). The Northern Adriatic Sea has a lithosphere thinner than the Central-Southern Adriatic Sea, where a band with moderate seismicity can be identified. The lithosphere is very thin, less than 20 km, in correspondence to Magnaghi, Vavilov and Marsili. The lithospheric thickness exceeding 250 km in the Campanian and Calabrian areas is associated with the subduction of the Ionian-Adria lithosphere, schematically represented, for depths larger than 250 km, by the isolines in Figure 3b.

The mantle wedge area shown in Figure 4a is in agreement with what proposed by Meletti et al. (2000) in their structural and kinematic model of Italy. In some cases, the lowest velocity material is not just below the Moho but below a thin mantle lid, possibly formed by thermal underplating. All the volcanic areas (see the caption of Figure 2), except those with the inactive volcanoes of Vulture and Ustica, are characterized by the presence of a low velocity layer just below the Moho or below a very thin lid.

The dark blue area in the northern part of Figure 4c, due to the plate collision process between Eurasian and African plates, contains the so-far proposed locations of the rotation pole of Adria versus Europe (Meletti et al., 2000 and references therein).

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Giuliano F. Panza, *Professor of seismology in the Department of Earth Sciences - University of Trieste, and head of SAND Group ICTP-Trieste. Laurea in physics from the University of Bologna in 1967; PostDoc at UCLA. He is fellow of Academia Nazionale dei Lincei, of Academia Europea, and Third World Academy of Sciences. He is winner of the EGS Beno Gutermberg medal in 2000, and received Laurea Honoris Causa in Physics in 2002 from the University of Bucharest. He is leader of several projects funded by EC related to seismic hazard assessment.*

Antonella Pontevivo, *PhD from the Department of Earth Sciences -University of Trieste in 2003. Laurea in physics from the University of Trieste in 1999. At present PostDoc in the Geological Institute - University of Copenhagen. Her PhD thesis is on surface-wave tomography, non-linear inversion and geophysical implications in the Italian area and surroundings.*

Giordano Chimer*a is PhD-student in the Department of Earth Sciences - University of Trieste. He received his Laurea in physics from the University of Trieste in 1998. His interests are: tomography and non-linear inversion in the Apenninic and Alpine areas.*

Reneta Raykova, *Laurea in physics from the University of Sofia in 1994. As PhD student in the Department of Seismology at the Geophysical Institute of Sofia, she received in 2003 from the Department of Earth Sciences - University of Trieste, a one year EU - Marie Curie fellowship. Her interests are surface wave tomography and structure of the crust and upper mantle.*

Abdelkrim Aoudia, *research scientist at the International Centre for Theoretical Physics at Trieste. He received his PhD in geophysics from the University of Trieste in 1998. His research interests revolve around using geophysical, geodetic and tectonic data to understand the mechanical behavior of earthquake faults and to constrain better conceptual and quantitative models for lithospheric deformation.*

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

Geological outline of the Alps

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

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

Introduction

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

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

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

Structural framework

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

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

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

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

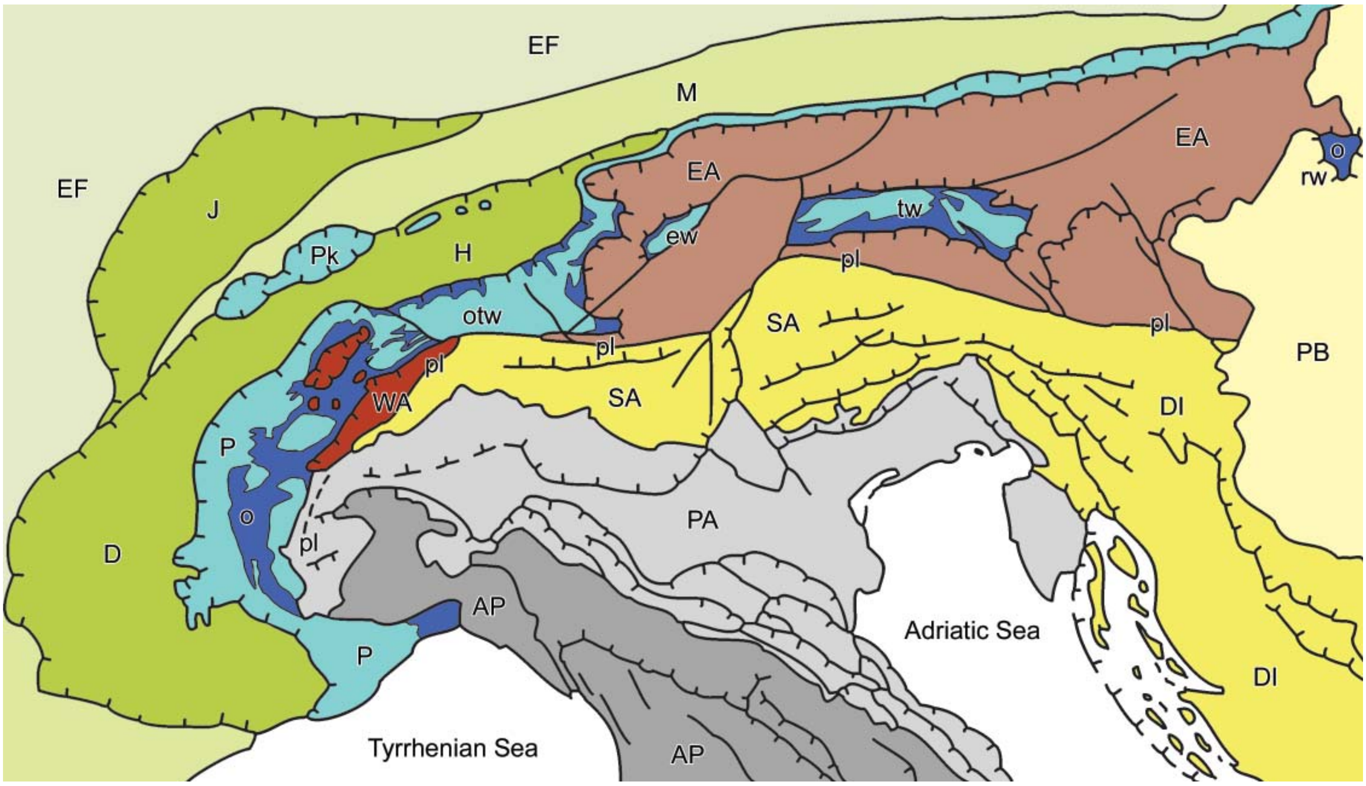

***Figure 1** Tectonic map of Alps - (1) Europe-vergent collisional belt: i) Western (WA) and Eastern (EA) Austroalpine; ii) Penninic domain: continental and ophiolitic (o) nappes in western Alpine arc (P) and tectonic windows (otw: Ossola-Ticino, ew: Engadine, tw: Tauern, rw: Rechnitz); Prealpine klippen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

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

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

The Austroalpine thrust units

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

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

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

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

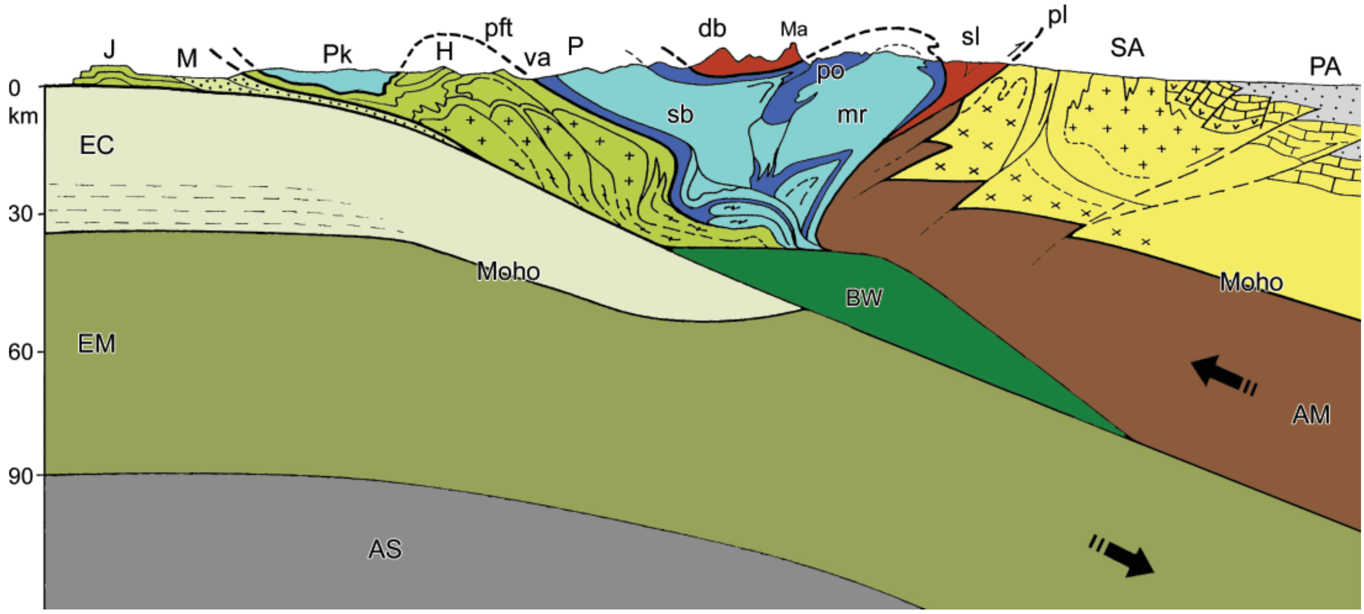

Figure 2 Lithospheric section of north-western Alps - 1) Austroalpine: Sesia-Lanzo inlier (sl) and Dent Blanche nappe s.l. (db), including Matterhorn (Ma); 2) Penninic domain (P): Piedmont ophiolitic units (po), Monte Rosa (mr) and Grand St. Bernard (sb) nappes, underlain by lower Penninic and outer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

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

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

The Penninic zone

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

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

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

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

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

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

The Helvetic-Dauphinois zone

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

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

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

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

Southern Alps

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

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

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

Geological history

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

Variscan and older evolution

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

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

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

Permian-Mesozoic evolution

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

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

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

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

Alpine orogeny

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

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

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

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

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

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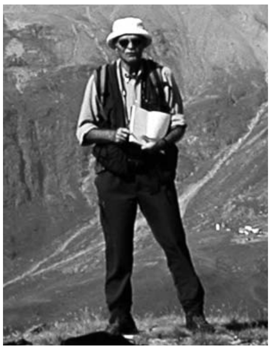
Giorgio V. Dal Piaz*, full professor of geology in the Faculty of Sciences of the University of Padova, Italy, was born in 1935 and his research activity focuses on field survey, tectonics, subduction metamorphism and hard rock geology in the Alps.*



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

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

by Alfonso Bosellini, Piero Gianolla, and Marco Stefani

Geology of the Dolomites

Dipartimento di Scienze della Terra, Corso Ercole I d'Este 32, 44100 Ferrara, Italy.

The Dolomites region is a spectacularly exposed portion of the Southern Alps, a northern Italian chain derived from the comparatively gentle deformation of the Tethyan passive continental margin of Adria. The region had an active Permo-Jurassic tectono-magmatic evolution, leading from Permian magmatism, through a Middle Triassic episode of fast subsidence and volcanism, to the Jurassic oceanic break-up. Although the sedimentary succession ranges in age from Middle Permian to Cretaceous, the geological landscape is largely dominated by the majestic Triassic carbonates, making the area a classical one for the early Mesozoic stratigraphy. Particularly noteworthy are the Anisian to Carnian carbonate platforms, recording an evolution from regional muddy banks to isolated high-relief buildups. The filling of the various basins and the development of a last generation of regional peritidal platform followed. The carbonate platforms of the Dolomites bear witness to a remarkable set of changes in the carbonate production and to significant palaeoclimatic fluctuations, from arid to moist conditions and vice versa; a great range of margin and slope depositional styles is therefore recorded. Alpine tectonic shortening strongly affected the area, with a first Eocene deformation, followed by later Neogene overthrusting and strike-slip movements.

Introduction

The purpose of this article is to introduce the reader to the general geology of the Dolomite Region. Discussion, however, will be focused on the majestic Triassic dolomite mountains, considered worldwide as typical examples of ancient carbonate platforms and buildups. Since the second half of the 19th century, the seminal studies by Richthofen (1860) and Mojsisovics (1879) recognized the reefal nature of the Dolomite Mountains, described and correctly interpreted the steep clinostratification pattern, identified several generations of platforms and provided a first biostratigraphic framework.

After dwindling research and reduced interest through the first half of the 20th century, great attention was again focused on the Triassic reefs and buildups, eventually leading to a modern synthesis of their depositional geometries and evolution (Bosellini, 1984). During the last twenty years, further substantial progress in the understanding of the geological evolution of the "reefs" has been achieved: the biostratigraphic and chronological framework was significantly improved (Brack and Rieber, 1993; Mietto and Manfrin, 1995), the main carbonate producing biota were recognized (Senobari-Daryan et al., 1993; Russo et al., 1998, 2000), and the origin of the platform-top sedimentary cyclicity has been the theme of a hot debate (Goldhammer et al., 1990; Brack et al., 1996; Egenhof et al., 1999; Preto et al., 2001), without, for the time being, reaching any firm conclusion. Finally, an improved knowledge of the basinal successions has made the integration with their platform counterparts possible, leading to a far better understanding of the sequence stratigraphic architecture (Gianolla et al., 1998a and references therein).

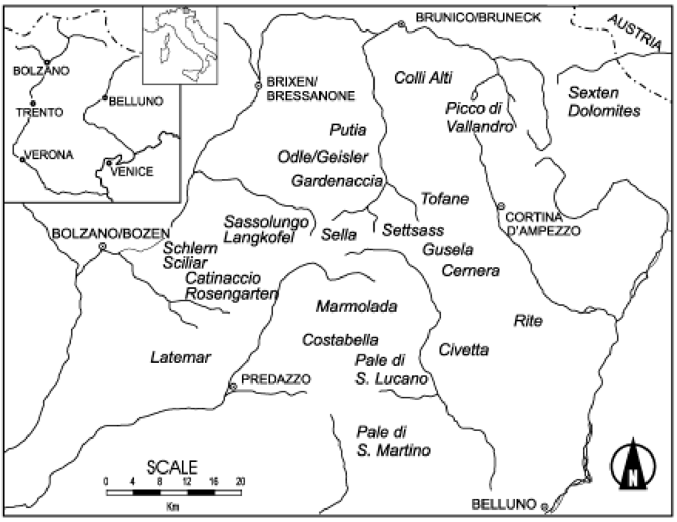


Figure 1 Location map of the Dolomites (northern Italy), with indication of the most important platforms and buildups.

Structural setting

The Dolomite Mountains are a group of carbonate edifices relatively well confined from the physiographic point of view (Figure 1). They are located in the eastern part of the so-called Southern Alps, a south-vergent fold-thrust belt (Doglioni, 1987; Castellarin, 1996), which constitutes a major structural unit of the Alpine Chain. The Dolomites themselves can be seen as a large pop-up related synclinorium of Neogene age (Doglioni, 1987), limited to the north by the dextral Insubric Lineament and to the south by the Neogen south-vergent Valsugana Overthrust (Figure 2). They constitute a relatively coherent slab of upper crust carried southward for at least 8–10 km. The sedimentary cover, preserved within this 60-km-wide synclinorium, is comparatively mildly deformed by tectonics; intense penetrative deformation as well as very large horizontal displacements do not occur. The region, however, records several magmatic and tectonic events including:

1. Volcanics and rifting of Permian and Early Triassic age, which produced N-S trending structural "highs" and "lows";
2. Late Ladinian magmatism and tectonics;
3. Rifting and continental margin evolution, associated with the opening of the western Tethys (Ligurian Ocean), which controlled differential thickness and facies during Late Triassic, Jurassic and Early Cretaceous times;
4. Paleogene N60ºE (Dinaric) compression producing a WSW-vergent thin-skinned thrust belt in the central-eastern Dolomites (Doglioni and Bosellini, 1987);
5. Finally, during the Neogene, the Dolomites became the innermost part of a south-vergent thrust belt (Figure 2); most of the present elevation has been generated during the last 10 million years.

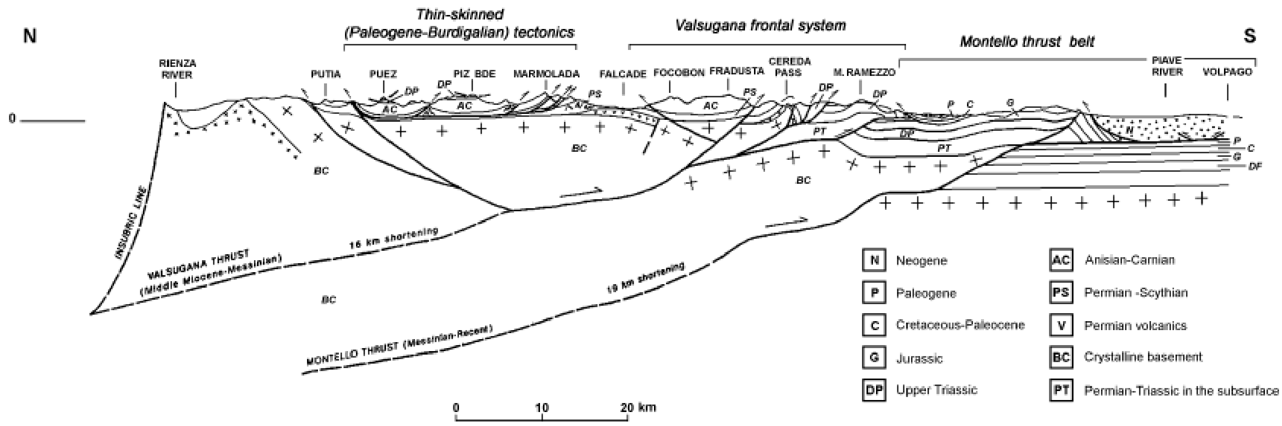


Figure 2 Geological profile across the western Dolomites (from Castellarin et al., 1998).

Regional stratigraphy

The stratigraphic framework of the Dolomite region includes Permian to Cretaceous terrains (Figure 3). Following the Carboniferous Variscan orogeny, deformed and metamorphosed Paleozoic rocks were uplifted and eroded, thus forming the regional basement.

The Early Permian rifting resulted in the accumulation of a thick volcanic package, the so-called Bozener Porphyry Plateau. This volcanic complex covers over 2000 sq. km, with thicknesses locally exceeding 2000 m. The sedimentary succession of the Dolomites unconformably overlies the Lower Permian volcanics or, where they are missing, rests directly on the crystalline basement. It begins with Upper Permian red beds (Gardena Sandstone) deposited in a semi-arid setting of alluvial fans, braided streams and meandering rivers (Massari and Neri, 1997). Following marine transgression from the Paleothethys to the east, transitional and shallow marine evaporites and carbonates (Bellerophon Formation) succeed upward in the Late Permian. Facies associations of the Bellerophon Formation suggest a wide spectrum of depositional environments, ranging from coastal sabkha to shallow shelf (Bosellini and Hardie, 1973; Massari and Neri, 1997).

The Lower Triassic Werfen Formation unconformably overlies the Permian sequence and consists of a complex succession of shallow-water carbonate and terrigenous deposits. The Werfen Formation is 300–400 m thick and is subdivided into several depositional sequences, recognizable over a wide portion of the Southern Alps.

During the early Middle Triassic, local uplifting, subaerial erosion and strong subsidence took place. Differential movements along fault blocks set the stage for localized carbonate production: the Middle Triassic carbonate platforms and buildups nucleated upon slightly elevated areas.

The Anisian platforms

A first widespread tidal flat unit (Lower Serla Dolomite), laterally grading into evaporitic environments, gave way to a complex framework of three partially superimposed carbonate platform systems: Monte Rite Formation, Upper Serla Dolomite and Contrìn Formation. Because of a general sea-level rise, the lower two platform systems drowned and long-lasting basinal environments, recorded by terrigenous-carbonate successions (Dont, Bivera and Ambata formations), succeeded in the eastern Dolomite. During the late Anisian, while subsidence was still active throughout the eastern Dolomites, the western area was significantly uplifted and subaerially eroded, locally exposing Permian sediments (Bosellini, 1968). Renewed transgression brought back marine environments to the western areas, where shallow-water carbonate platforms (Contrìn Formation) devel-

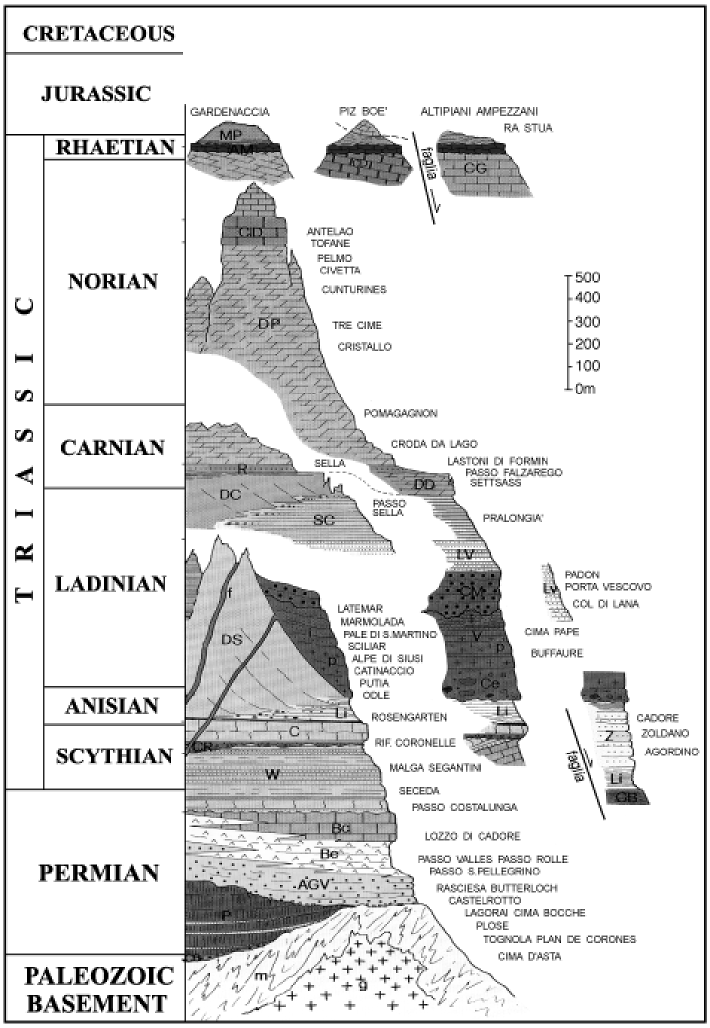


Figure 3 Composite stratigraphic succession of the central-western Dolomites. Granite (g); metamorphics (m); Basal Conglomerate (BC); Porphyry (P); Gardena Sandstone (GS); Bellerophon Fm: Black limestone (Bl), Evaporites (Be); Werfen Fm. (W); Braies Group (GS); Richthofen Conglomerate (CR); Contrin Fm. (C); Livinallongo Fm. (B); Zoppè Sandstone (Z); Sciliar Dolomite (SD); Volcanics (V): pillow lava (p), hyaloclastites (h), chaotic heterogeneous (Ch), dykes (d); La Valle Fm. (FM, LV), Marmolada Conglomerate (MC); San Cassiano Fm. (SC); Cassian Dolomite (CD); Dürrenstein Fm. (DD); Raibl Fm. (R); Dolomia Principale (DP); Dachstein Limestone (CD); Calcari Grigi (CG); Rosso Ammonitico (RA); Puez Marl (MP).

oped. These platforms, rich in dasycladacean algae, associated with encrusting and *problematica* organisms (*Tubiphytes*), widely prograded over lagoonal-basinal terrigenous-carbonate deposits (Morbiàc Formation).

The pre-volcanic carbonate buildups (late Anisian–late Ladinian)

A regional drowning terminated the previous Anisian platform and basin system, but shallow-water carbonate-producing environments "survived" at small isolated highs. Soon, however, these banks grew quickly upward by aggradation, forced by a phase of regional subsidence, which created a large accommodation space. These aggrading banks or buildups initially shared many facies similarities with the former and wider Contrìn platforms, being still rich in dasycladacean and *Tubiphytes* muddy sediments. The upward growth of some buildups was terminated by an early drowning, especially in the eastern, more subsiding portion of the region (Cernera, Casera Plotta, Tiarfin) and were covered by condensed ammonoid-bearing limestones (Gianolla et al., 1998a). The western buildups (referred to as Sciliar Dolomite or Marmolada Limestone, according to their composition), were on the contrary able to survive and catch up with the fast-growing relative sea level (e.g. Latemar, Catinaccio, Marmolada, Pale di San Martino). These platforms rapidly reached a thickness of 800–900 m, while just a few metres of cherty limestones were accumulating in the adjacent basins (lower portion of the Livinallongo Formation). In the eastern Dolomites, both the basinal and the platform successions are thicker then their western counterparts.

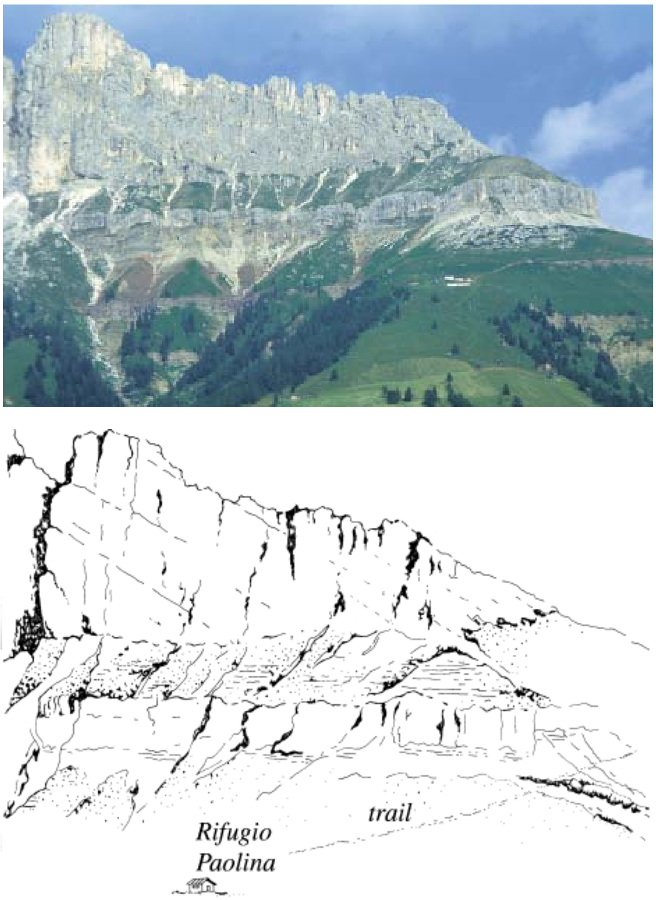


Figure 4 The southern end of the Catinaccio/Rosengarten platform, where the steep clinoforms show the horizontal progradation of the carbonate system.

The aggradation rate of these late Anisian-early Ladinian buildups was in the order of 200–400 m/Ma, but significant lateral variations did exist, being largely controlled by the regional differential subsidence. However, problems in the geochronometric evaluation of the different time intervals make the estimation of the aggradation and progradation rate somewhat uncertain. At the Fassian-Longobardian boundary, the subsidence slacked considerably and a massive progradation phase began, spanning over a comparatively short time interval of the late Ladinian. The progradational phase was characterized by pervasive phreatic marine cementation of the margin and upper slope sediments and by the development of very steep (40–45°), planar breccia slopes (Figure 4). Since the progradation rate largely exceeded the basinal accumulation rates, the surface of contact between the base of slope and the basinal unit can be sharp and sub-horizontal in geometry, simulating a pseudo-downlap relationship. Through this fast and remarkable progradation, the isolated buildups expanded considerably and became platforms with a width in the order of 5–10 km. In the western Dolomites, the average progradation rate of the base of slope was probably between 1400 and 2700 m/Ma. In the northeastern Dolomites, where subsidence was still quite active, the rate of the base-of-slope migration was considerably reduced.

During the same time, acidic volcanogenic layers (the so-called "pietra verde") were deposited in the entire Southern Alps, whereas the eastern Dolomites were the site of an important accumulation of turbiditic sands (Arenarie di Zoppè), deriving from the erosion of a Variscan metamorphic basement. The terrigenous and volcanic deposits document an active tectonic scenario that was soon to generate an important magmatic phase within the Dolomites themselves.

The volcanic event

The platforms were involved in the Ladinian tectono-magmatic event; they were cut by a great number of shoshonitic basaltic dykes and carved by large collapses, while huge heterogeneous megabreccia bodies (Caotico Eterogeneo Auct.) accumulated in the basins. The volcanic products (pillow lavas, hyaloclastites) partially infilled the basinal depressions, onlapping the platform slopes and "freezing" their original morphology (Figure 5). A few platforms of the western Dolomites (Agnello, Latemar, Viezzena, Marmolada) close to the volcanoes were even buried beneath the volcanic products. Some kind of carbonate production was nonetheless still active all the time, even close to the major magmatic centres (e.g. Sciliar). In areas far away from the volcanoes (eastern Dolomites) the carbonate sedimentation was not interrupted; here the lack of any true depositional break within the continuous platform-top successions makes the distinction between pre- and post-volcanic succession locally difficult.

The post-volcanic platforms (late Ladinian–early Carnian)

At the fading of the magmatic activity, an even healthier carbonate production developed, supporting the widespread progradation of several generations of carbonate platforms (Cassian Dolomite). The post-volcanic platforms record the progressive colonization of the margin environment by Techosmilia-like branching corals, which were however always subordinated to smaller sediment-producing organisms. Ooid grains reappeared during the earliest phases of the volcanic activity (Acquatona Formation), after being absent since the Early Triassic.

In the western Dolomites, the available accommodation space was not produced by subsidence but mainly inherited from the pre-volcanic "collapse" of the area. The platform, therefore, could only expand laterally, prograding over the adjacent deep-water basins. In the eastern Dolomites the subsidence was still ongoing and considerable. Clinostratifications are concave in shape and generally less

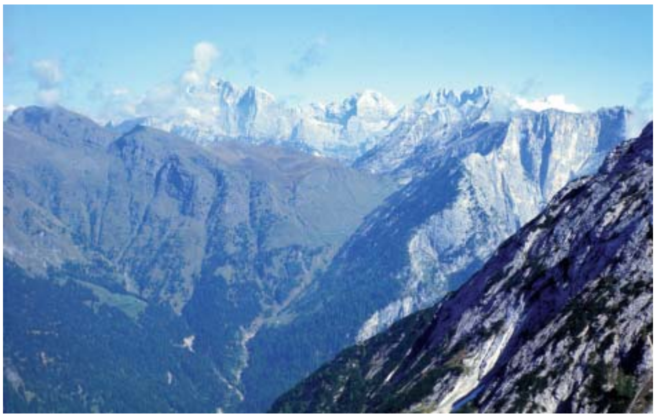


Figure 5 The western slope of the Pale di San Martino platform onlapped and "fossilized" by volcaniclastic products.

Figure 6 Oblique-tangential prograding pattern (Conturines Group-La Varella).

steep than the pre-volcanic ones (Figure 6). The high basinal sedimentation rates, owing to the large availability of volcaniclastic sediments, produced a shallowing evolution of the basinal areas. Moreover, the combined effect of the platform progradation and of the basinal aggradation resulted in climbing base-of-slope progradation, visible in areas facing major sediment sources (e.g. western Sella).

Early post-volcanic aggrading platform-top successions are relatively thick in the subsiding eastern Dolomites (e.g. Picco di Vallandro-Duerrenstein), whereas in the western Dolomites the same successions are very thin and associated with some terrigenous influx ("Schlern Plateau Beds"). In the central-eastern Dolomites, two platform generations (Cassian Dolomite I and II *Auct.*) are separated by a temporary interruption of the progradational evolution (Figure 7), matched with renewed transgression and with the onlap of the basinal beds onto the former carbonate slopes (e.g. Richthofen Riff and Settsass, etc.).

Figure 7 The "Richthofen Reef" (Piccolo Settsass). Thin-bedded shale and limestone of the San Cassiano Fm. onlap the slope of the buildup (Cassian Dolomite), which is wedging out towards the right with tongues of resedimented breccia and crinoidal grainstone.

The Carnian crisis of the rimmed carbonate platforms

During the early Carnian (younger Julian), the amount of loose carbonate mud available in the prograding slopes increased, while the platform slope elevation was progressively reduced by the shallowing-up of the basin and by some terrigenous clay content; these factors combined together to progressively reduce the slope angles, as visible in the latest Cassian Platforms (e.g. Lastoi di Formìn and Picco di Vallandro/Dürrenstein). The very late evolution of these platforms was matched with the appearance of patch reefs, for the first time rich in "modern" colonial corals, while true buildup systems disappeared. This evolution corresponds to a worldwide crisis of the rimmed carbonate platforms.

The basin eventually shallowed up into the photic zone, probably also because of a relative sea-level drop, starting an *in situ* active carbonate production, even in the deeper depocentre areas. This evolution triggered the deposition of the low-gradient Dürrenstein Formation, which records a complex palaeoenvironmental evolution. In the western Dolomites, this Carnian interval is however poorly recorded, mainly because of the lack of available accommodation space. These complex depositional systems witness important climatic fluctuations, marked by the development of moist phases. This interval is also relevant for its bearing the oldest known Mesozoic amber (Gianolla et al., 1998b).

The Upper Triassic carbonate platform: a regional peritidal succession

During the late Carnian (Tuvalian) the previous platform/basin systems were replaced by a variety of shallow-water environments (terrigenous, evaporite and carbonate sediments of the Raibl Formation). The preceding uneven morphology was levelled and a large carbonate platform was established over large portions of the Alpine region. In the central-western Dolomites, this carbonate system, the so-called Dolomia Principale, normally started with subtidal facies, grading upward into rapidly aggrading cyclic peritidal successions (Bosellini and Hardie, 1988). The eastern margin of this widespread peritidal platform lies in the Tarvisiano area, some 100 km at the east of the Dolomites. Here a well-preserved carbonate slope is documented by steep prograding clinostratifications, rich in serpulids, dasycladacean algae, microbial mats and pervasive phreatic cementation (De Zanche et al., 2000).

During the Norian Time, dis-anoxic intraplatform depressions, rich in carbonate mud and organic matter, developed in different areas of the vast Dolomia Principale platform, to the east (Friuli and Carnia), south (Bellunese) and west (Lombardia) of the Dolomites. In fact, differential subsidence, associated with widespread extensional processes, controlled the evolution of the Upper Triassic platforms, heralding the rifting stage of the Jurassic passive continental margin of Adria (African Promontory).

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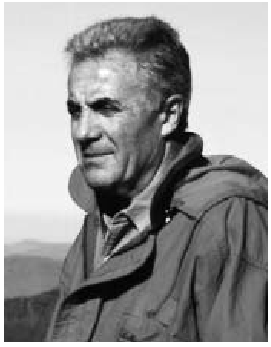
***Alfonso Bosellini** earned his degrees at the University of Padua. He held postdoctoral fellowships in the United States, at the Johns Hopkins University and Scripps Institution of Oceanography. His wide interests in structure, stratigraphy and carbonate sedimentology led him to study the Dolomites, Apulia, Somalia and Ethiopia. He is currently a professor of geology at the University of Ferrara, honorary fellow of the Geological Society of America and past president of the International Association of Sedimentologists.*



***Piero Gianolla** earned his PhD degree at the Padua University and currently holds a research position at the University of Ferrara. His main research fields consist in the sequence stratigraphy, paleoclimatological and sedimentological investigation of carbonate platforms and mixed basins. He is involved in a stratigraphic revision of the Triassic aimed at establishing a supra-regional sequence stratigraphic framework, through the correlation of sections from the Southern Alps, the Northern Calcareous Alps, the southern Appennines and Hungary.*

***Marco Stefani** received a PhD. degree in 1989, then worked in the Hydrocarbon Geology field and at the IFP, Rueil-Malmaison. He joined the Earth Sciences Department of the Ferrara University, Italy, in 1995, where he has been working ever since. His main research fields are the depositional dynamics of Triassic carbonate platform and basin systems and the Quaternary stratigraphy and G.I.S.-assisted geological mapping of the Po Delta Area.*

by Attilio Boriani[1], Francesco Sassi[2,3], and Raffaele Sassi[2]

The basement complexes in Italy, with special regards to those exposed in the Alps: a review

1 Dipartimento di Scienze della Terra, Università di Milano and Istituto per la Dinamica dei Processi Ambientali, Consiglio Nazionale delle Ricerche, Via Botticelli 23, 20133 Milano, Italy.
2 Dipartimento di Mineralogia e Petrologia, Università di Padova, Corso Garibaldi 37, 35137 Padova, Italy.
3 Istituto di Geoscienze e Georisorse, Consiglio Nazionale delle Ricerche, Sezione di Padova, Italy.

Most of the sedimentary rocks occurring in Italy are post-Carboniferous. All what lies below is considered basement, mostly metamorphic or igneous. Understanding the pre-Carboniferous evolution depends on the reconstruction of the sedimentary, metamorphic and igneous evolution of the basement. In general, the basement sedimentary protoliths were Lower to Middle Paleozoic siliciclastic rocks, while the igneous protoliths belong to an Ordovician cycle. The prevailing metamorphism, from very-low grade to granulite facies, is Variscan. It was followed by the formation of large amounts of granitic melts.

Introduction

The term "basement complexes" (**BCs**) is used here for the set of rock complexes underlying the post-Variscan unconformity (the age of the bottom of the related volcano-sedimentary cover being various, ranging in most cases from Upper Carboniferous to Lower Trias). These basement complexes show structural vestiges of older deformational events, and the effects of older depositional, erosional, metamorphic and magmatic events. The related old features are more or less modified by younger deformational and thermal overprints. The basements described here are mostly metamorphic and crosscut by Upper Paleozoic "younger granites"; however some unmetamorphosed old Paleozoic sequences are also present (but not described here).

Basement complexes in Italy are exposed in many places, as shown in Figure 1a, and with more details in Figures 1b to 1d.

To the North, along the Alps, BCs occur as the deepest lithostratigraphic element in each of the main structural complexes (or *nappe systems*) making up the Alpine chain: the Helvetides, the Pennides, the Austrides (or Austroalpine) and the Southern Alps (or Southalpine) (Figure 1b). In the peninsular and insular part of Italy, along which the Apennine mountain chain develops, BCs are exposed in ***Tuscany*** (surroundings of Florence in Figure 1a) and, more to the south, in ***Calabria and Sicily***. The latter BCs are part of the so-called ***Calabria-Peloritani Terrane*** (Figure 1c). Finally, BCs dominate as a component of the ***Sardinia*** island (Figure 1d), where they are classified in Axial Zone, Nappe Zone and External Zone.

The ***Calabria-Peloritani Terrane*** is a tectonic juxtaposition of a N subterrane (the Sila and the Serre massifs) and a S subterrane (Aspromonte massif and Peloritani Mts).

As regards ***Tuscany***, the data from the small and scattered occurrences are integrated with subsurface data obtained from drillholes in the Larderello geothermal field.

The main features of the Italian BCs are summarily outlined here, in the narrow limits of the necessarily small number of available pages. The literature is extremely large and, in order to save space the quotations throughout the text and the list of references are reduced to the minimum possible. The main source of data utilized here are the review volumes listed below. When single papers are quoted in the text marked by an asterisk, they are not included in the list of References, but their full citation may be found in one or more of the listed References. Anyway, the full list of the cited papers may be requested to the authors of the present paper.

The main source of data used here are: Flügel et al. (1987), Sassi and Zanferrari (1989), Carmignani and Sassi (1992), Peccerillo and Sassi (1994), von Raumer and Neubauer (1993), Frey et al. (1999), Carmignani et al. (2001), Vai and Martini (2001), Sassi et al. (2003) and, as regards radiometric geochronology, Anonimous (1985) for all age data available in Italy until 1985, and Thöni (1999) for more recent data on the Alps.

It is worthy to point out that the rock complexes occurring in the Alps cross the State boundaries with Austria, France and Switzerland. Furthermore, all other Italian BCs also attracted the attention of numerous not-Italian scientists. Therefore the description given below is also based on their numerous and significant contributions.

Both the discussion and the list of papers quoted in the text are more detailed for the BCs of the Alps than for the other BCs occurring in Italy. This fact reflects not only the literature, which is much more abundant for the Alps, but the personal experience of the authors of the present review. More specific data can be found in Pandeli et al. (1994) and Carmignani et al. (2001b*) concerning Tuscany, in Carmignani et al. (2001) and Sartori (2001*) as regards ***Sardinia***, in Bonardi et al. (2001*), Rottura et al. (1990*, 1993]), Grässner and Schenk (2001), Grässner et al. (2000*), Caggianelli and Prosser (2002) as regards the ***Calabria-Peloritani Terrane***.

General structural setting

The Italian mountain chains (Alps, Apennines), in which the BCs represent the oldest rocks (mostly Lower Paleozoic), formed during the Alpine Orogeny (mainly Tertiary). The largest BCs exposures occur in the Alps, i.e. in the complex, nappe structured chain which formed due to the closure of the Tethys ocean and the collision of the European plate with the African plate (specifically the Apulian or Adria microplate). As mentioned above, BCs occur, with different features, in each of the four main structural domains of the Alps. Before plate collision, the Helvetic and Pennidic BCs were part of the European plate, whereas the Austridic and Southalpine BCs were both part of the African plate.

Paleomagnetic and structural data, as well as lithological affinities, indicate that Sardinian BCs were a continuum with the S-European BCs and shared with them (particularly with Catalonia and Provence) the same evolution until Oligocene, before getting detached and drifting southwestwards to the present position. Struc-

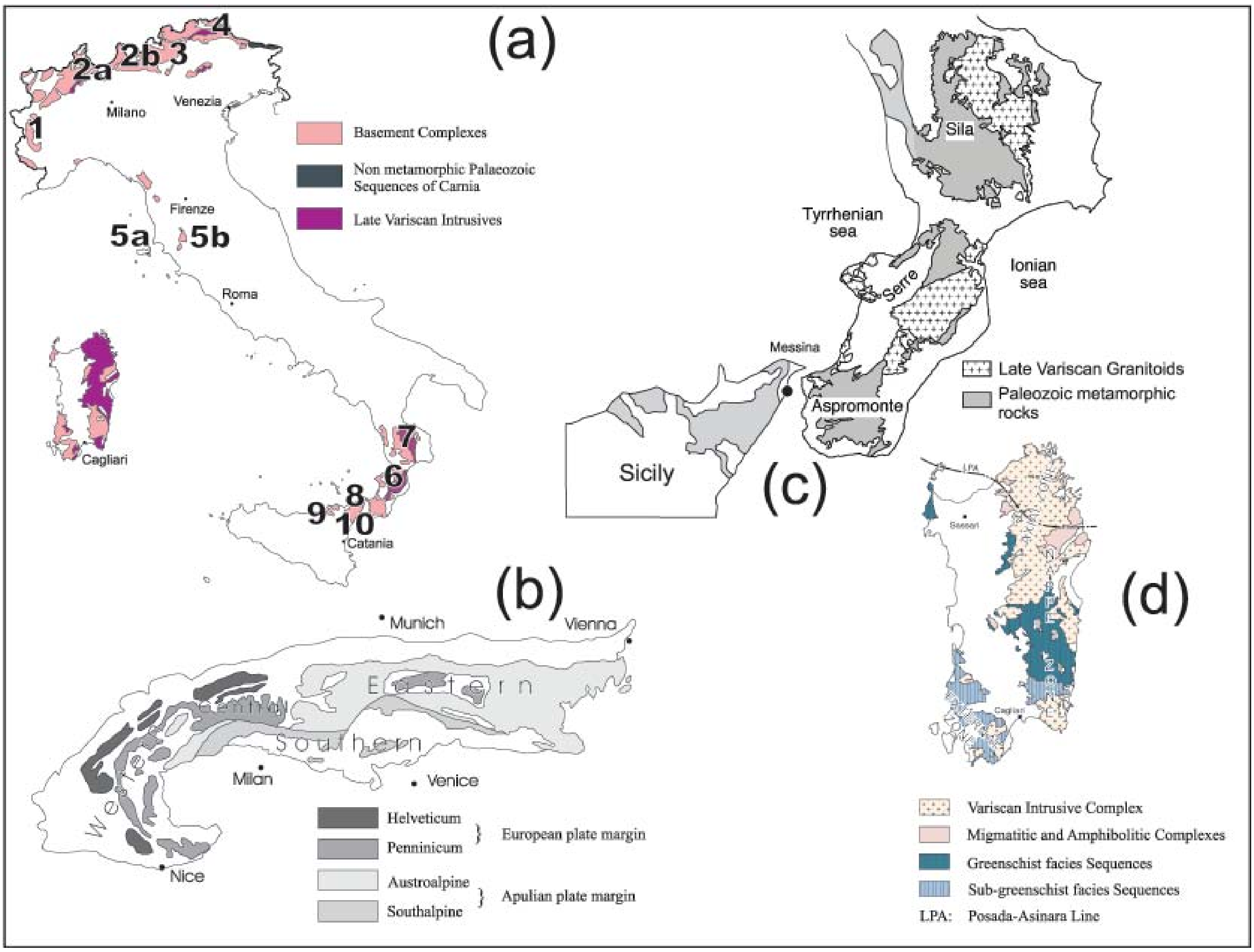


Figure 1 (a) Location of the basement complexes exposed in Italy. (b) Sketch map of the main structural units making up the Alps: the different grey areas are basement complexes in the different structural units (taken from Spiess et al., 2001, modif.). (c) Sketch map of the basement complexes of the so-called Calabria-Peloritani Terrane, i.e. in the southernmost part of the peninsular Italy and the northwest corner of Sicily (taken from Prosser et al., 2003, modif.). (d) Geological sketch map of Sardinia, showing that basement complexes outcrop over a dominant area of the island (taken from Carmignani et al., 2001, modif.).

tural data and lithological affinities suggest that the Ligurian Briançon units (Western Alps) also belonged to the Sardinia-Provence structure.

The ***Calabria-Peloritani Terrane*** is a fault-bounded exotic terrane, the origin of which is still debated, and probably is the result of amalgamation of two or three microterranes. An African origin has been suggested for the N sector, and a possible European origin for the S one.

BCs of ***Sardinia, Tuscany*** and ***Calabria-Peloritani Terrane*** were involved in the "alpidic" (Mesozoic and Tertiary) geodynamic precesses which took place in the Central-Western Mediterranean, but ***Sardinia*** escaped Alpine metamorphism.

As pointed out below, all BCs occurring in Italy underwent the Variscan thermal and tectonic events (with local exceptions: see below). Similarly to all other South-European basement terranes, they represent polydeformed, sometimes polymetamorphic domains which, since the Lower Paleozoic, underwent a complex and still debated history of drifting and amalgamation processes. They acquired the present setting due to Pre-Variscan, Variscan and Alpine events.

Lithology, stratigraphy, geochemistry

Most rock types indicate a shallow crustal nature, but rocks referred to an intermediate to deep crust occur both in the Alps (the so called Ivrea-Verbano Zone) and in ***Calabria***. Variously aged and more or less retrogressed eclogite lenses and boudins locally occur.

The shallow crustal sequences consist of metapelitic-psammitic rocks of various metamorphic grade with lenses or intercalations of leucocratic gneisses, amphibolites, acidic and basic metavolcanics and some marbles.

The deep crustal Ivrea-Verbano Zone consists of high-T amphibolite and granulite facies metamorphic rocks derived from pelites, mafic rocks and limestones with lenses of ultramafic rocks. The presence of lenses of spinel lherzolite near a large layered mafic pluton (Rivalenti et al., 1984) induced in the past many authors to consider the Ivrea-Verbano Zone a crust-mantle transition (Mehnert, 1975*; Zingg, 1983*). See also Schmid (1993*) and Quick et al. (1994*, 1995) for an overview on the Ivrea Zone. The Variscan deep crustal sections exposed in Northern Calabria (Sila and Serre) consist of granulites, migmatites, metapelites, metaultramafics, metabasites, and some marbles. In Sardinia and Ligurian Alps, eclogites affected by granulite overprint and associated to migmatites possibly represent deep crustal sections.

The sedimentation age of the protoliths of the metasedimentary sequences making up the bulk of the BCs has been determined by microfossil in very-low grade rocks. Sedimentation age ranges from Cambrian to Devonian. Sassi and Zanferrari (1989) present a general review of all biostratigraphic data available at that time for all BCs occurring in Italy. More recent biostratigraphic data concerning the very-low to low grade Austridic BCs of the Alps can be found in several papers in von Raumer and Neubauer (1993). In the metamorphic BC of the Southern Alps, only two fossil findings have been reported: Cambrian acritarchs in the most southeastern outcrop of the Southalpine BC (Sassi et al., 1984*; Kalvacheva et al., 1986*) and Ordovician-Silurian palynomorphs in the central Southalpine BC (Gansser and Pantic, 1988*).

In the medium to high grade metasedimentary sequences, a Lower Paleozoic sedimentation age of the protoliths is assumed. Precambrian age values have been only obtained radiometrically from detrital zircons in metasediments and in some zircon xenocrystals inherited in Paleozoic igneous rocks. However, some rocks and deformational events in the Silvretta Nappe are reported as Precambrian (Maggetti and Flisch, 1993*).

A special mention deserve the metabasites (mostly amphibolites, but also eclogites and metaultramafics), which occur in every BC, and which have been extensively used for the reconstruction of the evolutionary paths and geodynamic interpretations: e.g., in the Alps: von Raumer and Neubauer (1993, and refs. quoted therein), Miller and Thöni (1995*); in ***Sardinia***: Carmignani et al. (2001 and refs. quoted therein); in the ***Calabria-Peloritani Terrane:*** Bonardi et al. (2001*, and refs. quoted therein). Eclogites are commonly reported as pre-Alpine and sometimes pre-Variscan, but in the Austrides of the Eastern Alps evidence has been found that Eoalpine eclogites also occur (Miller and Thöni,1997). In the Ligurian Alps, a glaucophane-bearing eclogitic overprint occurring in the T. Visone Unit is considered of possible Eo-Alpine age.

Data and considerations on the pre-Cambrian development of the crust of the European Variscides in and outside the Alps can be found, e.g., in Gebauer (1993*).

As regards geochemical aspects, a large amount of data exists, mainly concerning igneous and orthometamorphics. In addition Sassi et al. (2003) present and discuss the abundance of 55 chemical elements and petrovolumetric models of the crust in 10 type areas from the BCs of Italy, with some geophysical and petrophysical data (see their location in Figure 1a).

Metamorphism

The Alpine Overprint

In many cases, the present metamorphic features of BCs are Paleozoic, in other cases Alpine metamorphism(s) affected various BCs even strongly. From this point of view, three extreme situations occur:

- BCs in which the present metamorphic features are Alpine, and the characters of the pre-Alpine metamorphism(s) cannot be defined: this is the case, e.g., of the main part of the Pennidic BCs in the Central and Eastern Alps;
- BCs in which an Alpine metamorphic overprint occurs, but the metamorphic pre-Alpine features can still be defined: this is the case, e.g., of parts of the BCs in the Austrides, or parts of the BCs of Calabria-Peloritani Terrane and Tuscany;
- BCs which were localized outside the Alpine metamorphic domain: e.g., the BCs of Sardinia and the Southern Alps.

Alpine metamorphic evolution developed in two multistage events. The complexity of such a scenario was increased by late-Alpine tectonic, post-metamorphic dismembering, which disconnected and variously displaced BC terrains. The features of the Alpine overprints are disregarded here.

The Variscan metamorphism

It affected all BCs occurring in Italy. The only exception known are: (i) to the west, the occurrence of some units in the Briançon basement (Desmons, 1992*); (ii) to the east, the occurrence, 5 km underneath Venice, of Ordovician (460–470 Ma on euhedral zircons) granodiorites (Meli and R. Sassi, 2003). We are inclined to consider as local but significant details these two situations, which do not weaken the concept that Variscan metamorphism regionally dominated over all terrains from which the south-European BCs derived.

As regards the features displayed by the Variscan metamorphism, it has been reported as a low pressure event in some areas (e.g. in the Austridic BCs of the Eastern Alps consistently with the common occurrence of andalusite and cordierite in rocks of appropriate bulk composition and metamorphic grade), and as a Barrovian-type event in other areas (e.g. in the Western-Central Alps, in which however sometimes a low-pressure, late Variscan metamorphic overprint occurs: Colombo and Tunesi, 1999). In the last decade, evidence was found of the existence of an early (about 350 Ma), HP stage of the Variscan metamorphism (Miller and Thöni, 1995*; Hauzenberger et al., 1996*; Godard et al., 1996*).

In the Central-Western Southern Alps the main metamorphic phase was followed by a retrograde metamorphism which occurred during uplift and erosion. This overprint, generally not pervasive, was particularly intense along late-orogenic shear zones (Colombo and Tunesi, 1999).

The age of the metamorphic peak in the Ivrea-Verbano Zone is early Permian (Boriani and Villa, 1997). It has been related to the thermal effect of the intrusion at the base of the crust of the mafic Ivrea body in a late-orogenic extensional geodynamic regime (mafic underplating).

In the Axial Zone of ***Sardinia***, the Posada-Asinara Line, a milonitic belt including relics of eclogites, is interpreted as a possibile Variscan oceanic suture, i.e. a part of the suture between Armorica and Gondwana. The amphibolite facies paragneisses and micaschists along it include amphibolite lenses preserving relics of a HP metamorphism, specifically relics of granulitic and eclogitic mineral assemblages. It separates two metamorphic domains: a Migmatitic Complex to the N and a Barrovian type, mainly Amphibolite Facies Complex to the S. In the Nappe Zone, the synkinematic Variscan metamorphism is of Barrovian-type, mainly in the greenschist facies, locally in the subgreenschist facies, or in the amphibolite facies. A postkinematic, low pressure overprint is well documented both in the Axial and in the Nappe zones. In the External Zone metamorphism, where detectable, is of very-low or low grade (Carmignani et al., 2001).

The ***Calabria-Peloritani Terrane*** includes various types of Variscan metamorphics, from weakly metamorphosed sequences to upper amphibolite facies rocks, granulites and migmatites. Relics of eclogites also occur. In the greenschist facies pelitic sequences, the mineral associations do not give definite indication of the P character. In the medium to high grade rocks, an early Variscan metamorphism under medium pressure conditions followed by a widespread low P overprint has been suggested (Schenk, 1990*).

The BCs in Tuscany mostly consists of Variscan low-grade sequences. A late Variscan, low-P/high T overprint locally occurs. Records of an early, barrovian-type stage of medium to high grade are reported (Pandeli et al., 1994).

The pre-Variscan metamorphism(s)

In some sites, as for example in the Austrides of the Eastern Alps, relics or traces of an older, Paleozoic (ca. 470–500 Ma), “Caledonian” metamorphic event have been reported (e.g. Sassi and Schmidt, 1982 * ; Ebner et al., 1987*; Frisch et al., 1987*; Sassi et al., 1987*; Hoinkes and Thöni, 1993*), although the matter is still debated (e.g. Schulz et al., 1993*). Some pieces of controversy deal with different interpretations of observed data, other on the name "Caledonian" (the use of "Cadomian" or "Panafrican" could make discussion smoother). Matte (1991) considers this event as Eo-Variscan, and refers it to the closure of the Rheic oceanic basin. The main Variscan metamorphism is instead interpreted as a fully intra-continental event.

In any case, the matter of facts is that, during upper Ordovician, an enormous amount of acidic melts (in some units more abundant than the Variscan ones) was formed through crustal melting and emplaced both as plutonic bodies and as volcanic products over a very large area. The degree and the extent of the regional metamorphism that may have accompanied or preceded this magmatic event are not well established. A few traces that may suggest a pre-Variscan metamorphic event are: (i) the occurrence in Oetztal of fragments of banded, partly amphibolitized eclogites within some marbles affected by low-pressure Variscan metamorphism; (ii) the occurrence in the Upper Ordovician orthogneisses of partly amphibolitized eclogites (W. Southern Alps) and of banded xenoliths, the folded foliation of which does not fit that of the host orthogneisses (E. Austrides). Furthermore, according to Hoinkes and Thöni (1993*), strong radiometric indications for a Caledonian anatexis accompanied by the formation of fibrolite support the view of a dominant Caledonian metamorphism in the Oetztal-Stubai BCs (E. Austrides).

It must be noted, however, that in the W. Southern Alps (Strona Ceneri Zone) Boriani et al. (1990b and 1995) report evidence that, at the moment of the intrusion of the Ordovician granites, the protoliths were unmetamorphosed.

In ***Sardinia***, it is not possible to exclude the occurrence, in the Migmatitic Complex occurring in the NE part of the island, of pre-Cambrian sequences preserving a pre-Variscan structuration. Anyway, a "Caledonian" event (but not including metamorphism) is documented at least in the SE ***Sardinia*** by an angular unconformity and a Middle Ordovician calc-alkaline magmatism (Carmignani et al., 2001).

In the **Calabria-Peloritani range,** some authors consider the eclogite facies relics and some granulites as representing the original, Cadomian or intra-Cambrian basement of the low-grade terraines. Data and interpretations about the metamorphic evolution recorded in the continental crust of the Calabria-Peloritani Terrane may be found in Paglionico and Piccarreta (1978*), Schenk (1990*), Caggianelli et al., (1991*), Grässner and Schenk (2001) and Fornelli et al. (2002) .

Magmatism

The Mesozoic and Tertiary igneous events are obviously disregarded here, because their products lie over the BCs (when emplaced by volcanic mechanisms) or crosscut them (when emplaced by intrusion). For the same reason, the post-Variscan, Upper Paleozoic, mainly acidic volcanics (e.g. the so-called "Bolzano Rhyolitic Plateau") are not considered here; they do not belong to any BCs, but lie over them, belonging to their non-metamorphic cover. Anyway, data on them may be found, e.g., in Cortesogno et al. (1998*), and Rottura et al. (1998*).

Late Variscan magmatism

Plutonism: Almost all BCs exposed in Italy are crosscut by granitoid intrusions, which were emplaced in the late stages of the Variscan before the above mentioned Permo-Triassic unconformity. They produced contact metamorphic aureoles on the surrounding, already metamorphic basement rocks. The literature on this subject is extremely abundant and cannot be easily summarized in this article. For the Alps Bonin et al. (1993*) wrote an extended and interesting review paper to which the reader is invited to refer. Granitoids are widespread in all tectonic units in most of which they experienced a pervasive Alpine overprint. In the Southalpine they were not affected by metamorphism and are therefore less problematic as far as age determinations and geochemical characters are concerned. The age of intrusion spans over a large time range. Their geochemical and isotopic signature point to a complex evolution of magma source and geodynamic regime. The oldest granitoids are peraluminous and reflect crustal thickening and more or less wet melting conditions. They are accompanied by high-K mafic magma. The Lower to Middle Carboniferous magmas are high-K calc-alkaline and reflect post-collisional rapid uplift and erosion. They are followed by Late-Carboniferous almost alkaline associations emplaced in a "Basin and Range" tectonic regime. Late-Carboniferous to Early Permian calc-alkaline (sometimes high-K) granitoids are reported to reflect a convergent setting. As far as the Southalpine is concerned, this interpretation conflicts with that given by Pinarelli et al. (2002*), who explain those characters with an origin of the parent mafic magma from an enriched mantle. Mid-Permian to Triassic A-type plutonic/volcanic complexes conclude this long story of granitoid magma formation. They are related to the post-orogenic continental consolidation of the European plate.

In ***Sardinia,*** the so-called "Sardinian Batolith" is exposed over a large part of the island and extends to a large part of Corsica, making up one of the most important batholiths of the European Variscan Chain. It has a composite structure, and its emplacement took place over a 40 MA time span, related to an extensional regime. Several rock types are represented in it (mainly granodiorites and monzodiorites related to a calc-alkaline association (Carmignani et al., 2001, and quoted liter.).

In the ***Calabria-Peloritani Terrane***, Late Carboniferous-Permian granitoid intrusions occur mostly in the N sector (Figure 1c). This magmatism shows a calc-alkaline character, and includes meta- and per-aluminous plutons emplaced at different levels from the lower to the upper crust (Rottura et al., 1990*, 1993*; Ayuso et al., 1994*).

Upper Ordovician magmatism

During Upper Ordovician a very strong, mainly acidic magmatic activity took place, both under plutonic and volcanic conditions. In the Eastern Alps, the possible co-genetic relation between volcanics and plutonics was first proposed by Peccerillo et al. (1979*) and Bellieni and Sassi (1981*), who also suggested the idea of an "Upper Ordovician Granite-Rhyolite Association", and further supported by Mazzoli and R. Sassi (1992*). Such a suggestive hypothesis, although consistent with the presently available data, still requires further analytical support.

In Sardinia, Ordovician magmatic rocks are also reported in some structural units. They include both plutonics (granodiorites) and acidic plus basic volcanics. The protoliths of the acidic orthogneisses are also referred to the Ordovician magmatic cycle (Carmignani et al., 2001, and quoted liter.).

Some felsic metavolcanics occurring in Northern Calabria are considered of Ordovician age. A widespread Middle Ordovician, acidic volcanism (metaignimbrites) is also documented in ***Tuscany,*** as well as late Ordovician intrusive metabasites.

Plutonism

Due to the Variscan, and sometimes also Alpine overprints, the products of this plutonism are presently orthogneisses, the chemistry of which ranges mainly within the fields of granites and granodiorites, but also tonalites and quartz diorites. In the Eastern Alps a huge amount of melts is related to this event. They display a calc-alkaline affinity. SiO_2 wt% covers almost continuously the 62–77% range The more acidic rock types show a higher HREE fractionation. Negative Eu anomaly is a common character. The lack of genetically related basic rocks is to be pointed out. All data available indicate an origin by a regional crustal anatexis. However, the Ordovician metagranitoids outcropping in the Western Southern Alps show evidence

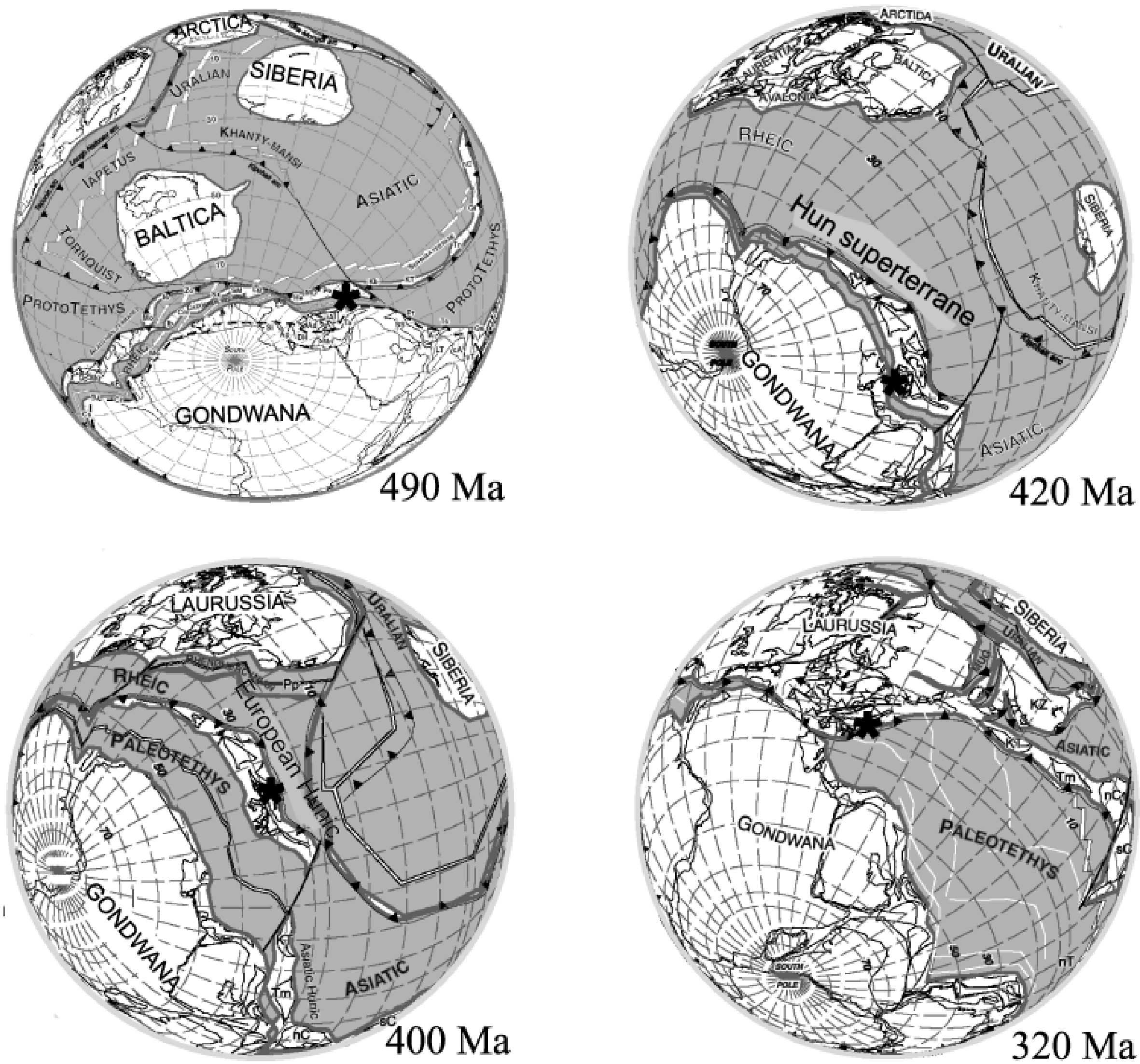


Figure 2 Plate drifting from 490 to 320 Ma ago according to Stamplfli et al. (2002), and related progressive migration of the Hun Superterrane from the margin of Gondwana to the margin of Laurussia. Asterisk shows roughly the location, in different times, of the microplates from which the basements occurring in Italy derived (a collage of the Figures 2 and 3 from Stamplfli et al., 2002).

of crust-mantle interaction, and are considered as records of a magmatic activity developed in a convergent plate boundary environnement (Pezzotta and Pinarelli, 1994*; Boriani et al., 1995*). Radiometric data (mostly Rb/Sr on large-sized rock samples) define a range of 470–420 Ma.

In ***Sardinia,*** worth of mentioning are the granodioritic orthogneisses of Capo Spartivento, in the southernmost part of the island (external Zone), and the similar orthogneisses in the north part of the Nappe Zone.

Volcanism

One of the most striking features in these basements is the widespread occurrence of mainly acidic volcanics and volcanoclastics which expanded, probably as ignimbrites, over large areas during Upper Ordovician (and somewhere Middle Ordovician). These rocks, then, underwent the Variscan metamorphism, which changed them into the so-called “porphyroids” in the low-grade areas, and in stratiform white gneisses in higher metamorphic grade areas. In every BCs these metavolcanics represent a well detectable key-horizon, also useful in inter-regional correlations.

In the Eastern Alps, they occur both in the South-Alpine and in the Autridic BCs. In the South-Alpine BCs, where they have been dated radiometrically on zircons (ca. 480 Ma, Meli and Kloetzli, 2001*) geochemistry and petrology based on parameters insensitive to later re-mobilization, indicate a crustal anatectic origin through dehydration melting reactions, in a possible late to post-orogenic scenario (Meli, 1998*).

The low grade Austridic BCs also record, in different areas, other Paleozoic volcanic processes (the products of which are obviously metamorphic). According to Loeschke and Heinisch (1993*), they are within-plate alkali basalts of Ordovician, Silurian, Devonian and also probably lower Carboniferous age, indicating a long-lasting, Paleozoic, extensional process. The protoliths of the amphibolites occurring in the medium to high grade metamorphics may be partly related to this process.

In Sardinia, in some areas of the Nappe Zone, four Ordovician volcano-sedimentary metamorphic complexes have been defined, including acidic metavolcanics, and intermediate and basic meta-epiclastites and metavolcanics.

Pre-Ordovician events

In the Western Southalpine basement, like in other parts of the Variscan belt, a peculiar bimodal lithological association is found, the Leptyno-Amphibolitic Group (LAG), which consist of alternating cm-thick, fine-grained leucocratic and melanocratic layers of amphibolite facies metamorphic rocks, with lenses of metagabbro, retrogressed eclogites and ultramafites. Their geochemistry is compatible with a derivation from back-arc tholeiites. These LAG may well represent the reworked products of an early Variscan suture, in which ophiolitic remnants that underwent HP metamorphism in a subduction zone have been incorporated in turbiditic sediments derived from back-arc bimodal volcanics before the intrusion of the Ordovician granites that cut across this unit (Giobbi Mancini et al., 2003). A similar situation seems to occur in the Ligurian Briançon units (Western Alps).

As regards the older history, records of major pre-Cambrian crust-forming events have been detected by radiometric dating of zircons: see Gebauer (1993*) for a review and discussion.

Cambrian sedimentary sequences (partly very-low grade metamorphics) are widespread in SW Sardinia. Furthermore, the protoliths of the Migmatitic Complex occurring to the North of the Posada-Asinara-Asinara Line are condidered as pre-Cambrian, and therefore should record pre-Cambrian events deeply oblitareted by the Variscan, high grade, polystage metamorphism and anatexis (Carmignani et al., 2001).

The possible existence of pre-Ordovician events in the Calabria-Peloritani Terrane is documented by zircon age values in the range 550–622 Ma obtained from metabasites and acidic orthogneisses in Serre and Aspromonte (Schenk, 1990*; Senesi, 1999*). In Tuscany, boreholes in the geothermal field reached gneisses and micaschists underlying Cambro-Ordovician rocks.

The BCs in Italy in the context of the Paleozoic organization of the European basements

As pointed out above, all BCs occurring in Italy acquired the present setting due to pre-Variscan, Variscan and Alpine events. These events are to be considered as critical records of a continuous plate drifting and various amalgamation processes. Such an evolution was common to all South-European BCs at least since Paleozoic.

General models of this Paleozoic evolution of the European BCs are available in the literature (e.g. Stampfli et al., 2001*; Vai, 2001*; von Raumer et al., 2002; Stampfli et al., 2002), The matter is still relatively fluid and rapidly evolving, although some pieces of the mosaic seem to be definitely established. However, significant pieces of knowledge are still lacking or poorly known, including specific field data. Furthermore, the interpretation and better dating of some specific igneous and metamorphic events are needed, in order to strengthen, better constrain or modify the various existing models. This is particularly necessary for the BCs occurring in Italy, the correlation among which and with other European BCs is at a rather early stage.

In a very simplified way Figure 2 shows, as an example, the model proposed by Stampfly et al. (2002). The asterisk roughly indicates the location, in different times, of the microplates from which the basements occurring in Italy derived. 490 Ma ago the South European basements were localized at the Gondwana margin, separated by the Early Rheic ocean. 420 Ma ago all continental fragments making up the present Sud-European BCs belonged to the Hun Superterrane, which progressively drifted towards Laurussia and finally, 320 Ma ago, accreted to it. All these basements display not only similarities because of their common Paleozoic history, but also peculiarities which depend on their individual evolution. The main goal of the present research in this field is to focus on both the similarities and the peculiarities, in order to better defining such type of models, and to contribute to their interdisciplinary discussion.

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***Attilio Boriani** (on the right) is full professor of Petrology at the Department of Earth Sciences of the University of Milano. His main research interest is the metamorphic and igneous evolution of the continental crust, with special attention to the Alps. He was member of the Geological Committee of the National Research Council (1976–1981). He has been Vice-President from 1992 to 1996 and Secretary General of the IUGS from 1996 to 2002. He is a member of the Italian National Academy (of Lincei). He is currently President of the 32nd International Geological Congress.*

***Francesco Sassi** (on the left) is full Professor of Petrography at the Department of Mineralogy and Petrology of the University of Padova. He led several national research projects concerning the basements in Italy from 1982 to 1999, and two IGCP projects on the Circum-Mediterranean basements (Leader of No. 5 and co-leader of No. 276). He also studied the petrologic mineralogy of white micas. He is a member of the Italian National Academy (of Lincei). In 1995 he received the National Prize for Physical, Mathematical and Natural Sciences.*

***Raffaele Sassi** (in the middle) from 1993 is lecturer of "Lithology and Geology" at the Faculty of Engineering, University of Padova. He studied the evolution of crystalline basements in the Eastern Alps, Sardinia, Calabrian-Peloritan Mountains, Slovakia and Hungary. His research activity includes also the compositional variations of metamorphic phyllosilicates, as a function of intensive and extensive variables in low-grade metamorphism. Since 2002 he has been Secretary General of the IUGS Commission on Systematics in Petrology (CSP).*

by Giovanni B. Piccardo

Mantle processes during ocean formation: Petrologic records in peridotites from the Alpine-Apennine ophiolites

Dept. for the Study of the Territory and Resources, University of Genova, Corso Europa 26, 16132 Genova, Italy.

Mantle peridotites were early exposed at the sea-floor of the Jurassic Tethys derived from the subcontinental mantle of the Europe-Adria system. During continental rifting and oceanic spreading, these lithospheric peridotites were percolated via diffuse reactive porous flow by melt fractions produced by near-fractional melting of the upwelling asthenosphere. Ascending melts interacted with the lower lithosphere, dissolving pyroxenes and precipitating olivine, and crystallized at shallower levels in the mantle column causing melt impregnation. Subsequent focused porous flow formed replacive dunite channels, cutting the impregnated peridotites, which were conduits for upward migration of MORB-type liquids. Melt migration produced depletion/refertilization and significant heating of the percolated/impregnated mantle, i.e the thermochemical erosion of the lithosphere. Impregnated and thermally modified lithospheric mantle was cooled by conductive heat loss during progressive lithosphere thinning and was intruded by MORB magmas, which formed Mg-rich and Fe-rich gabbroic dykes and bodies. Alpine-Apennine ophiolitic peridotites record the deep-seated migration of melts, which changed their compositions and dynamics during the rift evolution. The thermochemical erosion of the lithospheric mantle by the ascending asthenospheric melts, which induces significant compositional and rheological changes in the lower lithosphere, is a major process in the evolution of the continent-ocean transition towards a slow spreading oceanic system.

Introduction

Ophiolites exposed along the Western Alps-Northern Apennine orogenic chain represent the oceanic lithosphere of the Ligure-Piemontese (or Ligurian Tethys) basin which separated, during Late Jurassic- Cretaceous times, the Europe and Adria plates.

Since the early seventies (Bezzi and Piccardo, 1971; Decandia and Elter, 1972; Piccardo, 1976), it has been recognized that Alpine-Apennine ophiolites have anomalous pseudo-stratigraphy and lithological association, with respect to an idealised oceanic lithosphere formed at mature mid-ocean ridges (see reviews in: Rampone & Piccardo, 2000; Piccardo et al., 2002). In fact: i) mantle peridotites are both fertile cpx-rich lherzolites, and depleted cpx-poor peridotites; ii) MORB-type gabbroic rocks are intruded into mantle peridotites; iii) mantle rocks record a decompressional subsolidus evolution, from lithospheric mantle depths to the sea-floor; iv) the serpentinized mantle peridotites are directly covered by MORB lava flows and radiolarian cherts, i.e. the first oceanic sediments.

The radiolarian cherts, which are frequently interlayered with the MORB lavas, show a Middle to Upper Jurassic age (De Wever and Caby, 1981; Marcucci and Passerini, 1991; Bill et al., 2001). Accordingly, a general agreement exists on the idea that the Ligurian Tethys was floored by a peridotite-gabbro basement, subsequently covered by discontinuous lava flows and oceanic sediments (Decandia and Elter, 1969; Piccardo, 1983; Lemoine et al., 1987), and on the assumption that the inception of the oceanic stage was not older than Upper Jurassic.

Tectonic setting and opening mechanisms

The Ligurian Tethys is believed to have developed by progressive divergence of the Europe and Adria blocks, in connection with the pre-Jurassic rifting and Late Jurassic opening of the Northern Atlantic (Dewey et al., 1973; Lemoine et al., 1987).

Paleotectonic reconstructions of the Ligurian Tethys suggest that the oceanic basin was not wider than 400–500 km (Stampfli, 1993) and that plate convergence led to complete closure of the Ligurian Tethys in the Early Tertiary, by means of an east-dipping subduction zone.

The peculiar stratigraphy of the Alpine-Apennine ophiolites led researchers to propose various genetic models: 1) the transform fault model (Gianelli and Principi 1977; Lemoine, 1980; Weissert and Bernoulli 1985), 2) the slow-spreading ridge model (Barrett and Spooner, 1977; Lagabrielle and Cannat, 1990; Lagabrielle and Lemoine, 1997), and 3) the low-angle detachment fault model (Lemoine et al. 1987; Froitzheim and Eberli 1990; Piccardo et al. 1990, 1994; Froitzheim and Manatschal 1996). The subcontinental origin of the mantle peridotites from the Ligurian ophiolites was stressed by some Authors (Decandia and Elter, 1969, 1972; Piccardo, 1976), which outlined the diversity of the Alpine-Apennine ophiolites compared with mature oceanic lithosphere formed at mid-ocean ridges of modern oceans. Based on the atypical association of MORB magmatism and fertile subcontinental mantle, it was suggested (Piccardo, 1977; Beccaluva and Piccardo, 1978) that the Ligurian ophiolites were formed during early stages of opening of the oceanic basin, following rifting, thinning, and break-up of the continental crust, and were therefore located in a marginal, peri-continental position of the Jurassic oceanic basin.

The passive lithosphere extension

The passive extension of the lithosphere has been recognized (Elter, 1972; Piccardo, 1976; Lemoine et al., 1987; Piccardo et al., 1990, 1994) as the most suitable geodynamic process to account for the thinning and break-up of the Europe-Adria lithosphere and the opening of the Jurassic Ligurian Tethys. The passive extension caused (Piccardo et al., 1994): i) the tectonic exhumation and sea-floor exposure of the sub-continental lithospheric mantle, and ii) the almost adiabatic upwelling of the asthenosphere, which underwent decompressional partial melting.

Modern analogues

The presence, in the Alpine-Apennine ophiolites, of a sub-continental peridotite basement and relics of stretched continental crust, which have been injected by MORB-type basaltic dykes, strongly recall the present setting in the embryonic ocean of the Northern Red Sea, i.e. the association of sub-continental peridotites and continental gneisses, cut by MORB basaltic dykes, which is exposed on the Zabargad Island (Bonatti et al., 1983; Piccardo et al., 1988, 1994). The origin of the Northern Red Sea has been related to the passive and asymmetric extension of the Nubian-Arabian lithosphere (Bohannon et al., 1989; Voggenreiter et al., 1988).

Moreover, the association of rifted sub-continental mantle and discontinuous MORB magmatism characterises the ocean-continent transition in the magma-poor rifted margin of Galicia (Western Iberia) (see: Manatschal and Bernoulli, 1999, and quoted references).

These settings represent, therefore, suitable modern analogues for the early evolution of the Jurassic Ligurian Tethys (Piccardo, 1995; Rampone and Piccardo, 2000). As early envisaged (Piccardo, 1977; Beccaluva and Piccardo, 1978), the formation of the Alpine-Apennine ophiolites must be inserted in a geodynamic scenario of passive rifting to incipient oceanisation in a slow spreading system (Vissers et al., 1991; Piccardo et al., 1994; Rampone and Piccardo, 2000). A similar interpretation has been proposed for the segments of the Jurassic Tethys margin exposed in the Malenco and Platta-Err nappes of the Central Alps (Manatschal and Bernoulli, 1999; Müntener and Hermann, 2001; Schaltegger et al., 2002).

Main features of the Tethyan ophiolites

Basaltic volcanites

Petrologic and geochemical studies have provided clear evidence of the overall tholeiitic composition and MORB affinity of the basaltic volcanites, ranging from T-MORB to N-MORB (Piccardo et al., 2002, and quoted references). Geochemical modelling indicates that the most primitive T-MORB and N-MORB-type basalts are consistent with melts generated by variable degrees of fractional melting of a MORB-type asthenospheric mantle source (Vannucci et al., 1993). These basalts have fairly homogeneous Nd isotopic ratios, consistent with their MORB affinity (Rampone et al., 1998).

Gabbroic intrusives

The dominant intrusive rock types are (Serri, 1980; Hebert et al., 1989; Piccardo, 1995; Tribuzio et al., 2000): i) ultramafic cumulates (pl-cpx-bearing cumulus dunites); ii) Mg-Al-gabbros (troctolites, ol-gabbros and cpx-gabbros); iii) Fe-Ti-gabbros (Fe-Ti-oxide-bearing gabbros and diorites); iv) plagiogranites (diorites and trondhjemites). They show the crystallization sequence [olivine(ol) Æplagioclase(plg)Æclinopyroxene(cpx)] and covariations of Fo in ol, An in plg and Mg-number in cpx, which are typical of low pressure crystallization of olivine tholeiites. Clinopyroxenes of primitive ol-cumulates and ol-gabbros have rather flat HREE to MREE patterns, at about 9–10×C1, and significant LREE depletion (Ce_N/Sm_N = 0.21–0.29). Simple geochemical modelling indicates a MORB affinity for the primary liquids, in agreement with the Sr and Nd isotope ratios of ol-gabbros and their clinopyroxenes (Rampone et al., 1998).

Mantle peridotites

Most peridotites from the Alpine-Apennine ophiolites (Lanzo-Liguria-Corsica) are spinel-facies lherzolites: they are characterized by relict protogranular textures, km-scale tectonite-mylonite shear zones, presence and enrichment of plagioclase, spinel dunite bodies and channels, and late Mg-rich to Fe-Ti-rich gabbroic dykes and bodies.

Spinel peridotites vary in composition from rather fertile (i.e. External Liguride peridotites) to variably depleted [i.e. Erro-Tobbio (Voltri Massif), Internal Liguride, Monte Maggiore (Corsica), and Lanzo peridotites]. They show a complete equilibrium metamorphic recrystallisation, attained under spinel-facies conditions at T = 900–1100 ºC (Piccardo, 1976; Rampone et al., 1993, 1995). This event documents the accretion of their asthenospheric protoliths to the thermal lithosphere, and their annealing recrystallization along a conductive geotherm (Piccardo et al., 1994). Available Sm-Nd DM model ages suggest that the Ligurian peridotites were isolated from the convecting asthenosphere, and were accreted to the thermal lithosphere, during Proterozoic times (the External Liguride fertile lherzolites: Rampone et al., 1995), pre-Carboniferous times (the Erro-Tobbio depleted peridotites: Piccardo et al., 2002), Permian times (the Internal Liguride depleted peridotites: Rampone et al., 1996). The Southern Body of the Lanzo Massif has been interpreted as an asthenosphere diapir that rose from the garnet stability field and was emplaced in early Mesozoic, during the opening stages of the Ligure-Piemontese basin, whereas the Northern Body has been considered a fragment of the sub-continental lithosphere which became isolated by the convecting mantle 400–700 Ma ago (Bodinier et al., 1991, and the quoted references). Sm/Nd isotope data on the Monte Maggiore peridotites furnish Jurassic (165 Ma) DM model age of depletion (Rampone, 2002).

Most of these ophiolitic peridotite massifs show large areas where plagioclase is present and, frequently, rather abundant. Plagioclase formation in mantle peridotites have been differently interpreted, as deriving from: i) metamorphic recrystallisation, ii) low pressure partial melting, or iii) melt impregnation. The presence of plagioclase in the Alpine-Apennine ophiolitic peridotites has been related to: 1) incomplete melt extraction and crystallisation after low pressure partial melting (Lanzo: Boudier and Nicolas, 1972; Boudier, 1978; Nicolas, 1986); 2) subsolidus recrystallisation (External Ligurides: Piccardo, 1976; Rampone et al., 1993, 1995); 3) exotic melt percolation and impregnation (Monte Maggiore and Internal Ligurides: Rampone et al., 1997). Recent studies on Alpine-Apennine peridotites (Piccardo et al., 2002; Müntener and Piccardo, 2003) suggest that:

1) The presence of reduced amounts of plagioclase, confined to plg+ol reaction rims between spinel and pyroxenes and to plg+ol-rich granoblastic aggregates between spinel-facies minerals, documents the metamorphic transition from spinel- to plagioclase-facies conditions, according to the reaction: spinel+pyroxenesÆ olivine+plagioclase;
2) The presence of significant amounts of plagioclase, diffuse in the rock as both unstrained crystals and plg+opx(+cpx)-rich granular aggregates interstitial between the deformed mantle minerals, documents the interstitial crystallisation of exotic melts, which impregnated previous spinel peridotites.

Mantle processes during ocean opening

Ophiolitic mantle peridotites from the Jurassic Ligure-Piemontese basin show field, petrologic and geochemical features which reveal the processes they underwent after their accretion to the lithosphere and before their sea-floor exposure. The following sequence of mantle processes can been recognised and related to the rifting stage of the Ligure-Piemontese basin (see also: Piccardo et al., 2002; Müntener and Piccardo, 2003).

1) Decompressional evolution of the lithospheric mantle

The passive extension of the Europe-Adria lithosphere and the inception of rifting in the Ligure-Piemontese basin were recorded in the lithospheric mantle by development of extensional shear zones, leading to gradual upwelling of segments of the lithospheric mantle, and by incipient to extensive recrystallisation under decompression to plagioclase- and amphibole-facies conditions. Geothermometric estimates indicated that the temperature conditions were slightly to significantly decreasing during the decompressional evolution (Piccardo, 1976; Hoorgeduijn Strating et al., 1993; Rampone et al., 1993,1995).

2) Melt percolation and impregnation

Extension and thinning of the lithosphere was accompanied by almost adiabatic upwelling of the underlying asthenosphere, which underwent partial melting under decompression.

Melt percolation by porous flow and melt interstitial crystallisation

Melts formed in the asthenosphere migrated through the overlying lithospheric mantle column via porous flow (Piccardo et al., 2002; Müntener and Piccardo, 2003). Melt/peridotite interaction (i.e. olivine precipitation and pyroxenes dissolution) most probably transformed the percolated lherzolites to cpx-poor peridotites, forming a lower zone of "reactive" harzburgites (Xu et al., 2003) at the expense of the percolated lithospheric mantle. At shallower levels in the lithospheric mantle column, the percolating melts began to crystallise, when cooling down to their liquidus temperatures. Melt crystallisation produced mm-scale veins and interstitial granular aggregates of undeformed magmatic minerals between the deformed porphyroclastic mantle minerals and caused pervasive impregnation of the lithospheric mantle. At the impregnation level, melts either: 1) reacted with and partially replaced the mantle clinopyroxenes, forming opx+plg symplectites, and crystallised opx-rich, cpx-free noritic microgranular aggregates (Internal Liguride and Monte Maggiore peridotites: Rampone et al., 1997; Piccardo et al., 2002), or 2) did not react with the mantle clinopyroxenes and crystallised opx-rich, cpx-bearing gabbro-noritic microgranular aggregates (Lanzo peridotites: Piccardo et al., 2002; Müntener and Piccardo, 2003).

Regarding the chemistry and origin of the trapped melts in the Monte Maggiore impregnated peridotites, Rampone et al. (1997) suggested that they probably consisted of unmixed depleted melt increment produced by 6–7% fractional melting on an asthenospheric mantle source.

Early crystallisation and abundance of orthopyroxene in the interstitial magmatic aggregates, and orthopyroxene replacement on mantle olivine, indicates that the impregnating melts were silica- and opx-saturated (as in the Internal Liguride and Monte Maggiore peridotites), whereas the interstitial crystallisation of clinopyroxene evidences that melts attained cpx-saturation (as in the Lanzo peridotites). The pyroxenes-saturation of the impregnating melts supports the idea that they migrated upward by reactive porous flow (Piccardo et al., 2002) and reacted with the country peridotite, dissolving mantle pyroxenes and crystallising olivine, as proposed by Kelemen et al. (1995).

Porphyroclastic mantle pyroxenes in the impregnated peridotites from Monte Maggiore, Lanzo and Internal Ligurides have unusual trace element compositions, significantly enriched in many trace elements (i.e. REE, Ti, Sc, V, Zr, Y), with respect to porphyroclastic pyroxenes in the spinel lherzolites from the same peridotite body and to clinopyroxenes in equilibrium with MORB melts. Enriched clinopyroxenes show convex-upward REE patterns with a significant REE enrichment (MREE up to 30×C1 in some Lanzo samples), and both ortho- and clinopyroxenes frequently show a negative Eu_N anomaly (Figure 1). At Lanzo, clinopyroxenes in the same sample, as both mantle porphyroclasts and interstitial magmatic grains, show remarkably similar trace element compositions. Mantle and magmatic orthopyroxenes follow the same trend as clinopyroxenes. These features suggest that; 1) mantle and magmatic pyroxenes attained trace element equilibrium with the percolating melts, 2) the percolating melts changed their major/trace element composition via melt/rock reaction (olivine crystallisation and pyroxenes dissolution) during reactive porous flow in the mantle column.

At Lanzo, the trace element budgets of magmatic pyroxenes and plagioclases in the gabbroic microgranular aggregates of the impregnated peridotites vary from sample to sample. In fact, clinopyroxenes show a negative LREE fractionation which changes from slight to strong (La_N/Sm_N = from 0.19 to 0.01): this variation is accompanied by a progressive change, from positive to negative, of the plagioclase LREE fractionation (La_N/Sm_N = from 3.60 to 0.10; La_N/Pr_N = from 1.20 to 0.07). Plagioclase shows a parallel decrease in the Sr (from 154 to 5.7 ppm) and Na (from An67% to An87%) contents. The above compositional variations of the main magmatic minerals of the impregnated peridotites is, most probably, related to the progressive change in the composition of the primary melts, from slightly Sr-LREE enriched to significantly depleted, due to the progressive increase in the melting degrees during near-fractional melting of the upwelling asthenospheric mantle.

Melt impregnation produced the **chemical refertilization** of the lithospheric mantle, i.e. addition of basaltic components (the gabbroic microgranular aggregates) and significant trace element enrichment of the mantle minerals, when attaining geochemical equilibrium with the compositionally modified, impregnating melts. Thermometric estimates based on trace element (i.e. Sc and V) distribution between coexisting pyroxenes (Seitz et al., 1999) of the impregnated peridotites, indicate that magmatic aggregates and mantle porphyroclasts records the same equilibrium temperatures (about 1250–1300 ºC) (Piccardo et al., 2002; Müntener & Piccardo, 2003). This fact is in favour of a significant **heating** of the lithospheric mantle during melt percolation and impregnation. The high temperatures favoured attainment of the trace element equilibrium between melts and mantle minerals. In conclusion, the lithospheric mantle was subjected to significant **thermochemical erosion** (chemical refertilization plus heating) during the asthenosphere/ lithosphere interaction which accompanied the rifting stage of the basin (Piccardo et al., 2002; Müntener and Piccardo, 2003).

Melt percolation by focussed porous flow

Field evidence at Lanzo and Monte Maggiore indicates that the impregnated peridotites are locally cut and replaced by bodies and channels of spinel dunites (Boudier and Nicolas, 1972, Boudier, 1978; Piccardo et al., 2002; Müntener and Piccardo, 2003). This suggests that, after impregnation, the melt infiltration processes in the lithospheric mantle were focused in the dunite channels and the migration mechanisms changed from diffuse to channelled porous flow.

The melts migrating in dunite channels crystallized, small interstitial clinopyroxenes, at olivine triple points, plagioclase+clinopyroxene films surrounding olivine crystals and gabbroic veinlets with fuzzy contacts. At Lanzo, clinopyroxenes in dunites have almost flat M- HREE patterns (at <10×C1) and moderate LREE depletion, without any trace element enrichment. Geochemical modelling indicates that melts percolating in dunites were significantly different from the melts which produced impregnation in spinel peridotites. They were primary liquids similar to MORB (Piccardo et al., 2002; Müntener and Piccardo, 2003), which remain unmodified during upwelling and escaped melt/peridotite reaction. This is consistent with the idea that melts migrating in dunite channels may be not affected by melt/rock reaction and that focused flow in restricted conduits is required for MORB extraction from the mantle (Kelemen et al., 1995, 1997).

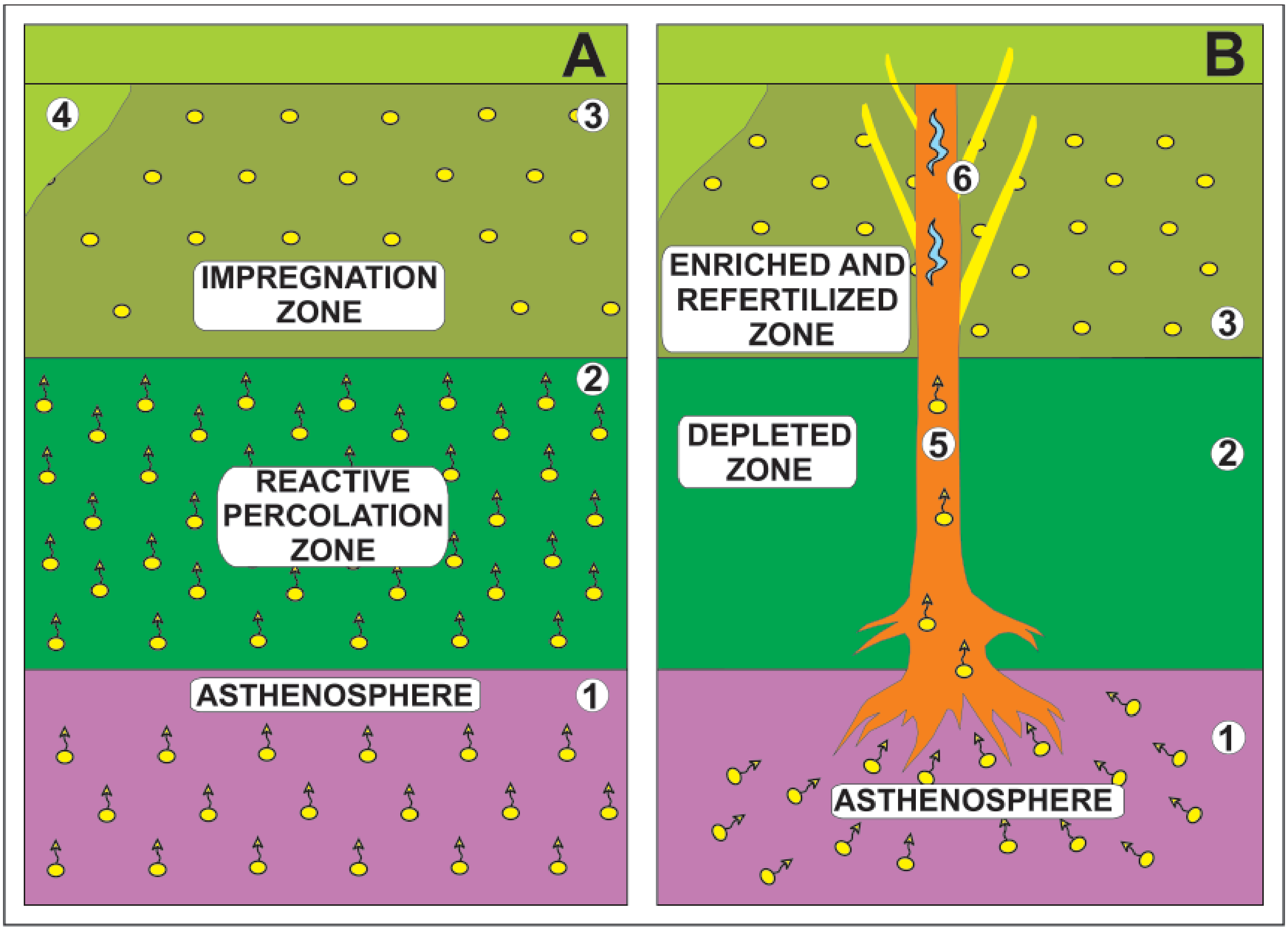

***Figure 1** Asthenosphere/lithosphere interaction via upward migration of hot asthenospheric melts, as evidenced by field, petrologic and geochemical data from the Lanzo peridotite massif.*

(A) Decompressional melting in the rising asthenosphere (1) produced variably enriched/depleted melts, most probably consisting of single melt fractions after fractional melting processes. These single melt increments migrated isolatedly through the overlying mantle lithosphere, consisting of spinel-facies depleted lherzolites, recrystallised at temperature of about 1000–1100˚C (4). Melts migrating by reactive porous flow heated the lower lithosphere to 1250–1300˚C and interacted with the peridotites (dissolving pyroxenes and crystallising olivine) (2): migrating melts attained pyroxenes- and silica-saturation and were, most probably, enriched in most trace elements, whereas lithospheric peridotites were transformed to reactive harburgites. At shallower levels (3), where the competing effects between heating by melt percolation and cooling by conductive heat loss caused interstitial melt crystallisation within the percolated mantle, spinel peridotites were strongly impregnated and refertilized, by addition of basaltic components (i.e. interstial micro-granular gabbroic aggregates) and by trace element enrichment of mantle minerals, when attaining trace element equilibrium with the impregnating melts.

(B) Following the early asthenosphere/lithosphere interaction by melt percolation, which formed a lower section of reactive harzburgites (depleted zone) and an upper section of impregnated peridotites (enriched and refertilized zone), the single melt fractions were, most probably, aggregated in the asthenosphere to form MORB-type aggregated melts, which migrated upward via focused porous flow within dunite channels (5). The dunite channels acted as preferential ways for upwards migration of melts: when cooling down to their liquidus temperatures, the migrating melts crystallised clinopyroxene+plagioclase, interstitial to the large olivine crystals. In places, clinopyroxene and plagioclase form gabbroic patches, veinlets and dykelets (6), which intrude, with fuzzy contacts, the surrounding impregnated peridotites. These features suggest that the interstitial crystallisation in the dunite channels caused progressive clogging of the melt migration ways and forced the migrating melts to intrude along cracks and fractures.

3) Early veining and dyking of gabbroic material

The dunite channels acted as preferential ways for upwards migration of melts: when cooling down to their liquidus temperatures, the migrating melts crystallised interstitial plagioclase+ clinopyroxene, gabbroic veinlets and dykelets, which intrude the surrounding impregnated peridotites, showing fuzzy contacts. These features suggest that the interstitial crystallisation in the dunite channels caused progressive clogging of the melt migration ways and forced the migrating melts to intrude along cracks and fractures.

Accordingly, following diffuse reactive porous flow, pervasive melt impregnation and focused flow in dunite channels, the melt migration mechanisms changed from diffuse/focused percolation to intrusion. At Lanzo, cm-scale gabbroic veins and dykelets were intruded in the impregnated peridotites, at Monte Maggiore, the early intrusion produced m-scale cumulate pods and cm-scale gabbroic dykelets. The gabbroic veins and dykelets in both peridotite massifs, and the cumulate pods at Monte Maggiore, have Mg-rich olivine (Fo90) and pyroxenes (Mg#90–92): they were formed, accordingly, by early crystallisation of rather primitive melts. These magmatic pyroxenes, differently to the magmatic pyroxenes of the previous impregnation, don't show any trace element enrichment, indicating that the intruding melts escaped significant melt/rock reaction during upwelling. The gabbroic dykelets at Lanzo South

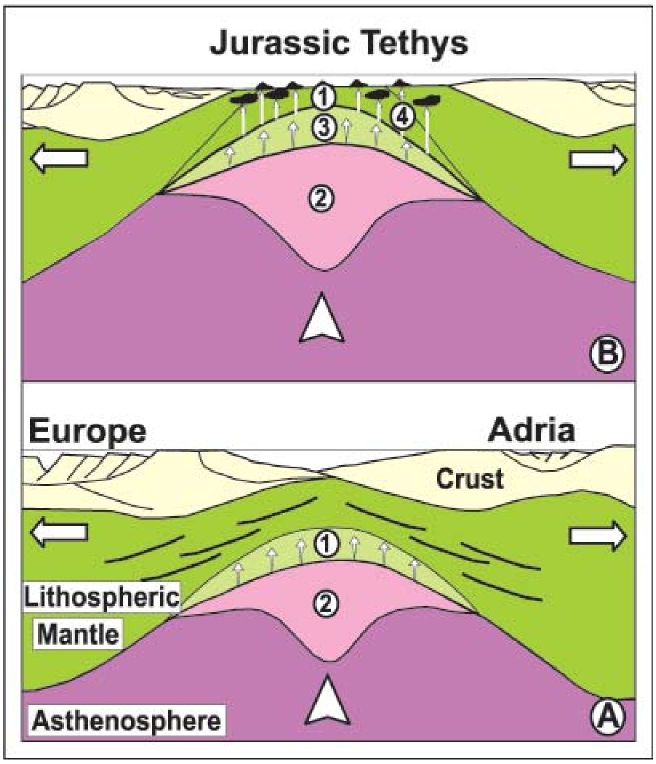

Figure 2 Jurassic rifting and drifting stages of the Ligurian Tethys. (A) Continental rifting: early percolation and impregnation of the pre-Jurassic subcontinental mantle lithosphere (Lanzo-type) (1) by ascending hot melts which are formed in the upwelling molten asthenosphere (2);
(B) Oceanic drifting: percolation and impregnation of Jurassic oceanic mantle lithosphere (M. Maggiore-type) (3), intrusion and extrusion of MORB-type magmas (4) (not to scale). The modified mantle lithosphere [(1) and (3)] most probably consists of i) a lower section of reactive harburgites, and of ii) an upper section of impregnated and refertilized peridotites.

have plagioclases with relatively high Na (An54–66%) and very high Sr (500–750 ppm) contents, whereas cumulates and dykelets at Monte Maggiore have Ca-rich (An88–96%), and extremely Sr-poor (20–30 ppm) plagioclases. Clinopyroxenes have almost flat REE patterns in the M-HREE region (at <10×C1), but significantly different LREE negative fractionation and incompatible trace element contents. Geochemical modelling indicates that: i) the primary melts of the Lanzo dykelets most probably correspond to low degrees (2–3%), melt increments, or low degrees aggregated melts, after near-fractional melting of an asthenospheric mantle source, ii) the primary melts of the Monte Maggiore dykelets and cumulates correspond to higher degrees (6–7%), strongly depleted melt increments, or higher degrees depleted aggregated melts, after near-fractional melting.

Above evidence suggests that the earliest melts which percolated and impregnated the ophiolitic peridotites were, most probably, single melt increments which escaped aggregation and survived unmixed during ascent; they migrated through the lithospheric mantle column by reactive porous flow, attaining pyroxene(s) saturation. Subsequent focused percolation in dunite channels and intrusion along fractures allowed both depleted melt fractions (Monte Maggiore) and MORB-type melts (Lanzo) to ascend in the lithospheric mantle without significant melt/peridotite reaction.

4) Intrusion of gabbroic dykes and bodies

The Alpine-Apennine ophiolitic peridotites are intruded by meter-wide gabbroic dykes and km-scale gabbroic bodies, showing sharp contacts and chilled margins to the country peridotite and cutting across all previous mantle and magmatic structures. They vary in composition from rather primitive troctolite to Mg-gabbros to Fe-Ti-gabbros, and rare plagiogranites. Computed liquids in equilibrium with clinopyroxenes from the most primitive olivine gabbros are closely similar to average aggregated MORBs. MORB-type evolved magmas were intruded when the lithospheric mantle was cold and brittle, at shallow levels in the conductive lithosphere.

Discussion and conclusion

Present knowledge on the Alpine-Apennine ophiolitic peridotites allow to constrain some major steps in the formation of the Jurassic Ligurian Tethys ocean, and to outline the mantle processes which accompanied the geodynamic evolution during rifting and opening of the basin.

After partial melting and accretion to the thermal lithosphere, the mantle sections of the Alpine-Apennine ophiolites were exhumed towards the sea-floor during passive extension of the Europe-Adria lithosphere. During lithosphere thinning, the underlying asthenosphere rose up and underwent near-adiabatic decompression melting: the resulting fractional melts migrated through and reacted with the overlying mantle lithosphere.

Alpine-Apennine ophiolitic peridotites record distinct magmatic cycles: (1) the diffuse porous flow and impregnation of depleted/enriched isolated fractional melt increments, which underwent pyroxenes-saturation and modification of the major/trace element composition by melt/rock reaction during upwelling, (2) the focused percolation in dunite channels and the early intrusion of depleted/enriched melt fractions and MORB-type primary melts, and (3) the late intrusion of variably fractionated magmas deriving from aggregated MORB melts.

Chemical depletion/refertilization and thermal erosion of the lithosphere

The chemical and rheological relevance of the enrichment processes in the lithospheric mantle peridotites has been stressed by Menzies and co-workers (1987). The combination of lithosphere extension in a rifting system and the intrusion of magmas from the upwelling asthenosphere have been considered as important factors which may accomplish the thermomechanical erosion of the lithosphere (Davies, 1994). Evidence of km-scale porous melt flow in the Ronda peridotites, related to pervasive infiltration of asthenospheric melts, has been regarded as a volumetrically important process accompanying the thermomechanical erosion of the lower lithosphere by the upwelling asthenosphere (Van der Wall and Bodinier, 1996). Infiltration of asthenospheric magmas in the lower lithosphere has been considered a peculiar feature of the early stages of continental rifting in eastern Africa (Bedini et al., 1997). The "reactive" formation of harburgites via melt percolation, as a consequence of lithosphere-asthenosphere interaction during lithospheric thinning, has been recently documented by Xu et al. (2003).

The outlined features of the Alpine-Apennine ophiolitic peridotites evidence that significant asthenosphere/lithosphere interaction followed the rifting stages of the Ligurian Tethys: the upward migration of hot asthenospheric melts caused significant chemical modifications (depletion/refertilization) and thermal erosion of the lithospheric mantle. Melt/peridotite reaction formed **a lower zone of depleted reactive harzburgites or cpx-poor lherzolites** at the base of the lithospheric mantle column. During melt upward migration, the competing effects of heating by melt percolation and cooling by ongoing exhumation caused the interstitial crystallisation of early liquidus phases in the percolated lithospheric mantle and the progressive clogging of the melt channels. This formed an **upper zone of refertilized, impregnated lherzolites**.

The thermochemical erosion of the lower lithospheric mantle was a fundamental step in the evolution of the Ligure-Piemontese basin. The thermal softening of the extending lithosphere could have played an important role in the dynamics of the rifting system during transition from passive lithosphere extension to active oceanic drifting in a slow spreading system.

Melt dynamics and mantle rheology

The change of the melt migration mechanisms from pervasive to focused porous flow and finally dyking suggests that the rheology of the lithospheric mantle was modified by the lithosphere/asthenosphere interaction during lithosphere extension and thinning. The relatively cold peridotites of the lithospheric mantle column attained more hot and plastic, asthenospheric characteristics, when they underwent thermochemical erosion during melt percolation.

Diffuse melt crystallisation under progressive cooling stopped porous flow and enhanced melt focusing into cracks and fractures. During ongoing upwelling in the thermal lithosphere, the thermochemically modified lithospheric mantle went back to more cold and brittle conditions, when it was subjected to increasing conductive heat loss. This is well consistent with the progressive thinning of the Europe-Adria lithosphere during the extension which governed the rifting stage of the Ligure-Piemontese basin.

The variations in rheology of the lithospheric mantle during rifting was accompanied by changes in the melt dynamics in the asthenosphere. During the early melting stages of the upwelling asthenosphere, single melt increments produced by fractional melting escaped aggregation, survived unmixedly and migrated isolatedly. Subsequently, the melt fractions were more efficiently mixed and completely aggregated to form MORB magmas. These aggregated MORBs underwent differentiation, most probably, within small magma chambers: variably fractionated magmas were formed and intruded as Mg- to Fe-Ti-rich gabbroic dykes with MORB affinity and, later, were extruded as variably evolved MORB-type lava flows.

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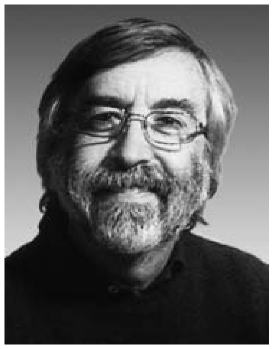
Giovanni B. Piccardo, *Full Professor of Petrology at the University of Genova since 1980. Research activity in the fields of Petrology and Geochemistry of (1) Ophiolites from the Northern Apennine and the Western Alps; (2) Eclogites and HP metamorphic ophiolites from the Voltri Massif and the Western Alps; (3) Mantle Peridotites from oceanic and extensional settings (Red Sea, Northern Apennine and Lanzo) and mantle xenoliths in alkaline lavas (Eritrea).*



by Roberto Compagnoni

HP metamorphic belt of the western Alps

The University of Turin, Dept. of Mineralogical and Petrological Sciences, Via Valperga Caluso, 35, 10125 TORINO, Italy.
E-mail: *roberto.compagnoni@unito.it*

The understanding of the subduction-related processes benefited by the studies of the high-pressure (HP) metamorphic rocks from the western Alps. The most stimulating information was obtained from the inner part of the western Alpine belt, where most tectonic units show an early Alpine eclogite-facies recrystallisation. This is especially true for the Austroalpine Sesia Zone and the Penninic Dora-Maira massif. From the Sesia zone, which consists of a wide spectrum of continental crust lithologies recrystallised to quartz-eclogite-facies mineral assemblages, the first finding of a jadeite-bearing meta-granitoid has been described, supporting evidence that even continental crust may subduct into the mantle. From the Dora-Maira massif the first occurrence of regional metamorphic coesite has been reported, opening the new fertile field of the ultrahigh-pressure metamorphism (UHPM), which is now becoming the rule in the collisional orogenic belts.

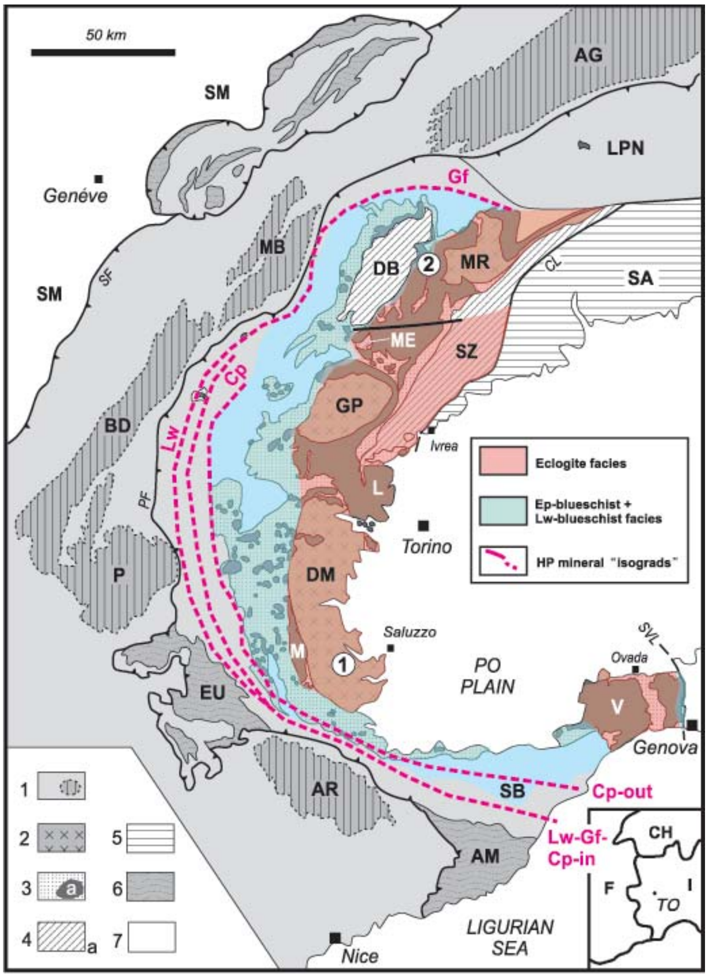


Figure 1** Simplified structural sketch-map of the Western Alps. 1: Jura, Helvetic Domain and external Penninic Domain. The dashed line contours the External Crystalline Massifs (AR: Argentera, P: Pelvoux, BD: Belledonne, MB: Mont Blanc-Aiguilles-Rouges, AG: Aar-Gotthard). SB: Grand St. Bernard Zone, LPN: lower Penninic nappes. 2: Internal Crystalline Massifs of the Penninic Domain (MR: Monte Rosa, GP: Gran Paradiso, DM: Dora-Maira). 3: Piemonte Zone (L: Lanzo Massif; M: Monviso Massif; V: Voltri Massif) and a) main meta-ophiolite bodies. 4: Austroalpine Domain (DB: Dent Blanche nappe, ME: Monte Emilius, SZ: Sesia Zone). 5: Southalpine Domain (SA). 6: Helminthoid Flysch nappes (EU: Embrunais-Ubaye, AM: Alpes maritimes). 7: Swiss Molasse (SM), Po Plain and Piemontese-Ligurian Tertiary basin. CL: Canavese line; SVL: Sestri-Voltaggio line; SF: Subalpine frontal thrust; PF: Penninic thrust front. (1) UHP Brossasco-Isasca Unit, Dora-Maira Massif; (2) UHP Lago di Cignana unit, upper Valtournenche, Piemonte Zone. Gf-in, from Bocquet (1974); Lw-in, and Cp-in and Cp-out from Goffé and Chopin (1986).
In the inset: CH: Switzerland; F: France; I: Italy; TO: Torino.

Introduction

The western Alps extends from the Sestri-Voltaggio tectonic Line, which separates it from the non-metamorphic rocks of the Apennine chain, to the Lower Penninic Nappes of the Lepontine dome (LPN, Figure 1). On the internal side, the western Alps are bounded by the Quaternary post-orogenic clastic deposits of the Po Plain up to about the latitude of Torino, and from there northwards by the Canavese tectonic Line (CL), the SW extension of the Insubric Line (or Periadriatic lineament), which separates the pre-Alpine domain of the Southern Alps (Ivrea Zone + Strona-Ceneri Zone) from the western Alpine chain reworked during the Alpine orogeny.

The HP belt of the western Alps comprises most tectonic units of the Penninic and Austroalpine domains (Figure 1).

The Penninic Domain is a heterogeneous realm, which consists of both continent- and ocean-derived tectonic units (see Dal Piaz et al., this issue). The continent-derived units are (from the internal concave side toward the external convex side): the *Austroalpine Sesia-Lanzo Zone—Dent Blanche nappe system*, the "*Internal Crystalline Massifs*" of Monte Rosa (MR), Gran Paradiso (GP) and Dora-Maira (DM), and the *Briançonnais Zone* (or Grand Saint Bernard nappe system), which overthrust the Helvetic Domain: this tectonic boundary is known as "Penninic Thrust Front" (PTF) (Figure 1). The units derived from the Mesozoic Tethys ocean make up the *Piemonte Zone*, also named "Zone of calc-schist (French: "*schistes lustrés*") with meta-ophiolite", which consists of a number of thrust sheets with different high-pressure metamorphic recrystallisations (Figure 1).

The Austroalpine domain includes the Sesia-Lanzo zone (in the following referred to as the Sesia Zone) and the Dent Blanche Nappe system, which are fragments of Variscan granulite to amphibolite-facies continental crust intruded by late-Variscan granitoids, derived from the Southalpine (or Insubric) plate. The "Eclogitic Micaschist Complex" (EMC) of the Sesia zone and the Monte Emilius klippe of the Dent Blanche nappe are the best preserved examples of continental crust basement recrystallised under eclogite-facies conditions (Compagnoni, 1977; Dal Piaz et al., 1983).

Due to the widespread occurrence of unaltered HP metamorphic rocks, the western Alps have been the ideal area for the study of this type of metamorphism. Contributions to the knowledge of the HP metamorphism go back to the end of 19th century. The first eclogite-facies metapelite (named "eclogitic micaschist") was reported by Stella (1894) from the EMC of the Sesia zone and later studied by Franchi (1900, 1902), who also described the reaction jadeite+quartz⇒albite (Franchi, 1902); the first eclogitised pillowed basalts were described by Bearth (1959) from the meta-ophiolites of the Piemonte zone from the Zermatt valley; the first eclogite-facies jadeite-bearing metagranite was reported by Compagnoni and Maffeo (1973) at Mt. Mucrone from the EMC of the Sesia zone; the first coesite in continental crust was reported by Chopin (1984) from the southern Dora-Maira massif; and the petrologic importance of Mg-Fe-carpholite$_{ss}$ for the blueschist-facies was first recognised by Goffé and Chopin (1986), studying Briançonnais lithologies previously considered unimportant for geobarometric estimates.

And finally, let's remember that the high-density, hard and tough Neolithic stone implements, excavated from all over the western Europe, are made of eclogite and jadeitite derived from the HP meta-ophiolites of the Piemonte zone (Compagnoni et al., 1996).

Regional distribution of the HP metamorphism

Since Bearth's (1962) pioneering work, a number of attempts were made to trace the "isograds" of the HP metamorphism (for a review see Desmons et al., 1999). Many published and unpublished petrologic data were first summarised in the *Metamorphic Map of the Alps* (sheet 17 of the 1:1,000,000 *Metamorphic Map of Europe*, edited by Zwart, 1973) and then in the 1:500,000 *Map of the Alpine Metamorphism* in the *New Metamorphic Map of the Alps* (Frey et al., 1999).

Broadly speaking, at a regional scale the HP metamorphism in the western Alps includes from E to W eclogite-facies, epidote- and lawsonite-blueschist facies, and lawsonite-albite-chlorite facies rocks (Figure 1). The quartz-eclogite facies units prevail, but two small coesite-eclogite facies units have been recognised: the Brossasco-Isasca unit (BIU) from the southern Dora-Maira massif (Figures 1 and 2) and the Lago di Cignana unit from the Piemonte zone (Figure 1). This large-scale metamorphic zoning is considered as evidence showing that the northern European plate was subducted below the Adrian (or Insubric) plate (Ernst, 1971; Dal Piaz et al., 1972).

"Isograds" for most significant blueschist-facies minerals, such as glucophane-in (Gf-in), lawsonite-in (Lw-in), and carpholite-in (Cp-in)-and-out (Cp-out) have been traced in the western Alpine belt, and provide a large-scale idea of their mineral zoneography in the most external Pennine zone (Figure 1). However, the increase of detailed petrographic studies showed that the HP mineral distribution is really much more complicated than originally supposed. For example, in the Aosta valley, the tectonometamorphic setting of the Piemonte zone consists of a quartz-eclogite facies unit or composite lithotectonic unit ("Zermatt-Saas zone": Bearth, 1967) overlain by an epidote-blueschist facies unit or composite lithotectonic unit association ("Combin zone"). The main tectonic contact between the two zones is marked by the presence of a thin coesite-eclogite facies meta-ophiolitic unit ("Lago di Cignana Unit": Reinecke, 1991) and of several Austroalpine continental crust slices, showing an Alpine quartz eclogite-facies overprint. This complex tectonometamorphic setting indicates the presence of a polyphase tectonic evolution, involving both compressional and extensional large-scale processes. Figure 2 illustrates the southern Dora-Maira massif, where a tectonic thrust sheet with Alpine coesite-eclogite facies overprint (Brossasco-Isasca Unit: T≈750˚C, P≈3.5 GPa) is sandwiched between two quartz-eclogite facies units (San Chiaffredo and Rocca Solei units: ≈550˚C, ≈1.5 GPa), bounded in turn by an epidote-blueschist facies unit (Pinerolo unit: ≈450˚C, ≈ 0.8 GPa) and a quartz-eclogite-facies unit (Dronero-Sampeyre unit: ≈550˚C, ≈1.5 GPa) (Compagnoni and Rolfo, 2003 with ref. therein).

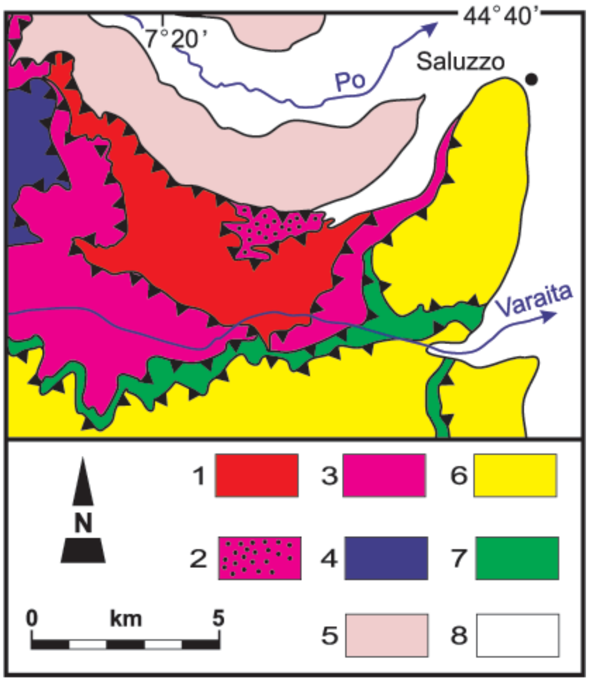


***Figure 2** Structural sketch map of southern Dora-Maira Massif. Continent-derived Dora-Maira Massif: Coesite-eclogite facies Brossasco-Isasca Unit (1); Quartz-eclogite facies San Chiaffredo Unit (2); Quartz-eclogite facies Rocca Solei Unit (3); Pre-Alpine basement and Permo-Triassic cover unit (4); Epidote-blueschist facies Pinerolo Unit (5); Quartz-eclogite facies Dronero-Sampeyre Unit (6); Ocean-derived Piemonte Zone: Epidote-blueschist facies metaophiolites and oceanic metasediments (7); Quaternary alluvial deposits (8).*

Peculiarities of the HP metamorphism

The Alpine eclogite-facies recrystallisation is especially well developed in the Eclogitic Micaschist Complex of the internal Sesia Zone, in the Monte Emilius klippe of the Dent Blanche nappe, and in the meta-ophiolites of the Piemonte Zone, while relics of eclogite-facies lithologies occur in the Internal Crystalline Massifs of Monte Rosa, Gran Paradiso, and Dora-Maira (Figures 1 and 2). Most units have been recrystallised under quartz-eclogite-facies conditions, except for the two small units of Brossasco-Isasca (BIU) (T=730 ˚C and P≥3.3 GPa) from the southern Dora-Maira Massif (Figure 2) and Lago di Cignana (T≈600 ˚C and P=2.6–2.8 GPa) from the Piemonte zone, which contain coesite relics.

HP metamorphism in the oceanic lithosphere

Historically important are the pillow metabasalts of the Piemonte zone (Zermatt Saas zone: see later on) described by Bearth (1959) for the Zermatt valley, in which the dry pillow core was converted to a coarse-grained bimineralic eclogite assemblage, whereas the hydrous rim was replaced by a glaucophane-rich rock. In the Monviso ophiolite of the Cottian Alps (Figure 1), a complete section of oceanic crust is exposed, including poorly deformed isotropic to layered cumulus gabbros, massive to pillowed basalts, and basaltic dykes, which recrystallised under eclogite-facies conditions (Schwartz et al., 2000).

The Lanzo ultramafic massif (Figure 1) is the largest portion of upper mantle peridotite exposed in the western Alps, which experienced Alpine eclogite facies metamorphism with P in excess of 2.0 GPa and T= 550˚C. Its central less-deformed portion still consists of spinel-plagioclase lherzolite and minor harzburgite—so fresh that the high-T petrology and deformation may be studied (Boudier, 1976), and gabbro dykes are locally found which may preserve the original Ca-rich plagioclase. By contrast, its marginal portion has been converted to sheared serpentinite, which contains the eclogite-facies assemblage: antigorite, metamorphic olivine (Fe-richer than peridotitic olivine), clinohumite (mostly red-brown titanian clinohumite) magnetite, Fe-Ti alloys±diopside±Mg-chlorite±apatite. Similar mineral assemblages may be found in the antigorite serpentinite of the whole eclogite-facies internal Piemonte zone, where olivine+Ti-clinohumite+Mg-chlorite±apatite metamorphic veins occur, which in part at least developed at the expense of primary igneous dykes rich in ilmenite (Scambelluri and Rampone, 1999).

Poorly deformed metagabbros are the best-preserved ophiolitic lithologies, since the coarse grain-size and dry composition favoured preservation of the igneous protolith. Let's first mention the Allalin metagabbro (Bearth, 1967) from the Swiss Valais, in which all reaction steps from the original olivine gabbronorite to a coarse-grained eclogite may be observed (Meyer, 1983). The poorly-deformed metagabbros are characterised by a more complex mineral association than the sheared gabbros, because pseudomorphous reactions develop after each igneous mineral and coronitic reactions at the original boundaries between igneous minerals, respectively. For example, in coronitic Mg-Al-metagabbros, diopside-rich omphacite+talc develops after igneous clinopyroxene, jadeite-rich omphacite+garnet+zoisite±quartz after plagioclase, and omphacite+talc+Na-amphibole after olivine. In coronitic Fe-Ti-metagabbros ferrian omphacite develops after igneous clinopyroxene, omphacite+garnet after plagioclase, and Na-amphibole+rutile after the hornblende-ilmenite intergrowth (Messiga and Scambelluri, 1988). In the pervasively deformed portions, Fe-Ti-metagabbros recrystallised to omphacite+garnet+rutile±glaucophane±epidote assemblages, whereas Mg-Al-metagabbros recrystallised to omphacite+garnet+rutile±chloritoid±talc±chlorite assemblages. In the so-called "smaragdite" metagabbros, the original igneous Cr-diopside has been replaced by bright-green "smaragdite" (i.e., a Cr-bearing omphacite±tremolite±talc), plagioclase by jadeite-rich pyroxene+zoisite, olivine by talc±tremolite, and ilmenite by rutile.

Mineral assemblages and mineral compositions are closely related to bulk rock chemistry: for example, in gabbros with high Mg/(Mg+Fe) ratio (such as Mg-Al-gabbros), Mg-Al minerals (such as chlorite and/or chloritoid) form, whereas the bimineralic omphacite+garnet (+rutile) eclogite assemblage is produced only in Al-poor, Fe-Ti-rich gabbros. Similarly, garnet is almandine-richer in Fe-Ti-metagabbros, whereas it is pyrope-richer in Mg-Al-metagabbros.

Lawsonite-eclogites are extremely rare in the western Alpine belt and confined to some small tectonic units of Liguria. On the contrary, the widespread occurrence of lozenge-shaped white-mica+paragonite pseudomorphs up to about two cm-long indicates that porphyroblastic lawsonite was common during prograde eclogite-facies metamorphism.

In eclogites, veins locally occur, which indicate that the eclogite-facies metamorphism developed in the presence of a hydrous fluid phase. However, the fluid flow was limited during eclogite-facies metamorphism and fluid was mainly released by devolatilisation reactions of dense hydrous silicates (such as lawsonite) or during plastic crystal flow (Philippot, 1993).

Associated with eclogites derived from Fe-Ti-gabbros, very minor felsic rocks (plagiogranite or sodagranite) are locally found, which exceptionally occur as km-sized bodies (Castelli et al., 2002 with ref. therein): they are the only lithology among meta-ophiolites containing the association jadeite+quartz±garnet.

In lithologies of unusual compositions belonging to the layered gabbros sequence, Cr-Mg-chloritoid and other Cr-bearing minerals have been found, most likely formed at the expense of chromite-rich bands of the cumulus layered gabbros (Messiga et al., 1999).

Rodingites, derived from both gabbroic or basaltic dykes, are ubiquitous within most serpentinised peridotites and have a typical paragenesis of diopside, vesuvianite, ugranditic garnet, epidote, and Mg-chlorite; however, the local presence of Alm-rich garnet and omphacite indicates that some of them formed during HP metamorphism.

In the calc-schists and associated metasediments, eclogite-facies assemblages are more difficult to identify because typically they are more easily retrogressed. However, for suitable compositions, the eclogite assemblage appears to be phengite+Alm-rich garnet+rutile±paragonite±zoisite±chloritoid±porphyroblastic lawsonite, mostly replaced by paragonite+epidote pseudomorphs. Noteworthy is also the Mn-deposit of Praborna, Saint Marcel (Aosta Valley), where a number of eclogite-facies Mn-bearing minerals or mineral varieties have been described (Martin and Kienast, 1987).

HP metamorphism in continental crust

The best preserved portion of eclogitised continental crust in the western Alps is the Eclogitic Micaschist Complex of the Sesia Zone. In this complex, derived from a Variscan amphibolite-facies basement intruded by late-Variscan granitoids (Compagnoni, 1977), the whole spectrum of continental lithologies from paragneiss to orthogneiss, from marble to metabasic rock, may contain one or more quartz-eclogite facies minerals, such as garnet, Na-pyroxene (omphacite and jadeite), high-celadonite phengite (3T polytype), paragonite, glaucophane, zoisite, and chloritoid.

Well-preserved relics of Variscan crystalline basement also occur in the Internal Crystalline Massifs. The best examples of undeformed lithologies are the coronitic metagranitoids, which are exposed in the central Gran Paradiso massif, at Monte Mucrone in the "Eclogitic micaschist Complex" (EMC) of the Sesia zone, and in the Brossasco-Isasca unit (BIU) of the Dora-Maira massif. In all three units, metabasic rocks have been converted to the eclogite assemblage, but metagranitoids show significant differences: in the Gran Paradiso metagranitoids (about 400–500 ˚C and about 1.0 GPa) the igneous structure and mineralogy are mainly preserved; in the Monte Mucrone metagranitoids, Sesia Zone (about 550 ºC and 1.6–1.8 GPa) igneous quartz and K-feldspar are preserved, but plagioclase is pseudomorphically replaced by jadeite+zoisite+quartz±kyanite±K-feldspar; in the BIU metagranitoids, Dora-Maira massif (about 750 ˚C and 3.5 GPa) igneous quartz is also replaced by a granoblastic polygonal aggregate of metamorphic quartz, inverted from former peak pressure coesite (Biino and Compagnoni, 1992).

However, the most interesting continental crust lithology of the western Alpine HP belt is the "silvery micaschist", which occurs as cm- to m-thick lens-like layers within the orthogneiss of the Internal Crystalline Massifs and is interpreted as a metasomatic rock formed at the expense of a granitoid protolith along shear zones (Compagnoni and Hirajima, 2001, with ref. therein). This rock, which is important petrologically because its phase relationships can be modelled in the relatively simple KMASH system (Chopin, 1981), is usually referred to as 'whiteschist" for its characteristic kyanite+talc (or talc+phengite) assemblage, indicative of pressures in excess of about 1.0 GPa (Schreyer, 1977). The best-known and unique white schist is that from the Dora-Maira massif first described by Chopin (1984), which contains in addition to kyanite, phengite, talc, rutile, and accessories, a pale-pink idioblastic pyrope (up to 98 mole% of the Mg-Al end-member). Pyrope crystals are up to 20 cm across and host, in addition to abundant kyanite, unusually large coesite relics. A number of new minerals (such as bearthite, ellenbergerite and phosphellenbergerite) or minerals of unusual composition (such as magnesiodumortierite, magnesiochoritoid, magnesiostaurolite), suggestive of UHPM conditions, have been found in the pyrope crystals (Chopin and Ferraris, 2003 with ref. therein).

Age of HP metamorphism

Up to the end of the 1980s, most radiometric ages relevant to the HP metamorphism from the western Alps, recognised as indicative of an Early-Alpine or Eoalpine metamorphic event fell into the range of 140–60 Ma, further subdivided into an early eclogite-facies stage (140–85 Ma) and a later blueschist-facies and cooling stage (85–60 Ma) (for a review see: Hunziker et al., 1992). The younger Tertiary ages (around 50 Ma) were interpreted as the evidence of two episodes of HP metamorphism (Monié et al., 1989).

The first work, which suggested the possibility of a Tertiary age for the eclogite-facies metamorphism from the western Alps, was that by Tilton et al. (1991, with ref. therein) who dated U-Pb on minerals from the UHPM Brossasco-Isasca Unit (BIU), southern Dora-Maira massif (DMM). However, the major change in the age determination of the eclogite-facies metamorphism in the western Alps was brought about by the *in-situ* U-Pb dating of high closure-temperature minerals, such as zircon, with the Sensitive High Resolution Ion-Microprobe (SHRIMP), assisted by cathodoluminescence imaging (Rubatto et al. 2003, with ref. therein). The most complete geochronological work was done on the UHPM rocks of the southern Dora-Maira massif, previously studied by Tilton et al. (1991) (Gebauer et al., 1997 with ref. therein): the new U-Pb ages, consistently fell at the Eocene-Oligocene boundary (~35 Ma) and were later confirmed by the results from Lu-Hf dating of garnet (Duchêne et al., 1997).

At present, with the exception of the Cretaceous age (~ 65 Ma) of the quartz eclogite-facies metamorphism of the Sesia zone (Rubatto et al. 2003), other U-Pb SHRIMP ages on the western Alps eclogites are all Tertiary (35–45 Ma).

P-T-t paths of the HP metamorphism

The tectonic units of the western Alps recrystallised under eclogite-facies conditions, independently of their peak P conditions, consistently show a similar P-T-path, which is characterised by two thermal peaks. The first thermal peak, which corresponds to the high- to ultrahigh-pressure climax, is followed by significant decompression coupled with moderate cooling. The second thermal peak is at low-P (about 0.4–0.6 GPa), and corresponds to the greenschist-facies event of the Alpine literature. It is followed by significant cooling coupled with moderate decompression. In Figure 3, the P-T paths of the most significant units of continental crust (i.e. Monte Rosa, Sesia zone and BIU) are reported. From their comparison, it is evident that all the P-T paths are clockwise, and that the highest is the peak pressure, the tightest is the P-T loop. This feature, initially difficult to explain, is now believed to be the best evidence that the whole subduction/exhumation process is much faster than formerly supposed. For example, subduction and exhumation speeds of about 2 cm/a have been suggested by Rubatto et al. (1999) for the EMC of the Sesia zone and by Rubatto and Hermann (2001) for the BIU of the Dora-Maira massif.

Conclusions

The existence of two age clusters, initially interpreted as evidence for two metamorphic events of Cretaceous and Tertiary ages, affecting the whole western Alps, have turned out to be the record of separate events in different units. Therefore, the different units were subducted and exhumed at different times, implying that at the same time subduction and exhumation were active in different portions of the orogenic belt. This diachrony of the HP and UHP metamorphism of the Western Alps gives rise also to a nomenclature problem, since

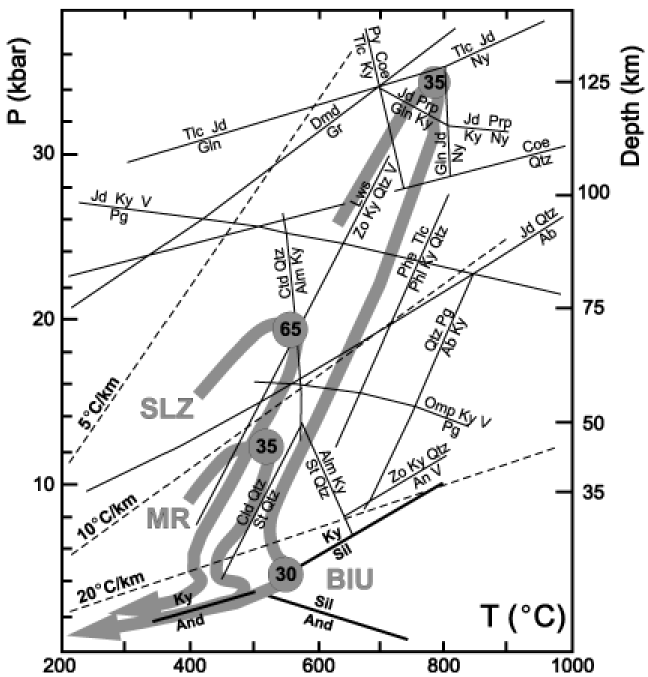


Figure 3 P-T-t paths of representative continental eclogite-facies units from the western Alps. BIU: UHPM Brossasco-Isasca Unit of Penninic Dora-Maira Massif (from Compagnoni et al., 1995, modified). SLZ: Austroalpine continental Sesia Lanzo Zone (from Tropper and Essene, 2002, modified). MR: Penninic Monte Rosa nappe of the Internal Crystalline Massifs (Borghi et al., 1996). Mineral abbreviations after Bucher and Frey (1994). Circled numbers are SHRIMP radiometric ages in Ma for BIU (Rubatto and Hermann, 2001), SLZ (Rubatto et al., 1999), and MR (Rubatto and Gebauer, 1999), respectively.

the so-called early-Alpine or Eoalpine metamorphism, previously considered to be Cretaceous in age and subduction-related, is not a single coeval event throughout the western Alps.

The preservation in both ocean- and continent-derived units of poorly deformed lithologies, which preserve the protolith structure and even part of the primary mineralogy, indicates that at any scale the HP deformation (and metamorphic recrystallisation) mainly occurred along shear zones. This mechanism was favoured by both the relatively low fluid content in continental and oceanic lithologies and the unusually high speed of the whole subduction/exhumation process, which took place in a time span of only a few millions of years. The high speed of the tectonic processes, the scarcity of a free fluid phase, and the continuous cooling during exhumation also account for the local superb preservation of the HP peak mineral assemblages.

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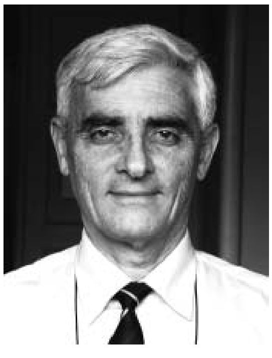
Dr. Roberto Compagnoni *is Professor of Petrology at the University of Torino, where he graduated in Geology in 1964. He has been teacher of Petrology and Metamorphic Petrology at the University of Torino, and at the University of Calabria. His research activity includes the study of metamorphism of meta-ophiolites and crystalline basements from the Western Alps and the Calabrian-Peloritanian Arc of southern Italy, with special attention to the relationships between mineral growth and deformation, and reconstruction of the P-T-t paths. During the last ten years his research has been mainly devoted to the study of the UHP metamorphism of western Alps, and Dabieshan-Sulu, China.*



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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

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

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

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

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

Introduction

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

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

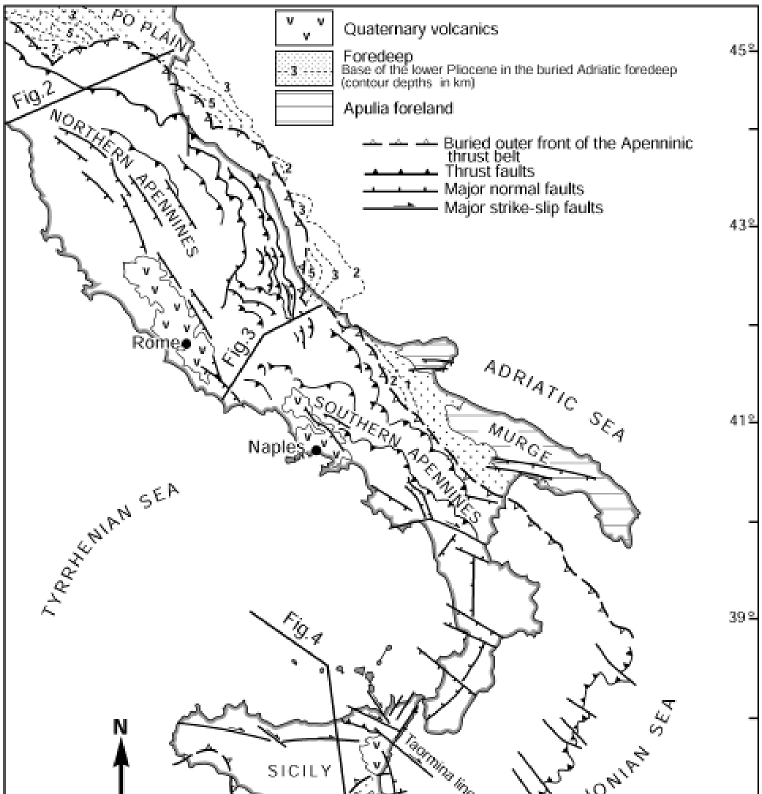

Figure 1 General structural map of the Apennine-Maghrebian chain. Fig.2, Fig.3, Fig.4: locations of the cross-sections of Figures 2–4. (After Ghisetti and Vezzani, 1999, modified)

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

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

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

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

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

The Kabylo-Calabride terranes

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

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

Internal domain

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

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

Liguride units

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

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

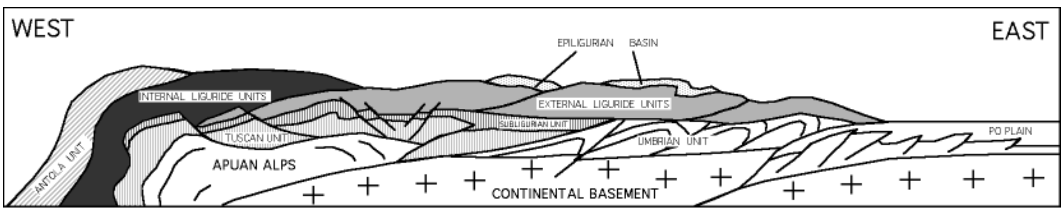

***Figure 2** Schematic cross-section showing the geometric relations among the major structural units of the northern Apennine. For location, see Figure 1. (P. Elter)*

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

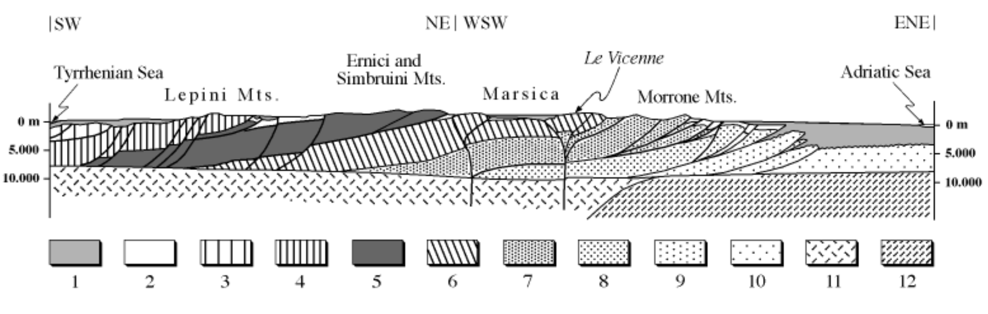

***Figure 3** Schematic cross-section of the central Apennine thrust belt. For location, see Figure 1. 1. marine and continental post-orogenic sedimentary cover (Tyrrhenian side) and intermountain basins (Late Messinian-Quaternary); marine syn- and post-orogenic deposits (Adriatic side: early Pliocene-Quaternary); 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 anti-clockwise rotations was responsible for several major arcuate structures of the Apennine thrust belt (e.g., the Gran Sasso-Mt. Picca thrust, see Ghisetti & Vezzani, 1991; Olevano-Antrodoco-Sibillini Mts. thrust, see Cipollari & Cosentino, 1992).

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

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

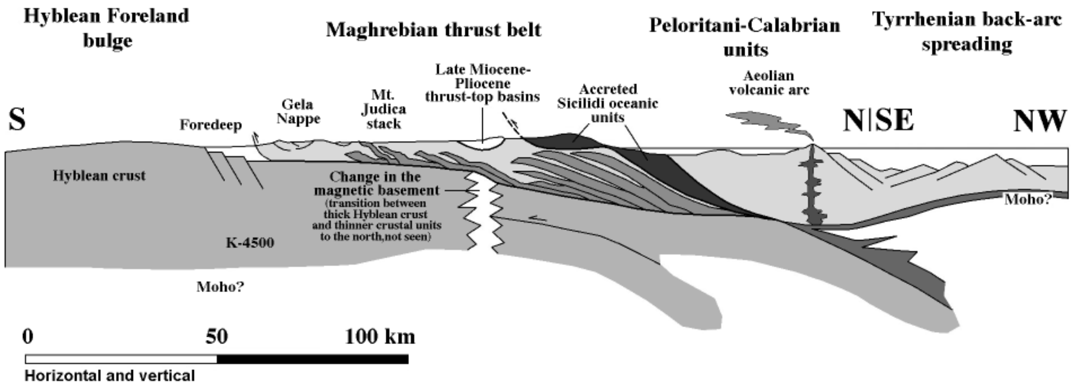

Figure 4 Simplified section across the eastern Sicily, from the Hyblean foreland to the southern Tyrrhenian back-arc basin; K indicates the magnetic susceptibility. For location, see Figure 1. (M. Grasso)

Open issues and discussion

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

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

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

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

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

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

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

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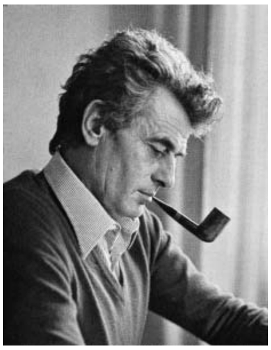
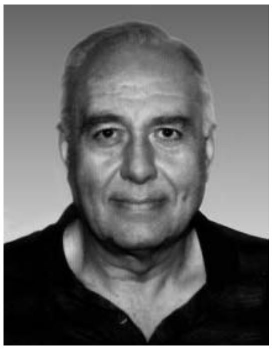
***Piero Elter**, born in 1927, retired Professor of Geology at the University of Pisa. He received the Degree in Geology at University of Geneva in 1954. Piero Elter was Director of the Centro di Geologia Strutturale e Dinamica dell'Appennino of the CNR (Italian National Council of Research) from 1972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

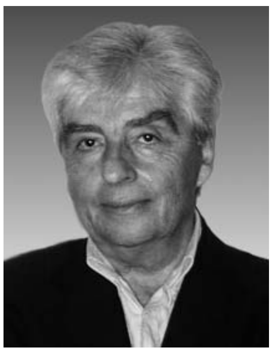


***Maurizio Parotto** is teaching Introduction to Geodynamics and Historical Geology at University "Roma Tre" (Italy). His research concentrates on stratigraphy and structural setting of central Apennine; at present he is Director of sub-project CROP 11 (CROsta Profonda, Deep Crust), a part of a CNR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



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

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



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

Topography, exhumation, and drainage network evolution of the Apennines

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

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

Introduction

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

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

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

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

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

Tectonic setting

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

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

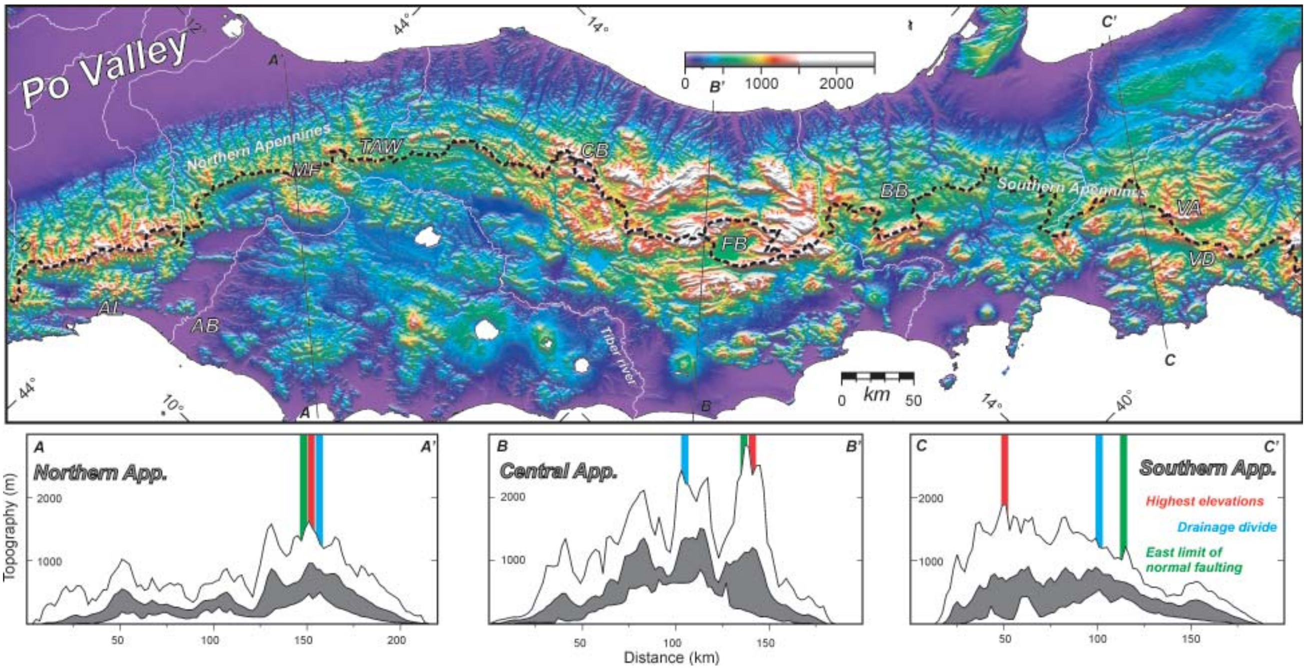


***Figure 1** Shaded topography of the Italian peninsula (Oblique Mercator Projection) with swath profiles across northern, central and southern Apennines. Upper, median, and lower lines in the profiles correspond to maximum, average, and minimum elevation perpendicular to a 40-km swath.*
Legend: AL,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

Thermochronological data and exhumation history

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

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

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

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

Evolution of the drainage network and the "watershed" basins

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

Detailed geological and geophysical investigations in the Fucino basin document (Figure 2) the persistent Plio-Quaternary lacustrine environments and the asymmetrical geometry of the basin fill thickening toward the master NW trending fault systems that ruptured during the 1915 Ms = 6.9 Fucino earthquake. The geometry of the basin fill shows that the fault segment which ruptured during that earthquake may have accumulated a vertical offset larger than 1000 m by repeated 1915-like seismic events. The remarkable correspondence between the distribution of the 1915 coseismic vertical deformation and the extent of internal drainage provides a striking example of the struggle of intramontane basin faults to preserve internal drainage in the "watershed" basins against integration by regional drainage network by repeated seismic events that "dam" externally flowing rivers or by providing local subsidence not compensated by basin aggradation.

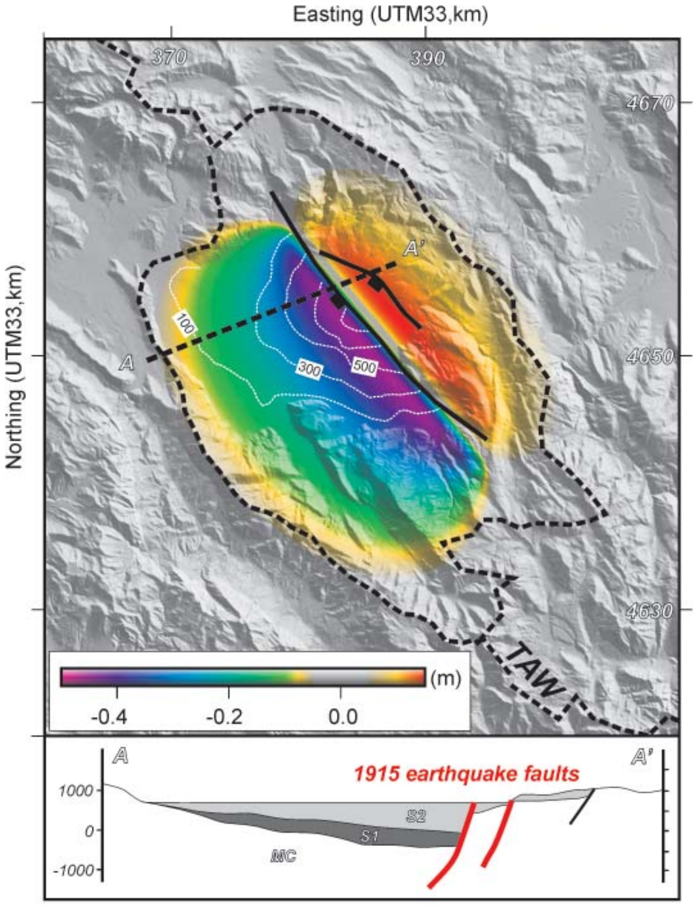


***Figure 2** Fucino basin schematic map. The color scale corresponds to the vertical coseismic deformation of the 1915 Ms 6.9 earthquake (fault parameters from Ward and Valensise, 1989). Main Quaternary basin-bounding faults are shown as black lines. Thin dashed white lines indicate isochron contours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

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

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

i) They are both far from the base levels of the streams that are trying to capture them. The wave of regressive erosion triggered by the base level fall has effected these internal regions later than those basins closer to the coasts, such as the Tiber, Sulmona, and Campo Imperatore basins. Once captured, the increased discharge through these well-integrated basins makes it likely that fluvial incision rates at their outlets will always dominate over local vertical motions related to faulting.

ii) Normal faulting is most active along the crest of the topographic bulge. Only in this location can faulting efficiently delay the capture of closed basins by causing subsidence that preserves internal drainage.

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

Discussion

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

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

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

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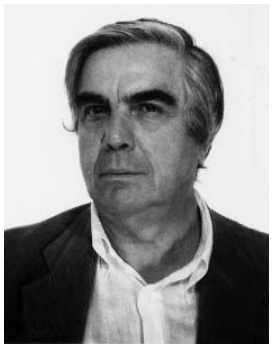
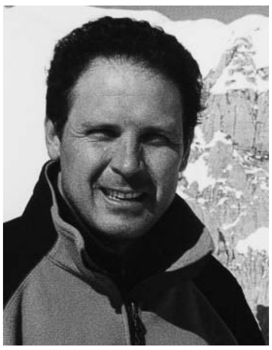
Carlo Bartolini*, professor of Physical Geography and Geomorphology at the University of Florence, Italy, has worked on marine geology and coastal geomorphology, neotectonics and morphotectonics. His present research is focused on exhumation/denudation rates over the Northern Apennine and on the impact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

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



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



by Renzo Sartori

The Tyrrhenian back-arc basin and subduction of the Ionian lithosphere

Dipartimento di Scienze della Terra e Geologico-Ambientali, University of Bologna, Via Zamboni 67, I-40127, Bologna, Italy.
E-mail: contact *capozzi@geomin.unibo.it*

A deep, narrow, and distorted Benioff zone, plunging from the Ionian Sea towards the southern Tyrrhenian basin, is the remnant of a long and eastward migrating subduction of eastern Mediterranean lithosphere. From Oligocene to Recent, subduction generated the Western Mediterranean and the Tyrrhenian back-arc basins, as well as an accretionary wedge constituting the Southern Apenninic Arc.

In the Tyrrhenian Sea, stretching started in late Miocene and eventually produced two small oceanic areas: the Vavilov Plain during Pliocene (in the central sector) and the Marsili Plain during Quaternary (in the southeastern sector). They are separated by a thicker crustal sector, called the Issel Bridge. Back-arc extension was rapid and discontinuous, and affected a land locked area where continental elements of various sizes occurred. Discontinuities in extension were mirrored by changes in nature of the lithosphere scraped off to form the Southern Apenninic Arc. Part of the tectonic units of the southern Apennines, accreted into the wedge from late Miocene to Pliocene, had originally been laid down on thinned continental lithosphere, which should constitute the deep portion of the present slab. After Pliocene, only Ionian oceanic lithosphere was subducted, because the large buoyancy of the wide and not thinned continental lithosphere of Apulia and Africa (Sicily) preserved these elements from roll back of subduction. After Pliocene, the passively retreating oceanic slab had to adjust and distort according to the geometry of these continental elements.

The late onset of arc volcanism in respect to the duration of extension in the Tyrrhenian-Ionian system may find an explanation considering an initial stage of subduction of thinned continental lithosphere. The strong Pleistocene vertical movements that occurred in the whole southeastern system (subsidence in the back-arc basin and uplift in the orogenic arc) may instead be related to the distortion of the oceanic slab.

Introduction

The Tyrrhenian and the Aegean Seas are back-arc basins connected to two Benioff zones (Calabrian and Aegean) indicating subduction of the Eastern Mediterranean lithosphere, an old, largely oceanic area. The Aegean trench and arc roughly trend E-W and here subduction reflects the N-S convergence between the African and European plates, as well as coaxial extrusion of Anatolia (Le Pichon and Angelier, 1979).

The Calabrian subduction system (Figure 1) is instead almost normal to the Aegean one and appears independent from the motion of the large plates. Here, subduction has been imputed, first by Ritsema 1979 (passive subduction) and subsequently by Malinverno and Ryan, 1986 (roll-back of subduction) to the slab pull of a retreating, old and dense oceanic lithosphere.

In this frame, the marine areas to the W of Italy represent a set of eastward younging back-arc basins. The Western Mediterranean basin formed from Oligocene to Middle Miocene at the rear of the Corsica-Sardinia continental block, which acted in the meantime as anti-clockwise rotating volcanic arc. Since late Miocene, back-arc extension resumed to the E of Sardinia, and the Tyrrhenian basin started forming.

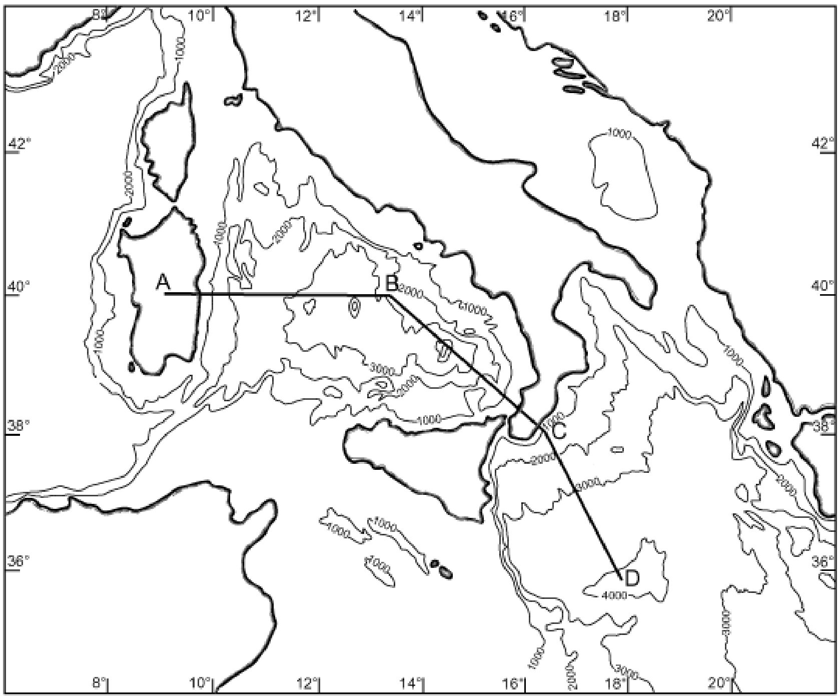


Figure 1 Location map of geological section across the Tyrrhenian to Ionian Seas (central Mediterranean) shown in Figure 3.

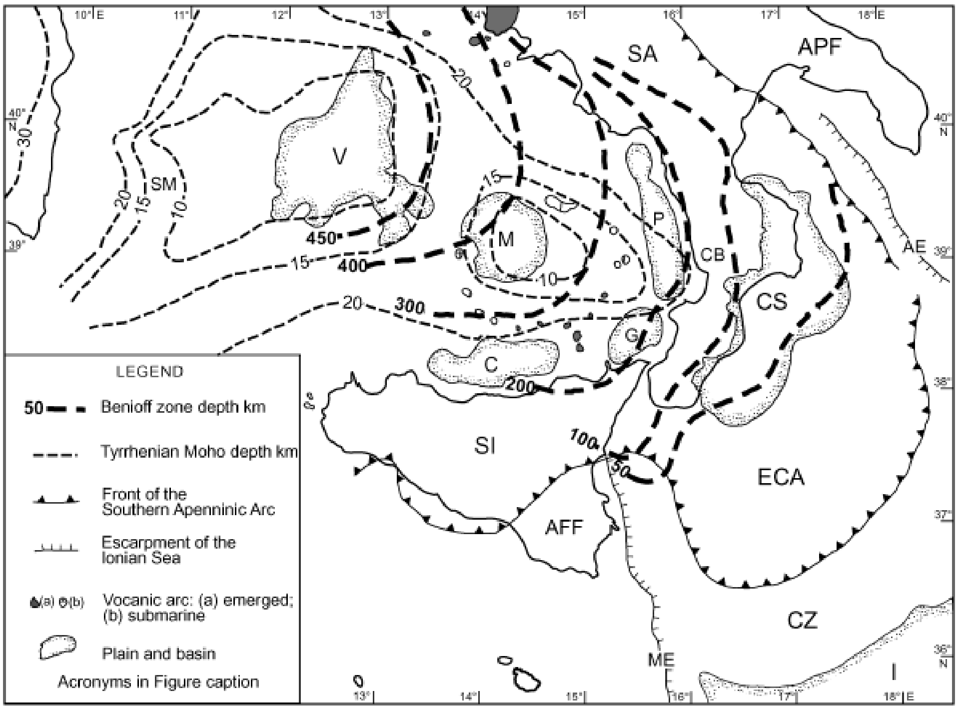


Figure 2 Moho depth and Benioff zone contours in the Tyrrhenian to Ionian area. Legend: AE, Apulian escarpment; AFF, African foreland; APF, Apulian foreland; C, Cefalù basin; CS, Crotone-Spartivento basin; CZ, Cobblestone zone; ECA, External Calabria Arc; G, Gioia basin; M, Marsili basin; ME, Malta escarpment; P, Paola basin; SA, Southern Apennines; SI, Sicily; SM, Sardinia margin; V, Vavilov basin.

Running with continuity from peninsular Italy to Sicily, the Southern Apenninic Orogenic Arc developed and migrated eastwards accompanying extension in the back-arc regions (see Vai and Martini, 2001, for an updated account on the Apennines). In respect to subduction processes, the arc behaved as upper plate or as off-scraped tectonic wedge. The Southern Arc has migrated towards the Apulian (Southern Apennines) and African (Sicily) continental forelands; but also towards the deep marine areas off Calabria, where off-scraping of the Ionian lithosphere (“oceanic foreland?”) is occurring (Figure 2).

This paper describes the present configuration of the Tyrrhenian-Ionian system; it tries to interpret, also on the base of data recently achieved in marine and emerged areas, the history of its progressive evolution during the last 10 Ma. It also takes into account some unusual features of the system, as a late development of arc volcanism and the onset of large vertical movements coupled to extension and shortening.

The Ionian lithosphere and the Calabrian Benioff zone

Geometry and characters of subduction

The Ionian Sea, NW portion of the Eastern Mediterranean, reaches some 4,100 m depth in the Messina Abyssal Plain and is bordered by two steep scarps 3 km high (Apulian and Malta Escarpments), where mostly Mesozoic carbonate sequences crop out (Bigi et al., 1991).

Nature and age of the crust in the deep Ionian basin are still a matter of debate. Refraction data (DeVoogd et al., 1992, with previous references) indicate either an oceanic or a thinned continental crust, depending on the location of seismic experiments. In addition, magnetic signature is poor and the basement is covered by several km of sedimentary rocks. Across the borders of the basin, a crust transitional to normal continental (Africa and Apulia) occurs, but break-up unconformities do not crop out and, in addition, the present steep scarps were strongly reactivated during Neogene, so that they do not contain information as to the age of the oceanic crust in the Ionian basin. The proposed ages for this element span therefore from Permian to Tertiary (with cluster around late Mesozoic), depending on the various indirect criteria adopted (Catalano et al., 2000, with previous references).

At any rate, deep reflection seismic lines clearly image a flexing of the Ionian lithosphere beneath Calabria (Cernobori et al., 1996, with previous references) to join a seismogenic slab that extends down to some 500 km beneath the SE Tyrrhenian basin (Anderson and Jackson, 1987). This NW plunging Benioff zone is distorted, long (700 km) and narrow (less than 250 km), abruptly disappearing laterally beneath the Apennines and Sicily (Ritsema, 1979; Gasparini et al., 1982). The depth reached by the Benioff zone (and the presence of a volcanic arc, see later) indicates that oceanic lithosphere has been subducted in this system.

The trench-arc region

A deep, continuous trench is not observed in the Ionian basin, as is the case for other converging margins carrying thick sediments. Where the Ionian basement starts plunging NW-ward, a gentle and irregular slope brings water depth from more than 4,000 m to less than 2,000 m. The external portion of the slope (cobblestone zone in Figure 3) displays surficial deformations probably triggered by the presence of the Messinian evaporites. The major portion of the slope, named External Calabria Arc, hosts thrust systems, olistostromes, and possibly mud volcanoes, recalling an accretionary wedge (Rossi and Sartori, 1981; Finetti, 1982). Between the upper slope and Calabria, the wide Crotone-Spartivento fore-arc basin occurs. This is partly emerged, and its sedimentary sequences start with middle or upper Miocene (Rossi and Sartori, 1981).

Calabria is a sort of terrane largely made of Palaeozoic crystalline-metamorphic units assembled from Cretaceous till middle Miocene. This arc portion was bent and extended since Tortonian times, and experienced strong uplift (up to more than 1 mm/y) during Pleistocene. Inner to Calabria, but still fore of the volcanic arc, the Paola, Gioia, and Cefalù basins developed via extensional listric faults during or after Tortonian times, as parts of the eastern Tyrrhenian rifted margin (see later). The Paola basin shows a strong Pliocene-Quaternary subsidence, with thick sediments in part very gently folded. In the Gioia basin subsidence was largely Tortonian, while in the Cefalù basin, offshore Sicily, post-Tortonian stretching interplayed with wrench tectonics (Barone et al., 1982).

The Southern Apenninic Arc

Since Oligocene-Miocene, the continuous orogenic system forming the Southern Apenninic Arc was shortened while migrating towards the Apulian (Southern Apennines) and African (Sicilian Maghrebids) forelands, as well as towards the Ionian basin (Calabria and External Calabria Arc). The Tortonian to Recent part of this evolution is coeval with the development of the Tyrrhenian back-arc basin. Accepting that the units of the arc represent a tectonic wedge generated by off-scraping of the subducting slab, the pattern of arc shortening and migration should reflect the progressive deformation of the system. In the Southern Apennines, shortening and migration rates up to 6–8 cm/y from Tortonian to Pliocene have been calculated, with Pleistocene shortening turned to sinistral strike-slip deformation (Patacca et al., 1992, 1993). Again with Pleistocene, large portions of the Southern Arc (including the Apulian foreland and Sicily) were affected by strong uplift. These observations call for a marked change in the deformation regime of the arc across the Pliocene-Pleistocene boundary.

The volcanic arc

The Eolian Islands are the volcanic arc of the converging system (Barberi et al., 1974). It is located in correspondence to a depth of the Benioff plane of about 250–300 km (Figure 3). The islands are part of a wider, largely submarine, ring-shaped complex of volcanic edifices whose activity started less than 1.7–1.5 Ma ago and still goes on in Stromboli Island and in other localities. Coeval arc type volcanism is typical of the whole south-eastern Tyrrhenian plain, including the large Marsili volcano, while manifestations of arc type older than Quaternary are quite rare elsewhere in the Tyrrhenian.

Slightly modified arc-type volcanism (Serri et al., 2001) also occurred since 2 Ma and still is active (Vesuvius, etc.) in the Neaples-Phlegrean area, quite N of the Eolian Islands. This volcanism is located at the abrupt NE closure of the slab, where hypocenters suddenly pass from more than 400 km to crustal depths.

The southern Tyrrhenian back-arc basin

General characters

The northern Tyrrhenian Sea, between Corsica and the Northern Apennines, is shallow, its crust is only moderately thinned, and its present evolution is not related to the Ionian subduction. The southern Tyrrhenian is included between Sardinia and the Southern Apenninic arc. It displays an irregular seafloor, marked by important seamounts, and reaches a depth of about 3,600 m. In two discrete areas, around the Vavilov and Marsili deep plains, MOHO depth is around 10 km; and crustal thickness increases to 30 km and more towards the continental areas encircling the basin.

The southern Tyrrhenian was developed by stretching of an area previously occupied, from W to E, by the Sardinia continental block (Variscan basement), by the Europe-verging Alpine chain (Cretaceous-Paleogene) and by the portion of Apenninic arc formed from Oligocene to middle Miocene (Sartori, 1986). These early formed Apenninic units were generated by deformation of the Apulian and African margins during the development of the Western Mediterranean basin, in front of the rotating Corsica-Sardinia block.

A poly-phase evolution

From Sardinia to the central Tyrrhenian

Timing of rifting has been investigated across the Sardinia margin of the southern Tyrrhenian, during ODP Leg 107 (Kastens, Mascle et al., 1990). Lying north of a prominent lithospheric feature called Orosei Canyon Line (OCL) by Sartori et al. (2001) and occurring at about 40˚N latitude, the upper portion of the margin was rifted from intra-Tortonian to mid-Messinian, while the lower one

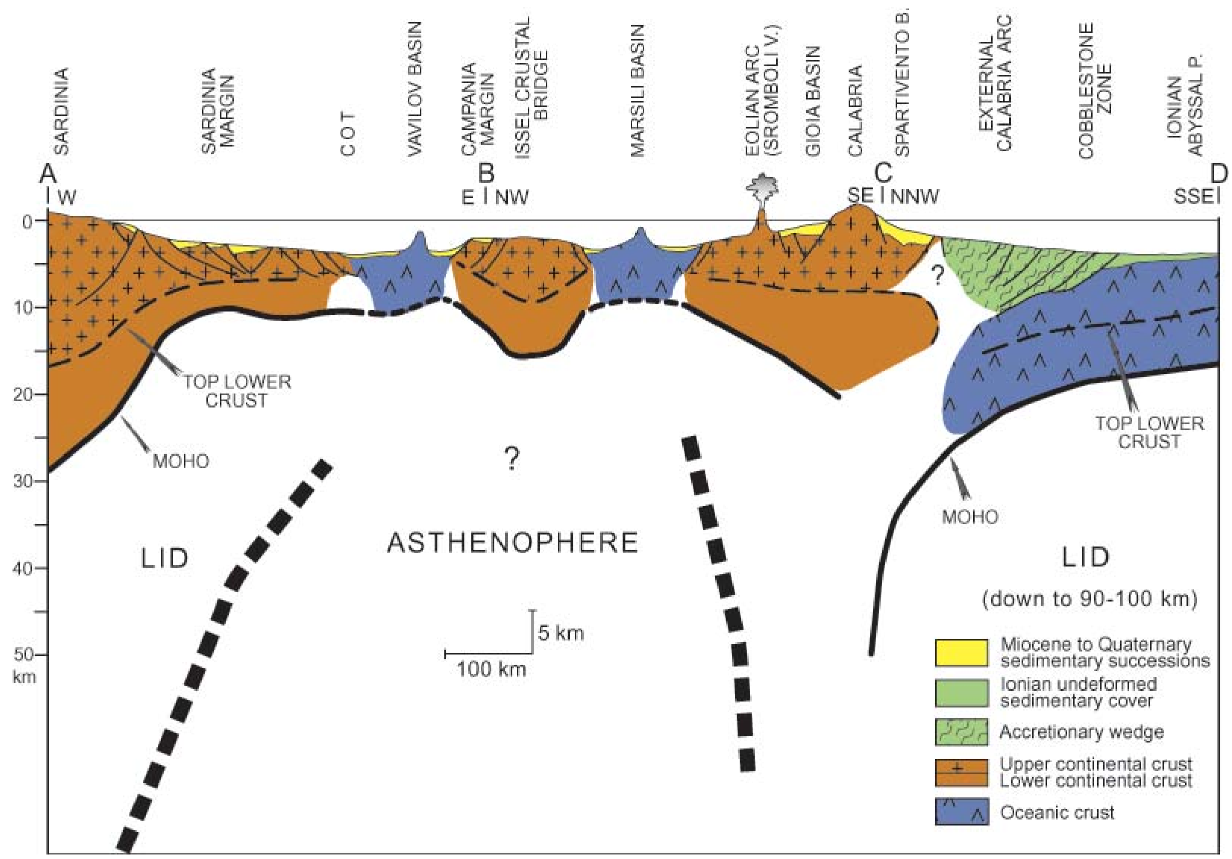


Figure 3 Crustal geological section (vertical exaggeration × 10) and lithospheric/asthenopheric interpretation from Sardinia through the Tyrrhenian Sea, Calabria and the Ionian Sea.

experienced stretching from mid-Messinian to lower Pliocene time. A wide continent-ocean transition (COT) was generated, with MOHO lying at 10 km depth beneath both the middle and lower faulted margin and the Vavilov plain (Sartori et al., 2001 and in press). In the central Tyrrhenian COT, extensional deformation rates were high, bringing crustal thickness from 30 km (or more) to 10 km, in some 5 Ma. Abundant magmatism also accompanied stretching in the central Tyrrhenian. South of the OCL, rifting was continuous from Toronian to intra-Pliocene times and wider sectors of oceanic crust were generated in the southern Vavilov basin (Figure 3).

The areal distribution of oceanic crust (continent-ocean boundary) inside the central Tyrrhenian is not obvious: the COT is wide and linear magnetic anomalies are confuse or absent. In addition, the basin hosts huge volcanic and sub-volcanic bodies not related to spreading, whose seismic signature interplays with that of oceanic crust (Sartori et al., in press). The oceanic crustal domain in the central Tyrrhenian has a marked triangular shape and should be essentially of Pliocene age.

From the central Tyrrhenian to the Southern Apenninic Arc

A thicker (15 km) crustal sector, named Issel bridge, separates the Vavilov and the Marsili plains, where MOHO depth is again less than 10 km. The Marsili basin has a rhombic shape and is located in the SE Tyrrhenian, just above the Ionian Benioff zone. It is included within the Eolian submerged ring of calcalkaline volcanics, with the huge Magnaghi calcalkaline volcano in the center, and should have been stretched essentially during Quaternary. It was also affected by strong (up to 1mm/y) subsidence since its formation (Kastens, Mascle, et al., 1988).

Timing of rifting along the opposed narrow margins of peninsular Italy and of Sicily is not well constrained, although in most segments stretching had to predate Messinian and possibly interplayed with episodes of wrench deformation (see above).

Summarising age and trends of rifting-spreading evolution, a Tortonian to Pliocene episode of back-arc extension, when the Sardinia margin and the Vavilov plain were formed, indicates migration roughly from W to E; a Pleistocene episode, when the Marsili basin was generated, indicates migration from NW to SE (Sartori and Capozzi, 1998).

Discussion and conclusions

The deep Benioff zone beneath Calabria witnesses a long period of slab retreat, characterized by grossly eastward migration, but punctuated, inside the back-arc regions, by interruptions and changes of extensional vectors. Roll-back of the slab begun during Oligocene, when stretching started in the Western Mediterranean back-arc basin. Here spreading stopped in middle Miocene, when rotation and arc volcanism ceased in Sardinia (Vigliotti and Langenheim, 1995). Back-arc extension resumed with Tortonian, jumping to the East of Sardinia. The Tyrrhenian basin then formed, with multistage rifting and with the eventual generation of two discrete oceanic areas, extended under different migration vectors.

Episodicity in back-arc development is observed in intra-oceanic settings, as in the Western Pacific, due to changes in the configuration of the retreating slab with time. In the Mediterranean region, all land-locked, it is feasible that interruptions and changes in back-arc extension were produced by the interference of the retreating oceanic slab with intervening continental lithosphere.

In the Southern Apenninic tectonic wedge, palinspastic analyses point out that a carbonate platform was offscraped and included into the chain from late Miocene to Pliocene. These tectonic units are made up by Triassic to Miocene shallow water sequences, detached from a continental basement, since the platform was part of the inner Apulian continental margin of Tethys which experienced rifting during Mesozoic. External to this platform, but internal to a further carbonate platform making up the Apulian foreland of the chain, the passive margin of Tethys also included deep-water domains hosting at places basic volcanics. These basins, floored by thinned continental or oceanic crust, were included into the eastward migrating chain essentially during Pliocene. With Pleistocene, shortening ceased in the Southern Apennines and the tectonic regime turned to sinistral NW-SE strike-slip, accompanied by strong uplift of the whole orogenic system.

In the Tyrrhenian basin, after a Tortonian to middle Pliocene W to E directed stretching, which generated the Sardinia margin and the Vavilov plain, by the end of Pliocene extension rapidly turned to NW-SE and was limited to the southeastern Marsili plain.

Assuming that back-arc deformation and migration of the orogenic arc (accretionary wedge) were connected, we are forced to conclude that from Tortonian to Pliocene the sectors of thinned continental lithosphere, underlying the inner Apenninic carbonate platform, experienced subduction. With Pliocene-Pleistocene, it was instead Ionian-type oceanic lithosphere, branching possibly northwards in the deep-water areas located between the internal and the Apulian carbonate platforms, that was involved in subduction.

The Pliocene-Pleistocene changes in Tyrrhenian back-arc extension and in deformation of the Southern Apennines may reflect the resistance opposed to slab retreat by the wide Apulian and Sicilian (African) forelands. These elements were not thinned by the Mesozoic rifting and are characterized by normal continental lithosphere. During its eastward migration, passively retreating oceanic slab had to adjust and deform, in relation to the occurrence of these large and buoyant continental sectors. Post-Pliocene migration was therefore eventually driven only towards the narrow (250 km) corridor represented by the present Ionian Sea, largely oceanic in nature, and separating the Apulian from the African sectors.

This reconstruction implies that the present Benioff plain, although rather continuous at depth, has not a homogeneous oceanic nature, since its deep portion should consist of the thinned continental lithosphere that floored the inner carbonate platform of the Southern Apennines.

This hypothesis could explain the late onset of arc volcanism (2–1.5 Ma) in respect to inception of back-arc extension in the Tyrrhenian basin (8–9 Ma), assuming that no foundered remnant arcs exist inside the system. Arc volcanism could not develop in the central Tyrrhenian area, produced by back-arc extension between Tortonian and intra-Pliocene, because in the meantime the thinned continental lithosphere underlying the inner Apenninic platform was subducted. This portion of the Tyrrhenian hosts abundant Pliocene-Pleistocene volcanism, but this is essentially of Ocean Island Basalt-type, with no influence from subduction (Serri et al., 2001; Sartori, in press).

Arc volcanism can only be triggered when oceanic lithosphere is subducting, and after it reaches at least some 100–150 km depth. With Pliocene, subduction affected the Ionian basin and the deep-water sectors of the Southern Apennines. Arc volcanism resumed in the system only when this oceanic lithosphere reached the depth necessary for magma generation. Due to the high velocity of slab retreat, this critical depth was reached towards the beginning of Pleistocene. At this time, back-arc extension was controlled by the not thinned continental sectors (present chain forelands), and migrated NW-SE into the Marsili basin. A calculation can be made taking an average dip of the Benioff plain of about 60–70˚, and assuming that the age of inception of arc volcanism in the Eolian area (2–1.5 Ma) indicates the time when the oceanic portion of the subducting slab had reached 100–150 km depth. Since the Eolian volcanism is still active above a slab some 300 km deep, this would indicate a horizontal component of slab retreat (at constant dip) of some 100–120 km in 2 Ma, which is roughly the width of the Marsili basin. The derived NW-SE Pleistocene extensional rate of 5–6 cm/y is well in line with deformation-migration rates calculated in the Southern Apennines (Patacca et al., 1992, 1993).

The present distortion of the slab should result from its different trends of retreat: W to E from Tortonian to Pliocene, and NW to SE afterwards (see also Ritsema, 1979), implying that the geometry of the subducting slab can adjust rapidly (in less than 2 Ma) to changes in the boundary conditions dictated by the occurrence of large, normal continental lithosphere. This is not surprising considering the high deformation rates shown by this land-locked system. A working hypothesis is that the large Pleistocene vertical movements, observed in the Southern Apenninic Arc (uplift) and in the Marsili plain (subsidence) are related to slab distortion. The subsidence in the Marsili plain (up to more than 1 mm/y) is much higher than expected by thermal cooling of the oceanic crust and is the same order of magnitude of the uplift in the Southern Apenninic Arc. Observing that the arc lies on the convex part of the distorted slab while the Marsili plain corresponds to the concave one (see hypocentral isobaths in Figure 2), it is tempting to relate such opposed vertical movements to Pleistocene slab distortion.

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*The late **Renzo Sartori** was Professor of Marine Geology at the Bologna University (Italy), where he got a Laurea in Geological Sciences in 1970. He participated several cruises in the Mediterranean Sea and in the Atlantic and Pacific Oceans, including DSDP Leg 59 and ODP Leg 107. In ODP, he served as member of EXCOM and SSP; he also chaired the European Consortium (ECOD) Management Committee for ODP (EMCO). He was author or co-author of more than 100 research papers and geological maps, mainly related to the various sub-basins of the Mediterranean and to Equatorial Pacific and Eastern Atlantic areas.*

Renzo Sartori passed away on January 5, 2003, in Bologna. With his passing the geological fraternity has lost an excellent geologist and a fine gentleman.

— The Editor of this Special Issue

by Angelo Peccerillo

Plio-Quaternary magmatism in Italy

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

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

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

Introduction

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

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

Petrological characteristics of Plio-Quaternary magmatism in Italy

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

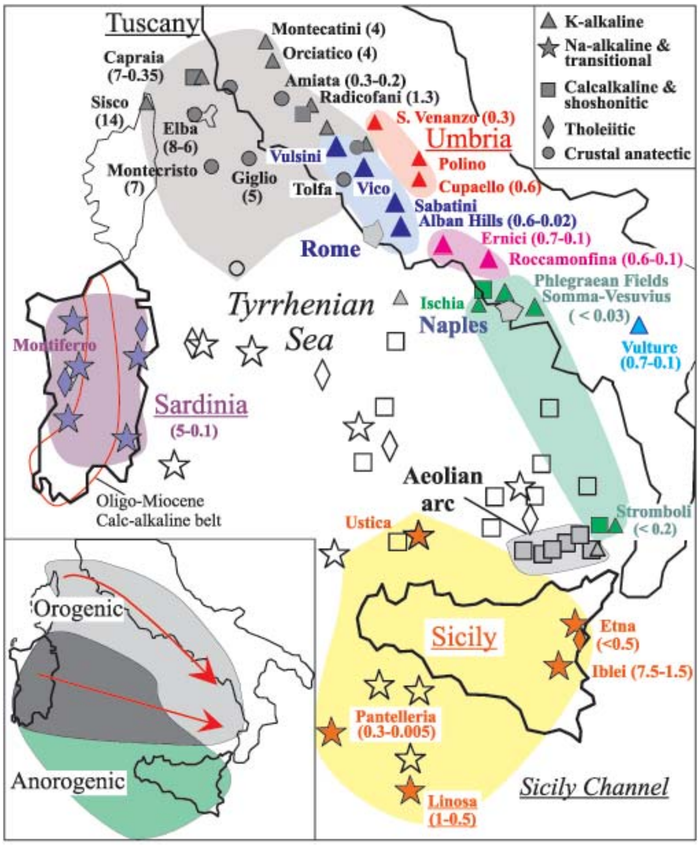


***Figure 1** Distribution of Recent magmatism in Italy. Open symbols indicate seamounts. Ages (in Ma) are given in parentheses. Different colours denote various magmatic provinces. Inset: schematic distribution of orogenic and anorogenic volcanism: red arrows indicate migration of orogenic magmatism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

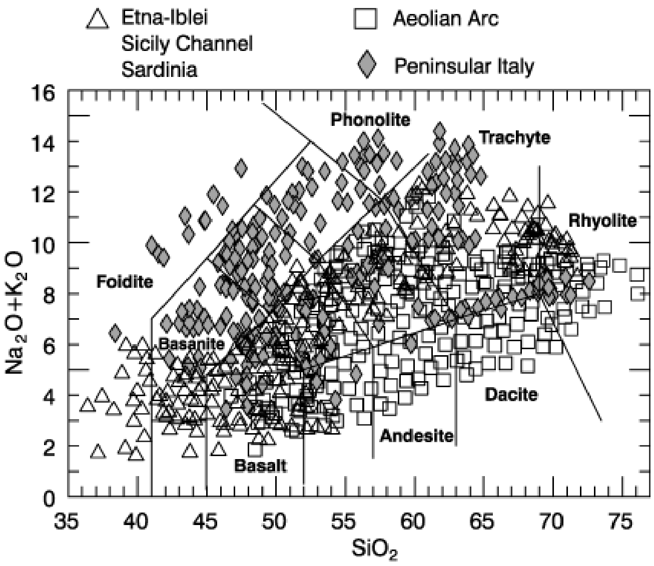


Figure 2 Total alkali vs. silica classification diagram for Italian Plio-Quaternary magmatic rocks. For source of data see Peccerillo (2002).

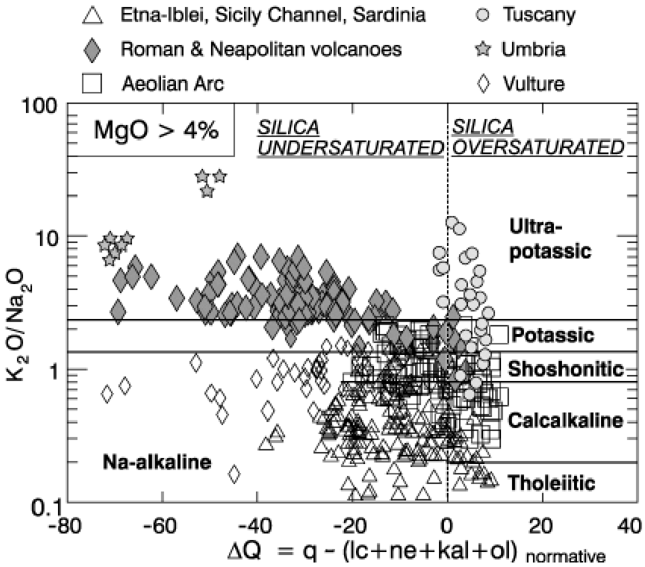


Figure 3 ΔQ vs. K_2O/Na_2O classification diagram for Plio-Quaternary mafic (MgO > 4%) volcanic rocks from Italy. ΔQ is the algebraic sum of normative quartz (q), minus leucite (lc), nepheline (ne), kalsilite (kal) and olivine (ol). Silica oversaturated rocks have $\Delta Q > 0$, whereas silica undersaturated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

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

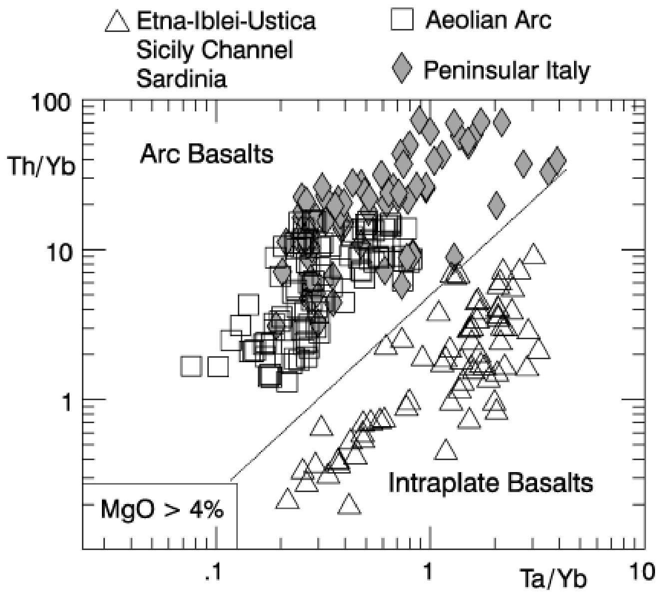
A large proportion of Italian Plio-Quaternary volcanic rocks have high-silica, low-MgO compositions. However, mafic rocks (MgO>3–4 wt%) deserve particular attention, since they are the closest relatives of primary mantle-derived magmas that were parental to erupted lavas, and can furnish the maximum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

Regional distribution of magma types

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

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

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



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

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

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

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

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

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

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

Petrogenesis

Low-pressure magma evolution

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

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

Genesis of mafic magmas

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

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

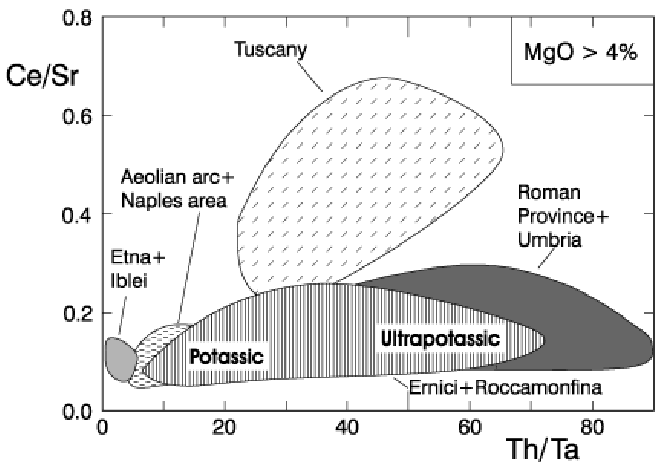


Figure 5 Variation of trace element ratios in Plio-Quaternary Italian mafic rocks. Note strong regional variation.

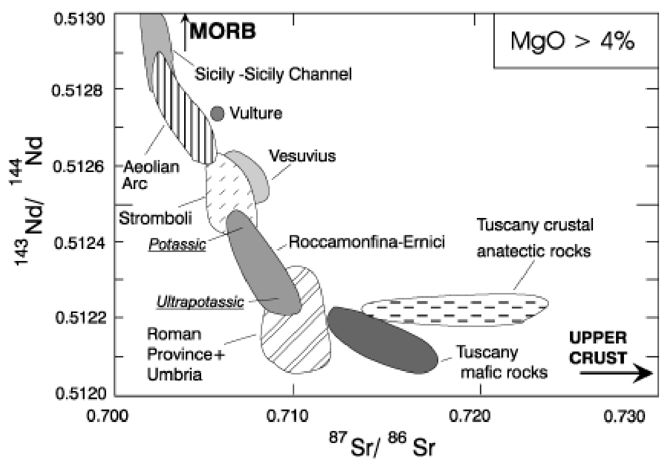


Figure 6 Sr vs. Nd isotope diagram for Plio-Quaternary mafic volcanic rocks from Italy. Note strong regional variation.

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

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

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

Geodynamic significance

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

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

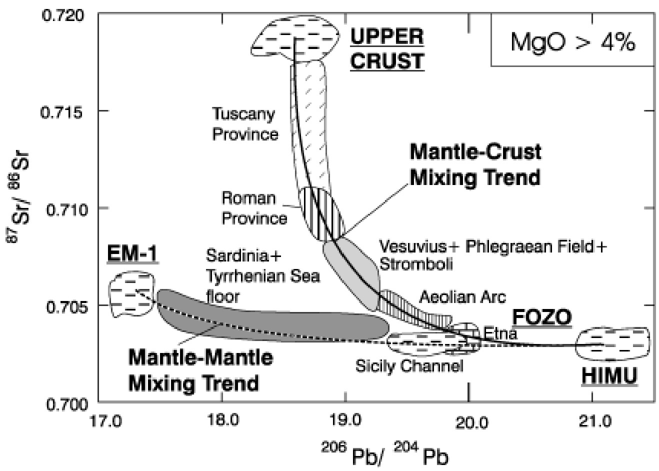


***Figure 7** $^{87}Sr/^{86}Sr$ vs. $^{206}Pb/^{204}Pb$ variations of Plio-Quaternary Italian mafic rocks. Central Italy orogenic magmatism falls along a mantle-crust mixing trend involving HIMU-FOZO and Upper Crust. Sicily, Sicily Channel, Sardinia and some Tyrrhenian Sea seamounts (anorogenic magmatism)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

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

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

Conclusions

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

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

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

by Roberto Santacroce[1], Renato Cristofolini[2], Luigi La Volpe[3], Giovanni Orsi[4], and Mauro Rosi[1]

Italian active volcanoes

1 Dipartimento di Scienze della Terra, Università di Pisa, Italy.
2 Dipartimento di Scienze della Terra, Università di Catania, Italy.
3 Dipartimento Geomineralogico, Università di Bari, Italy.
4 Osservatorio Vesuviano, Istituto Nazionale di Geofisica e Vulcanologia, Napoli, Italy.

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 inter-plate 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).

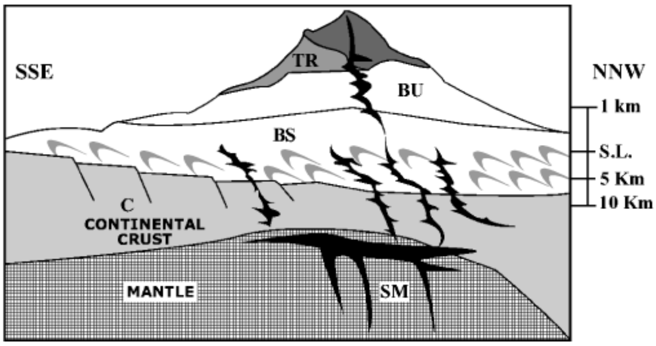

***Figure 1** Sketch NNW-SSE section of Mount Etna. Note the different scale of elevations above and below sea level (lm). MB: Mongibello Unit; TR: Trifoglietto Unit; UB: Basal volcanic Units (older than 80 ka); BS: sedimentary Basement Successions; SM: main magma reservoirs.*

The 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.

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.

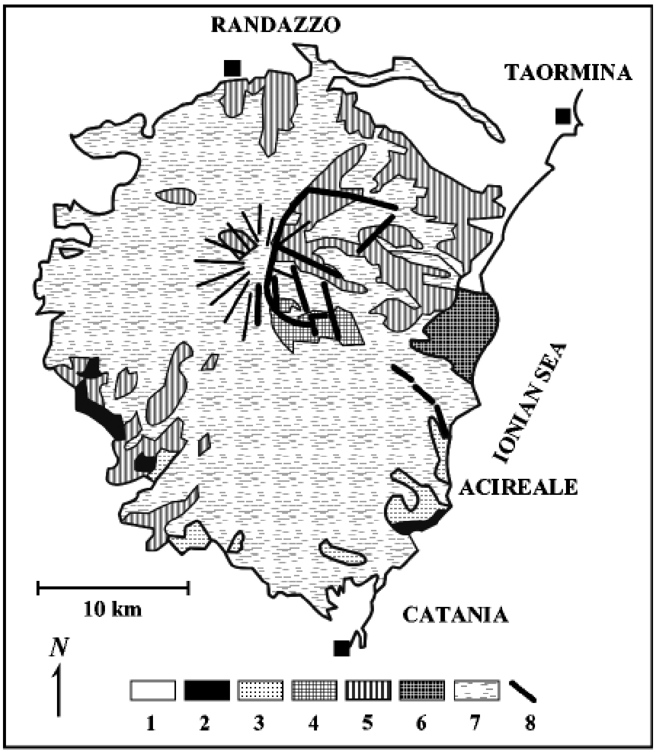

***Figure 2** Sketch map showing the distribution of the main units of the Etnean volcano. 1) Sedimentary basal levels; 2) Tholeiites; 3) Ancient Na-basalts and hawaiites; 4) Trifoglietto Unit (mugearites); 5) Ellittico volcano (Mongibello Unit; hawaiites to trachytes); 6) Detrital alluvial fan from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

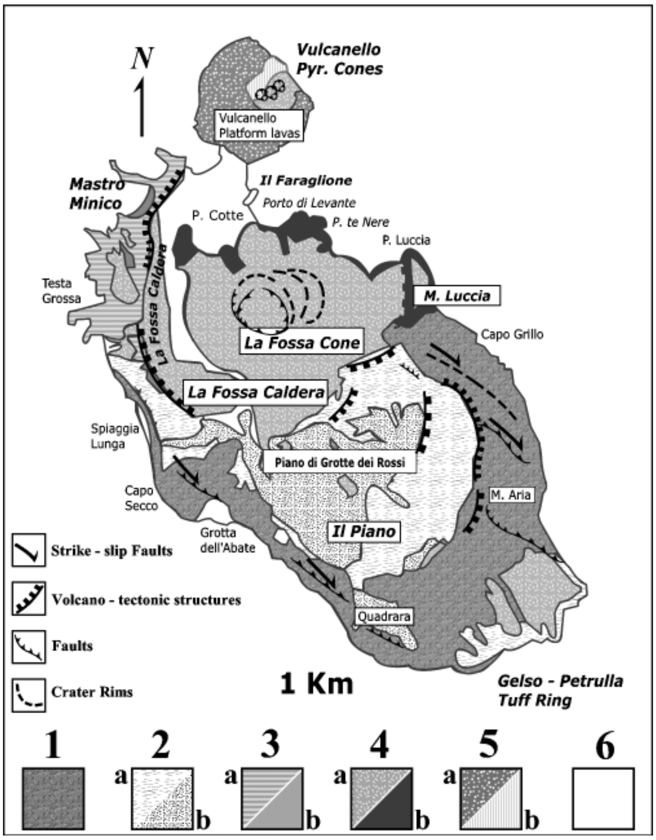

Figure 3 Sketch map of Vulcano Is. 1. Primordial Vulcano; 2. Piano Caldera infilling products; 3. Mastro Minico–Lentia Complex: 3a. rhyolitic lavas, 3b. latitic and trachytic lavas; 4. La Fossa Caldera and La Fossa cone products: 4a. pyroclastics, 4b. lava flows; Vulcanello Cones: 5a. leucite-bearing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

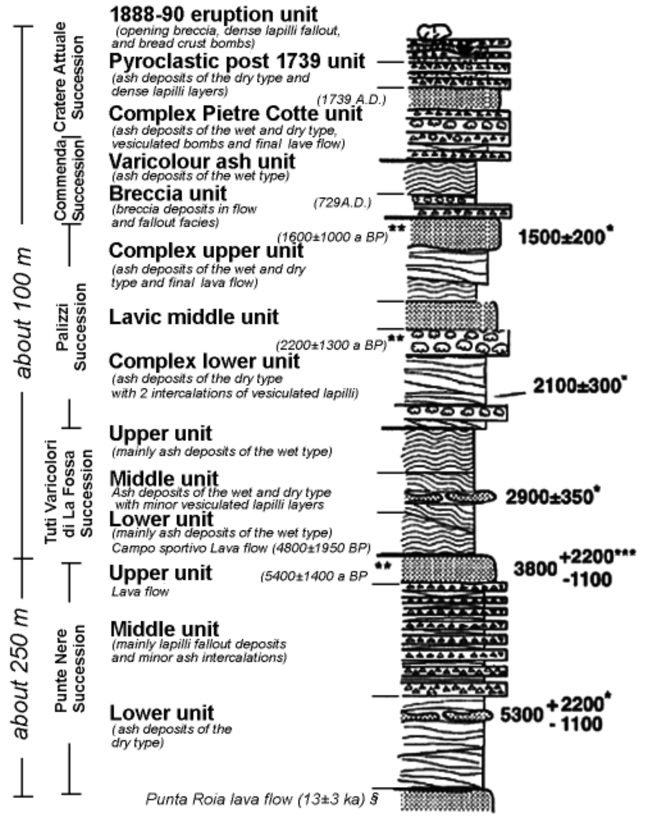

Figure 4 Composite columnar section of La Fossa cone.

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.

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

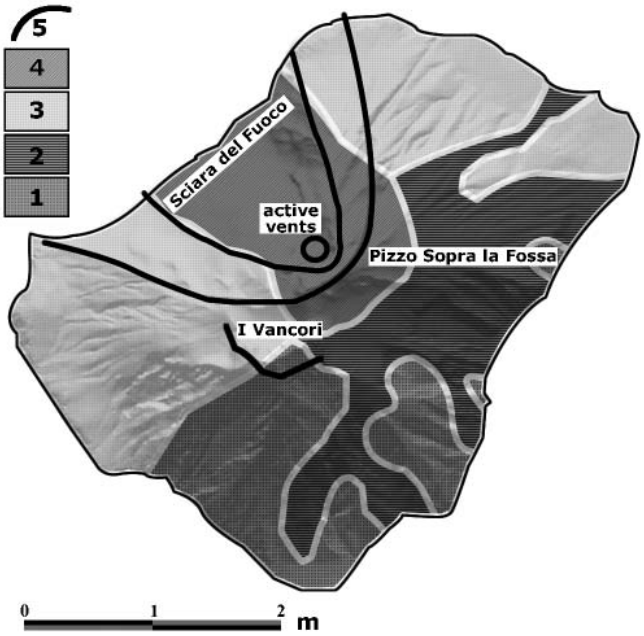

Figure 5 Sketch of Stromboli Island showing the areas covered by products of different periods of activity: 1. Paleo-stromboli (100–35 ka); 2. Vancori (?25–13 ka); 3. Neostromboli (13–5 ka); 4. recent Stromboli (5.0 ka–present); 5. flank collapses

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

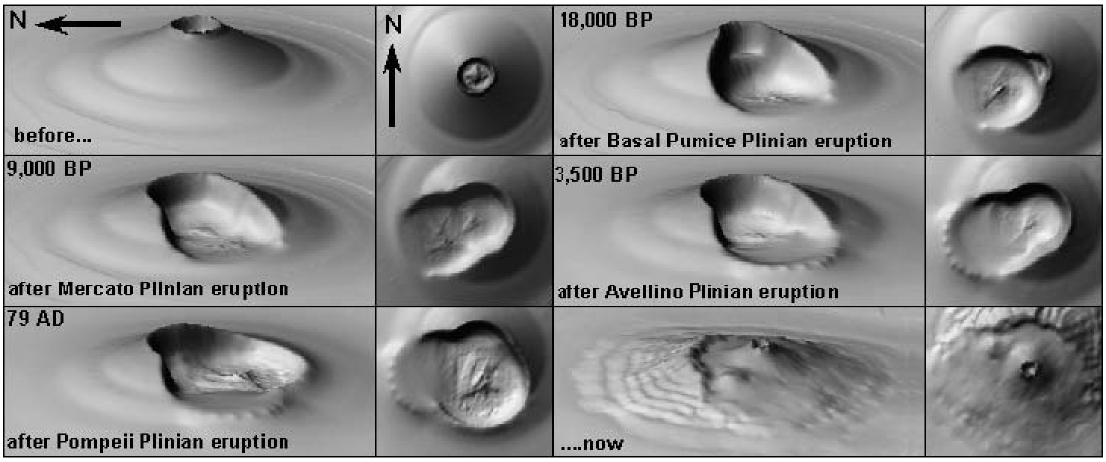

Figure 6 Schematic reconstruction of the morphological evolution of SV caldera.

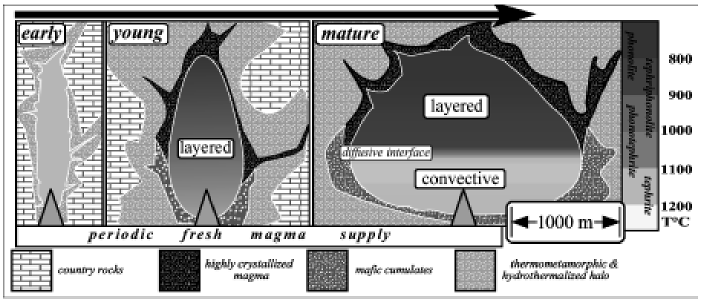

Figure 7 Highly speculative sketch of the evolution of Vesuvio magma chambers growing following the periodic arrival of periodic tephritic inputs (modified after Cioni et al., 1997).

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

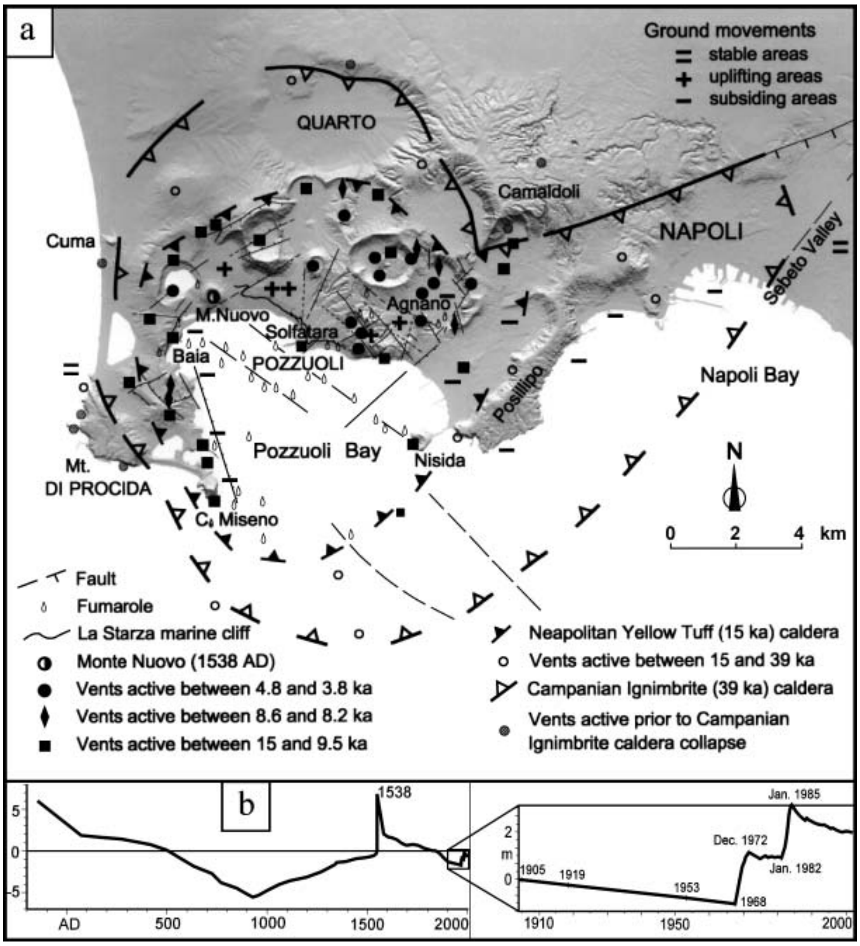

Figure 8 a) Structural map of the Campi Flegrei caldera; b) Vertical ground movements at Serapis Roman market in Pozzuoli.

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.

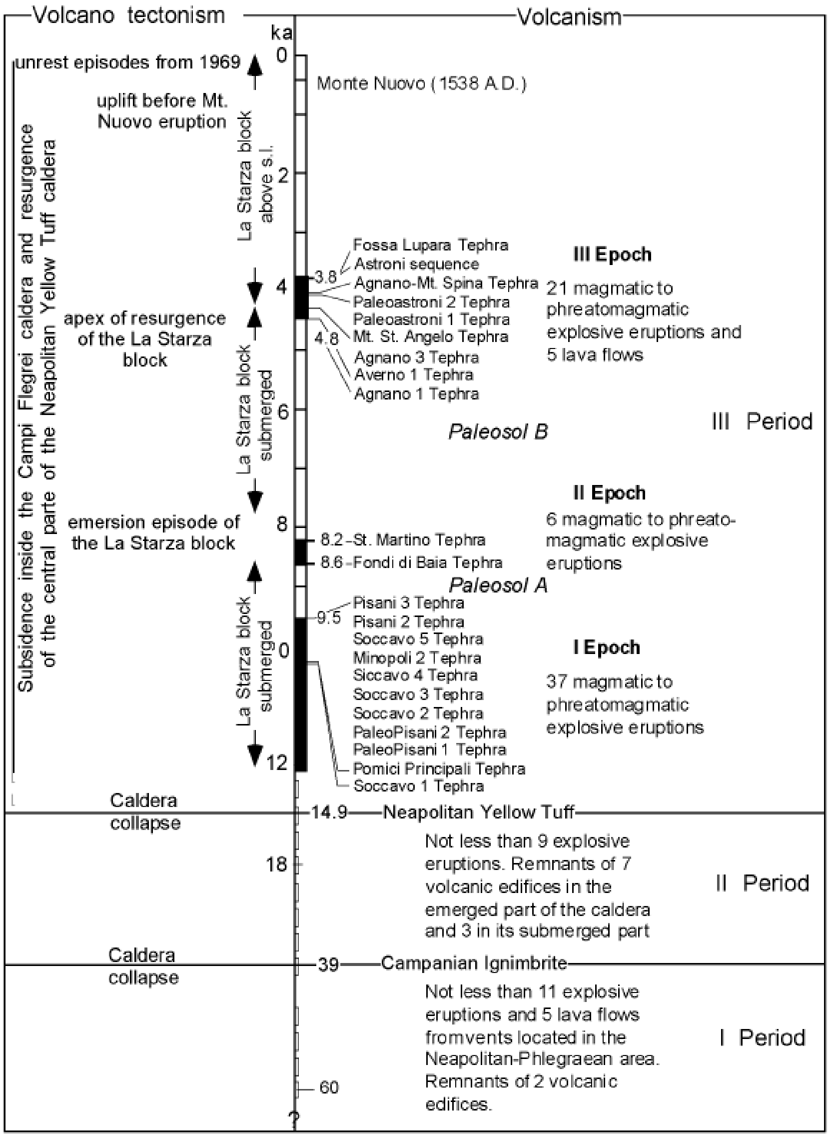

Figure 9 Chronogram of volcanic and deformational history of the Campi Flegrei caldera.

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

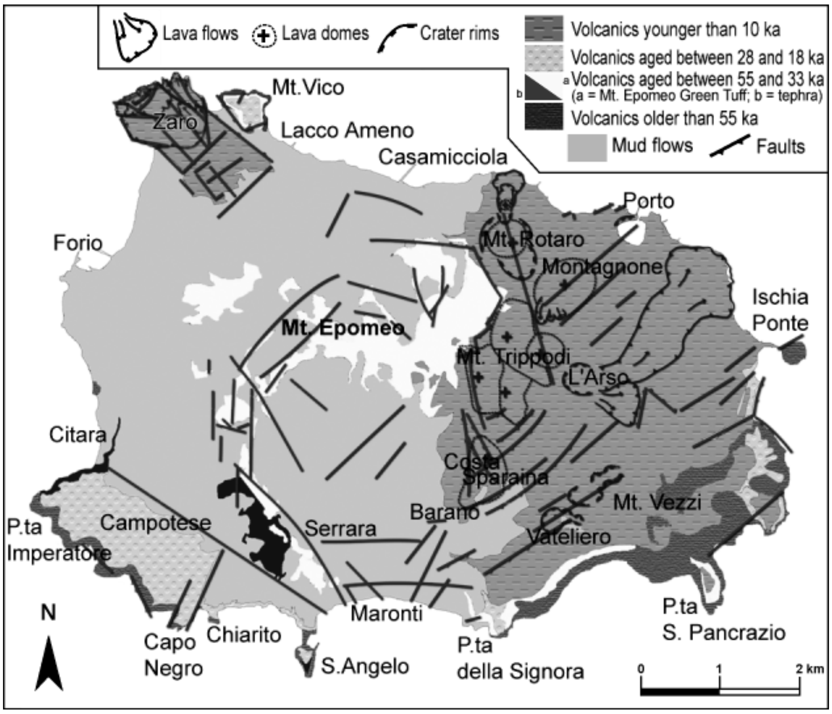

Figure 10 Geological sketch map of Ischia.

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.

Acknowledgments

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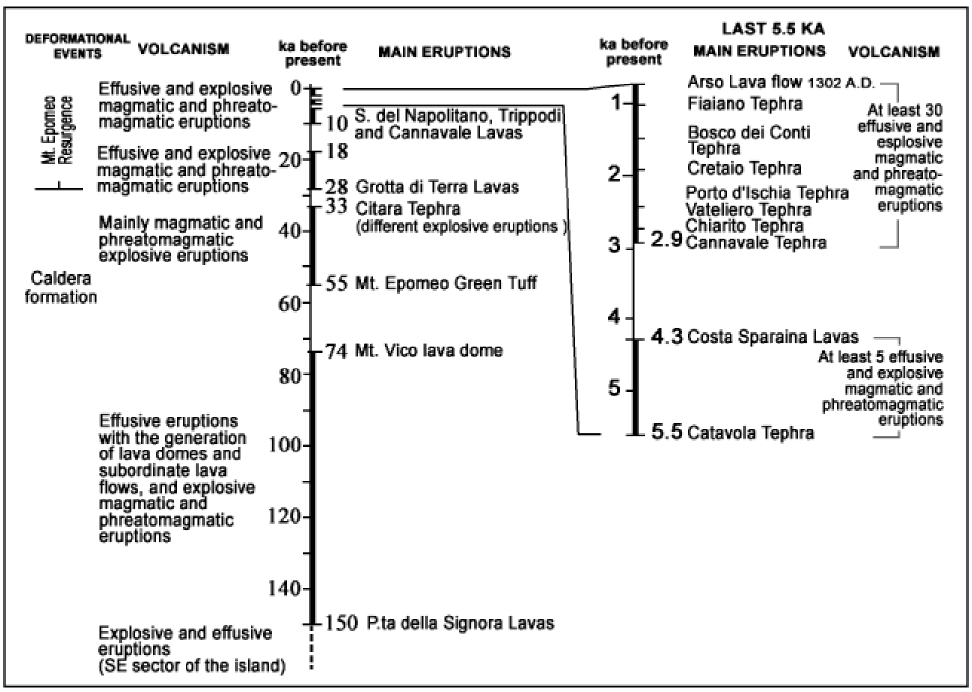

Figure 11 Chronogram of volcanic and deformational history of Ischia. (After de Vita et al., 2003)

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Roberto Santacroce *is full professor of Volcanology in the University of Pisa, formerly associate professor of Applied Petrology, head of the International Institute of Volcanology (IIV) of Catania, head of the Dipartimento di Scienze della Terra in the University of Pisa, chairman of the IAVCEI Commission for the Mitigation of Volcanic Disaster. Its scientific activity covered several aspects of Volcanology, Geochemistry and Igneous Petrology, from the origin of magmas and the relationships between volcanism and tectonics (Italy, East Africa and Red Sea, Arabian peninsula, Central and Southern America), to geothermics (Italy, Ethiopia, Central America), and to volcanic hazard assessment and zonation (Italy).*

by Lisetta Giacomelli[1], Annamaria Perrotta[2], Roberto Scandone[3], and Claudio Scarpati[2]

The eruption of Vesuvius of 79 AD and its impact on human environment in Pompeii

1 Gruppo Nazionale di Vulcanologia, Via Nizza 128, 00193 Roma, Italy.
2 Dipartimento di Geofisica e Vulcanologia, Largo San Marcellino 10, 80138, Napoli, Italy.
3 Dipartimento di Fisica, Università di Roma Tre, 00146, Roma, Italy.

The eruption of Vesuvius of 79 AD caused extensive destructions all over the Campanian area, engulfing the cities of Pompeii, Herculaneum, Oplonti and Stabiae. The eruption followed a long quiescence period and the inhabitants of the area were surprised by the volcanic events. The first part of the eruption was characterized by a widespread dispersal of pumices from a high eruptive column. The second part of the eruption, characterized by pyroclastic flows emplacement, caused the major damages with extensive life losses in most of the towns surrounding the volcano. In Pompeii, the major casualties during the first phase resulted from roof collapses; during the second phase, people were killed either by physical trauma due to the kinetic energy of the flow or by suffucation because of the ash-rich atmosphere.

The sequence of events during the 79 AD eruption of Vesuvius

Vesuvius is one of the most studied volcanoes in the world, because of its long time interval with historic eruptions (2000 years), its easy accessibility, and the first well-documented historic explosive eruption: that of 79 AD. The eruption destroyed Pompeii, Herculaneum, Oplonti and Stabiae and caused the death of Pliny the Elder among many other people.

Before the eruption, earthquakes occurred for some time, but were disregarded by local inhabitants because of their familiarity with the phenomenon. As the younger Pliny testified, “for several days before (the eruption) the earth had been shaken, but this fact did not cause fear because this was a feature commonly observed in Campania”. The effects of these earthquakes are still visible in several buildings in Pompeii, and Villa Regina, where hastened repair works were underway in the days immediately preceeding the eruption.

The main phases of the eruption have been described by Pliny the Younger, who observed the eruption from a distance of more than 25 km, basing also on contemporary testimonies and closer view accounts, especially for what regards the death of the uncle, Pliny the Elder, gone to the rescue of the inhabitants of the area.

The beginning of the eruption is uncertain: the two Plinys observed the cloud at the seventh hour of the day (1 PM). We must presume that the eruption began sometime earlier to allow the arrival, at about the same hour, of a messenger sent from the Vesuvian area.

The eruptive cloud was directly observed from Misenum at a distance of 21 km, so that they could fully appreciate its total extent and behavior. (“*It resembled a pine* [Mediterranean pine] *more than any other tree. Like a very high tree, the cloud went high and expanded in different branches. I believe, because it was first driven by a sudden gust of air then, with its diminution or because of the weight, the cloud expanded laterally, sometimes white, sometimes dark and stained by the sustained sand and ash*”.

During the night of the first day of the eruption, and for most of the morning of the next day, the houses of Misenum were shaken by earthquakes that caused much panic.

In the morning of the second day of eruption, Pliny the Younger observed the development of pyroclastic flows descending down the flanks of Vesuvius and flowing on the sea. “*From the other side, black and horrible clouds, broken by sinuous shapes of flaming winds, were opening with long tongues of fire … . After a little while descended onto the land, opened the sea, covered Capri and prevented the sight of Misenum …*”

The sequence of events described by the Younger Pliny fits well the geologic record of the eruption (Lirer et al., 1973; Sigurdsson et al., 1985).

We can summaryze the temporal evolution of the eruption into major phases which are typical of most large scale explosive eruptions.

1 The first phase, after minor phreatic explosions, is characterized by the development of an high, sustained column where the erupted mixture of juvenile gases and pyroclasts, mixing turbulently with atmospheric air, rises convectively into the stratosphere reaching an estimated maximum height of 32 km.
2 The second phase is characterized by the collapse of the eruptive column with the emplacement of pyroclastic flows and surges which destroyed every settlement within a radius of 10–15 km from the volcano.
3 Collapse of the magma chamber, ingression of water into the feeding system, magma water interaction and final phreato-magmatic activity.
4 Post-eruption remobilization of ashes and pumice by rain water during the following years.

The four phases are identified by their typical deposits (Figure 1).

- The first phase produced a fall deposits consisting of a lower part of well-sorted white pumice and an upper part of gray pumice dispersed to the southeast of the volcano and traced on land to a distance of more than 70 km (Lirer et al., 1973).
- The deposits of the second phase consist of surge deposits made of layers of thin, poorly-sorted ash with cross bedding, and dune structures alternated with pyroclastic flow deposits made by thick and massive layers partly indurated and poorly sorted (Sigurdsson et al., 1985).
- Silty sands beds with abundant accretionary lapilli form the deposit of the third phase. In proximal areas a debris flow deposit consisting of angular lava and carbonate blocks supported by an ash matrix with minor pumices is correlated with this phase (Sigurdsson et al., 1985, Sheridan et al., 1981).

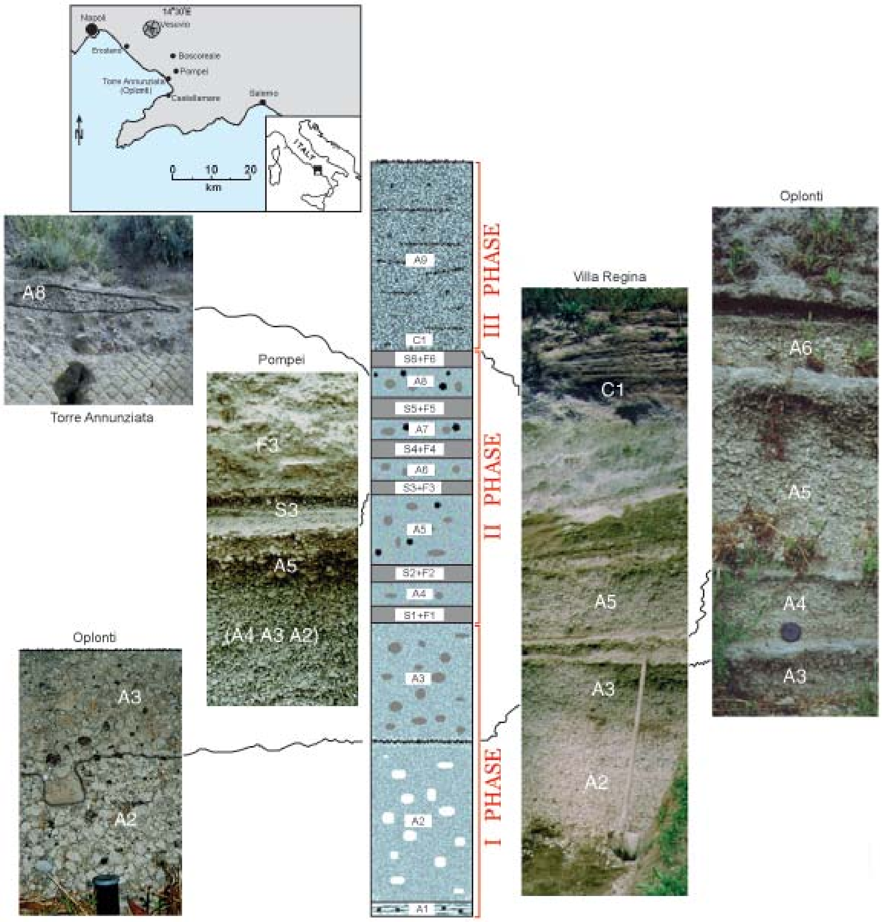

Figure 1 ***General stratigraphic section of the 79 AD eruption deposits based on the reconstruction of Sigurdsson et al. (1985) in the main outcrops surrounding the Vesuvian area.***

- The deposits of the fourth phase are a succession of lahars made up of a conglomerate composed of coarse pebbles with a matrix composed of small pebbles and coarse sand (Lirer et al., 2001).

Carey and Sigurdsson (1987) estimated the height of the eruption column during the development of the Plinian phase basing on the isopleth distribution of maximum diameters of pumice and lithic fragments. They estimated that the eruption column rose from an height of 14 km to 26 km during the emission of white pumice and then to 32 km during the emission of grey pumice immediately before the deposition of pyroclastic flows. The estimates of the column heights permitted the evaluation of the corresponding magma discharge rates.

Scandone and Giacomelli (2001) used the estimates of Carey and Sigurdsson (1987) to evaluate the temporal evolution of the eruption fed by a 7–12 km deep magma chamber. The progressive removal of magma in the course of the eruption caused a slow boiling of magma within the chamber because of decompression. This in turn produced a faster and faster emission rate until the final collapse of the wall surrounding the reservoir. Scandone and Giacomelli (2001) evaluated the duration of the first phase at approximately 22 hours (several hours longer than previously estimated basing on average effusion rates). During this phase there was a steady increase in magma discharge rate. The second and most destructive phase with the massive emplacement of the major pyroclastic flows and surges lasted only about 5–6 hours.

State of the buildings and distribution of victims inside the city of Pompeii

The city of Pompeii was destroyed and many of its inhabitants were killed during the 79 AD eruption. Several authors have reconstructed the succession of products emplaced during the eruption (e.g. Lirer et al., 1973; Sigurdsson et al., 1985; Carey and Sigurdsson, 1987) but the stratigraphic framework used here largely follows that of Luongo et al. (2002a, 2002b) which specifically studied the impact of this eruption on Pompeii. In the following paragraphs we summarise the damage suffered by population and buildings during the first two phases of the eruption and report the stratigraphic height at which were recovered human bodies and crumbled walls. The main sedimentological characteristics of the 79 AD deposit are reported in Figure 2. On the basis of these characteristics the deposit has been subdivided in 8 units named A to H from base upwards; a soil at the top of the sequence (unit I) is also reported.

The state of the buildings all over the city is summarised in the following observations:

a) the amount of destruction is not the same throughout the city, some buildings were more affected than others;
b) the northern (relatively proximal) and southern (relatively distal) sectors in the city were generally affected in the same way;
c) the ground floor is partly intact in most of the buildings, whereas the upper floor is almost completely demolished;
d) the E-W oriented walls are by far more damaged than those striking N-S;
e) in many cases, the northern vent-facing part of the buildings was more damaged than the southern one;
f) most of the destruction is stratigraphically related to unit E.

As shown in Table 1, 394 corpses were found in the pumice fall deposit and 650 in the pyroclastic density currents (PDCs) deposit. So a total of 1044 victims were recovered inside 2/3 of the city of Pompeii (the excavated part). Other 100 victims are estimated on the basis of many groups of scattered bones. Finally, considering the regions partially excavated (I, III, IV, V, IX) an estimate of 464 corpses still buried is obtained. Furthermore, it is meaningful to document the amount of victims with respect to their location (e.g. inside or outside the buildings). Most of the corpses within the pumice fall deposit were found inside buildings (80% as shown in Table 1), whereas of the 650 corpses recovered in the PDCs deposit 334 were found inside buildings and 316 outdoor (Table 1). Luongo et al. (2002b), on the basis of recent excavation, state that all human casts in the PDCs deposit lie over the well-recognisable lithic-rich unit D, enclosed within unit E (Figure 2). These corpses are mostly intact and only few corpses are partially or fully dismembered. In the still preserved outcrops of Pompeii and in the photographs of the Pompeii Archive, most of the casts lay prone in the attempt to shelter their face; it is noteworthy that in some places (e.g. garden of fugitives, Regio I, Insula 21; Stabian house, Regio I Insula 22) human casts show the head and the bust supported by arms, with this raised part of the body at higher stratigraphic level within E1.

Table 1 Total of corpses found in the pumice fall deposit and in the stratified ash and pumice PDCs deposit (into buildings and outdoor places) at Pompeii.

	Victims in the Fall deposit		Victims in the PDC deposit		
	Indoor areas (a)	Outdoor places (b)	Indoor areas (a)	Outdoor places (b)	Total
External areas	17	17	49	70	153
Regio I	66	9	86	41	202
Regio II	12	7	26	73	118
Regio III	9			4	13
Regio IV	1	1			2
Regio V	40		11	16	67
Regio VI	41	2	39	35	117
Regio VII	59	5	33	14	111
Regio VIII	21	5	43	51	120
Regio IX	69	1	43	7	120
Unknown location	10	2	4	5	21
Subtotal	345 **(88%)**	49 **(12%)**	334 **(51%)**	316 **(49%)**	
Total	394 **(38%)**		650 **(62%)**		1044

The effects of the 79 AD eruption on Pompeii

The process of Pompeii's destruction and burial started with the accumulation of a thick pumice lapilli deposit (layers A and B in Figure 1) resulting from the column fallout. The rate of deposition in the city ranged from 15 cm per hour in open areas to 25/30 cm per hour in places accumulating additional pyroclasts rolling from the steeper roofs. Within six hours from the beginning of the eruption the roofs and part of the walls of the buildings had collapsed under the pumice load. By the morning of 25 August most structures were seriously damaged; the pumice fall deposit, generally 3 m thick, totally buried the lower part of the buildings. The percentage of victims (38%) found in this deposit at Pompeii is anomalously high with respect to a mean of 4% of deaths caused by tephra fallout in the last four centuries during explosive eruptions (Blong, 1984; Tanguy et al., 1999). This high percentage of deaths is possibly due to the attempt of some people to take shelter into buildings where roofs and walls collapsed under the load of the pyroclastic material. The small percentage of people found dead outdoor within the pumice fall deposit was probably killed by the crumbling roof tiles or by the largest lithic fragments following ballistic paths.

The first PDC flowed through the city depositing the basal ash layer C and causing irrelevant damages. Based on the evidence that all of the human remains lie above this deposit, it can be deduced that people were not killed by the earlier PDC (units S4 and S5 of Sigurdsson et al., 1985). The inhabitants survived also to the successive fallout phase that emplaced the lithic-rich layer D and some

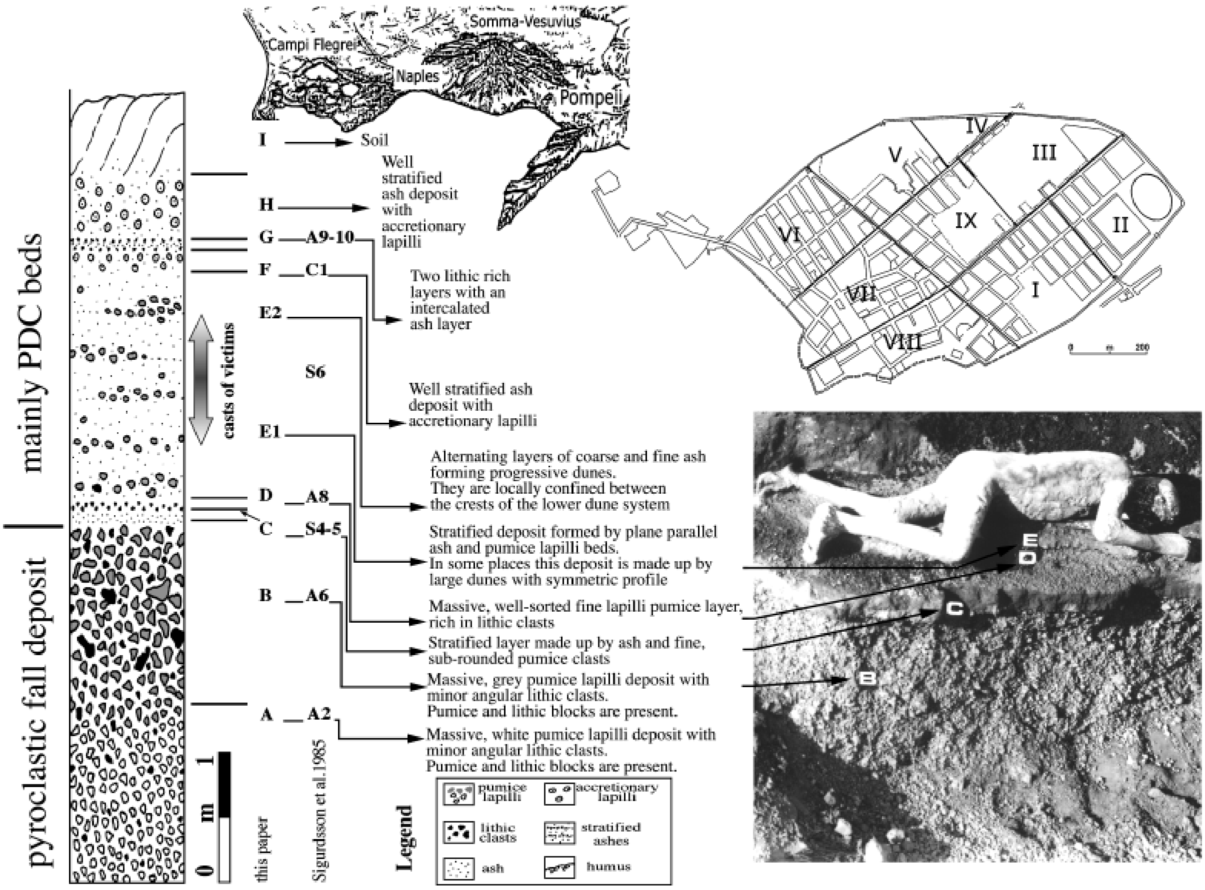

Figure 2 Stratigraphic section of the 79 AD deposits at Pompeii with lithofacies characteristics and emplacement mechanisms reported for each layer. The maximum thickness for each unit is represented. On the right, a human cast near the casa del Criptoportico (photo from the Archives of the Soprintendenza Archeologica di Pompeii) stays above a well-exposed stratigraphic sequence (unit B to E1). Above: Campanian Plain with Somma-Vesuvius volcanic complex and the city of Pompeii. Pompeii map and location of regions (roman numbers). Unornamented areas represent unexcavated regions.

were able to walk outdoor during the emplacement of the basal part of the unit E. In fact, we found the victims several centimetres above the base of this unit. Possibly, the parental pyroclastic current ran over the city with a low-temperature, dilute frontal part settling progressively few centimetres of ash. The rear part of the current had a non-uniform behaviour in terms of concentration, possibly due to the canalization of the basal part of the current along the longitudinal walls of the buildings. Inside these areas the current showed a greater destructive power, flattening most of the (especially transversal) walls, standing out of the pumice fallout deposit, in its north-south path. In the areas outside the channels the current had essentially a depositional behaviour engulfing the city and suffocating the inhabitants. These opposed behaviour of the PDC in very close areas testify to the different kinetic conditions undergone by the Pompeii inhabitants and hence the different physical integrity of their corpses. Observations on objects, cloths, frescoes and skeletons rule out the possibility that burn injuries contributed to kill Pompeii inhabitants, as recently proposed for Herculaneum inhabitants (Capasso et al., 2000; Mastrolorenzo et al., 2001). Furthermore, the proposed non-uniform behaviour of PDCs, due to the interaction with the urban structures, justifies the described different state of destruction of the buildings throughout the city.

A final phreatomagmatic phase, punctuated by two minor lithic fall episodes, emplaced the upper part of the succession (F to H units). Field features, such as the presence of accretionary lapilli in the upper part of the ash and pumice deposit and the lack of high temperature evidences in the buildings, support the idea of low emplacement temperature for the pyroclastic currents during the final phase of the eruption.

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Lisetta Giacomelli, *geologist, is associated with Gruppo Nazionale di Vulcanologia; the main field of interest is in general volcanology, and currently is involved in the creation and maintanance of didactic web pages on earth sciences.*

Annamaria Perrotta, *geologist, is associated with Dipartimento di Geofisica e Vulcanologia, Universita' "Federico II" of Naples; the main field of interest is related with the depositional mechanisms of the products of explosive eruptions.*

Roberto Scandone, *geophysicist, is Professor of Physical Volcanology at Universita' "Roma Tre" of Rome; his main field of interest is related with the mechanisms of eruptions.*

Claudio Scarpati, *geologist, is Researcher at Dipartimento di Geofisica e Vulcanologia, Universita' "Federico II" of Naples; the main field of interest is related with the depositional mechanisms of the products of explosive eruptions.*

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

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

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

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

Introduction

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

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

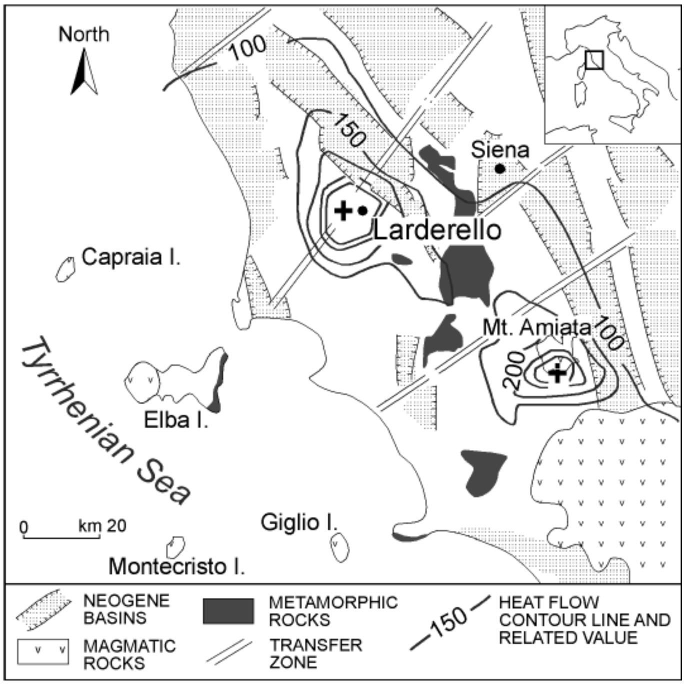

Figure 1 Structural sketch map of southern Tuscany with the regional heat flow contour lines (equidistance: 50 mW/m²). The Larderello-Travale and Mt. Amiata geothermal fields are located in areas where heat flow reaches 1 W/m² and 0.6 W/m², respectively. (after Baldi et al., 1995)

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

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

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

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

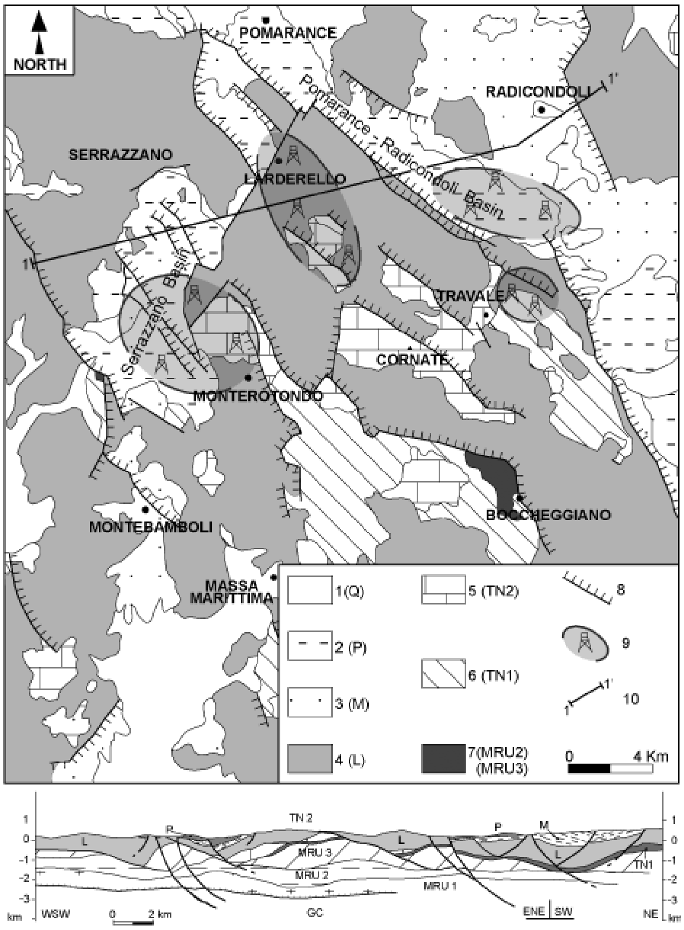

***Figure 3** Geological sketch map of the Larderello-Travale area. Key: Neogene and Quaternary deposits: 1—Quaternary continental sediments; 2—Pliocene marine sediments; 3—Miocene continental and marine sediments; 4—Ligurian units l.s. (Jurassic- Eocene); 5—Tuscan Nappe: Late Triassic-Early Miocene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

Geological features of the Larderello-Travale and Monte Amiata areas

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

Larderello-Travale Area

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

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

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

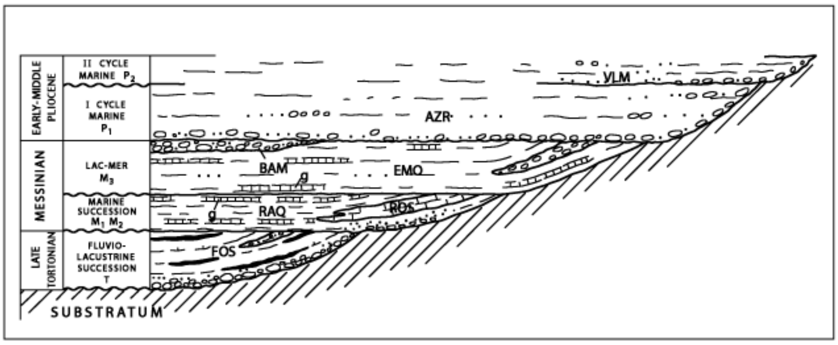

***Figure 4** Stratigraphic relationships among the Neogene formations cropping out in the surroundings of the Larderello-Travale geothermal area; RAQ: Raquese Stream Fm. (Early Messinian), ROS: Rosignano Limestone (Early Messinian), EMO: Clays and gypsum of Era Morta River (Late Turolian), BAM: Montebamboli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

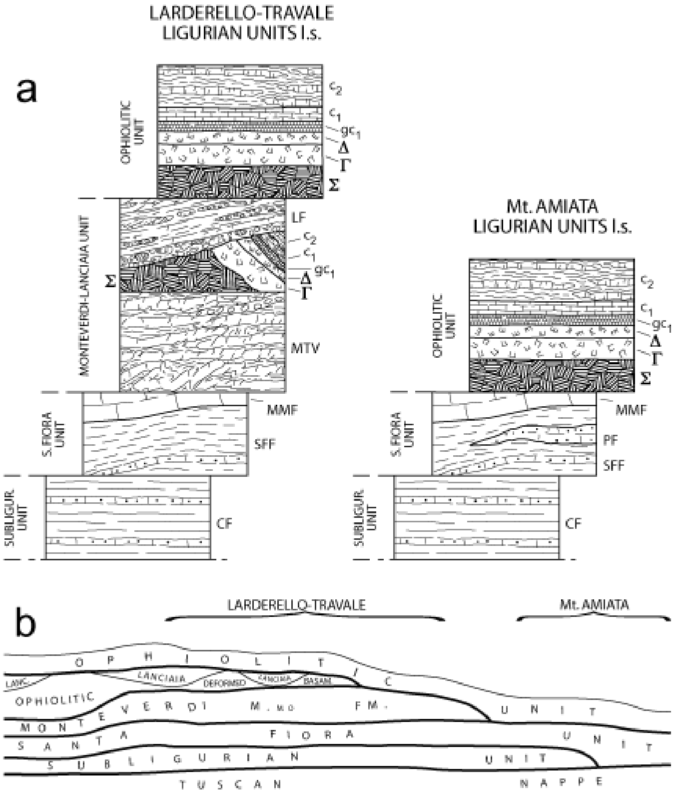

Figure 5 a) Structural and stratigraphic relationships between the Ligurian and Subligurian Units in the Larderello-Travale and Mt. Amiata geothermal areas. Ophiolitic Unit: Δ, Γ, ∑: Ophiolites (serpentinites, gabbros, basalts—Middle-Late Jurassic), gc_1: Mt. Alpe cherts (Late Jurassic), c_1: Calpionella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

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

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

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

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

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

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

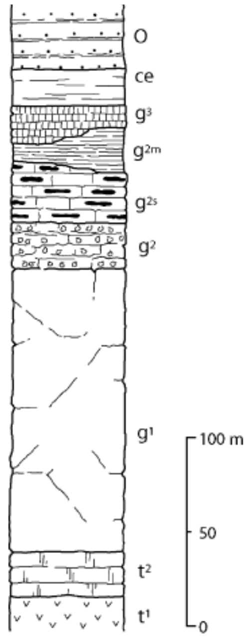

Figure 6 Stratigraphic succession of the Tuscan Nappe.
Symbols: O: Macigno Fm. (Late Oligocene-Early Miocene); ce: Scaglia toscana Fm. (Cretaceous-Oligocene); c^1: Maiolica Fm. (Early Cretaceous); g^3: Diaspri Fm. (Malm); g^{2m}: Marne a Posidonia (Dogger); g^{2s}: Calcare Selcifero Fm. (Middle-Late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

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

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

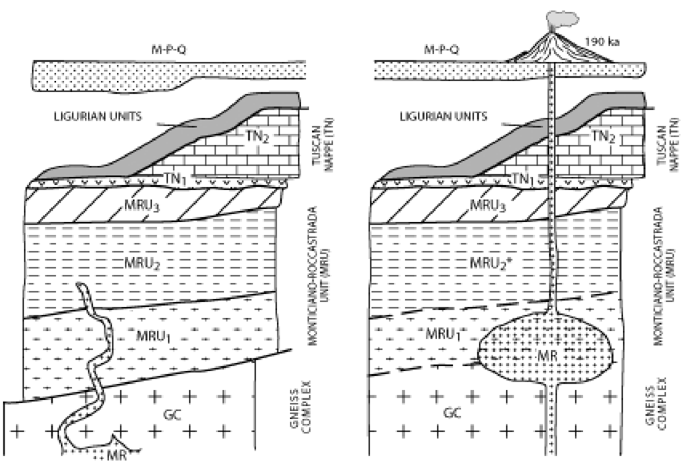

Figure 7 Tectonostratigraphic units in the Larderello-Travale and Mt. Amiata areas. Q-P-M: Quaternary, Pliocene and Miocene sediments; MR—Magmatic rocks; Tuscan Nappe (TN): TN_2—Early Miocene-Rhaetian sequence; TN_1—Late Triassic evaporite (Burano Fm.); Monticiano-Roccastrada Unit (MRU): MRU_3—Triassic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

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

Mt. Amiata Area

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

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

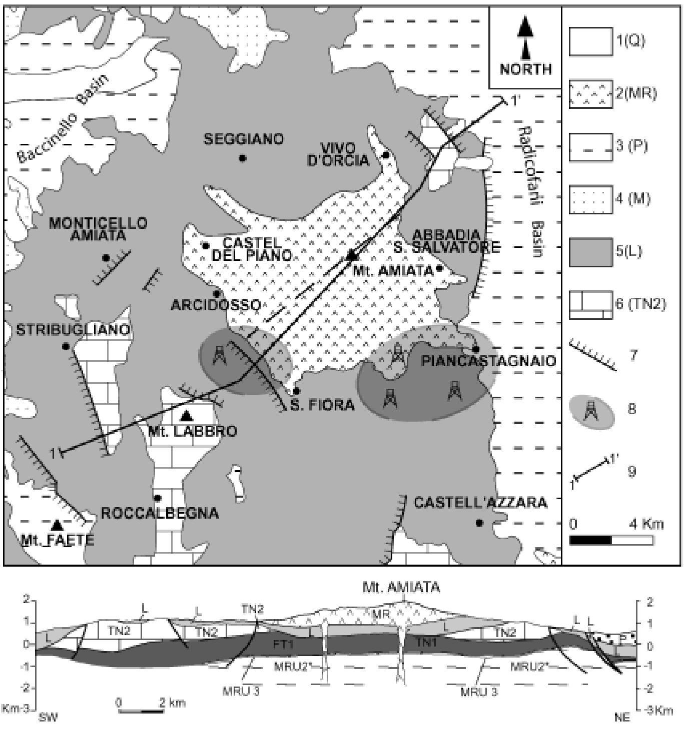

Figure 8 Geological sketch map of the Mt. Amiata area. Keys: 1—Quaternary continental sediments; 2—Magmatic rocks; 3—Pliocene marine sediments; 4—Miocene continental, brackish and marine sediments; 5—Ligurian Units l.s. (Jurassic-Eocene); 6—Tuscan Nappe (Late Trias-Early Miocene); 7—normal faults; 8—Main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

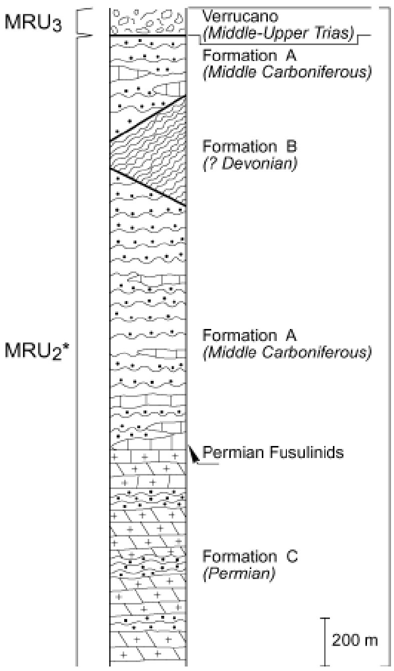

***Figure 9** Relationship among the Triassic and Palaeozoic formations belonging to Monticiano-Roccastrada Unit encountered by geothermal wells in the Mt. Amiata area (after Elter & Pandeli, 1991).*

Extensional structural features

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

Reflection seismic features

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

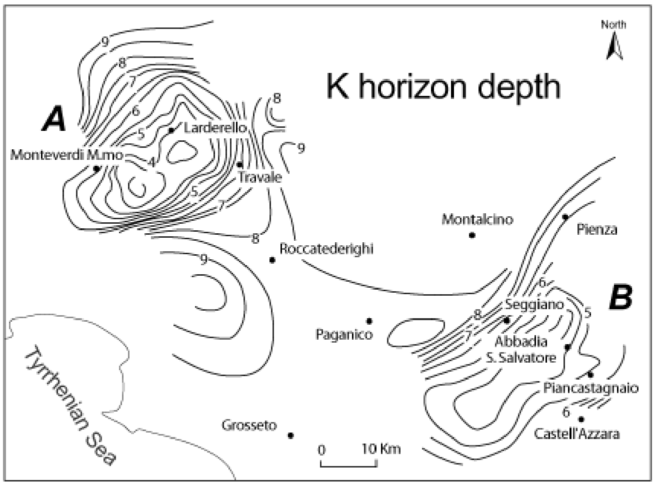

Figure 10 Contour lines in kilometres (equidistance: 0.5 km) of the K-horizon depth. A and B show respectively the Larderello-Travale and Mt. Amiata geothermal areas (after Cameli et al., 1998).

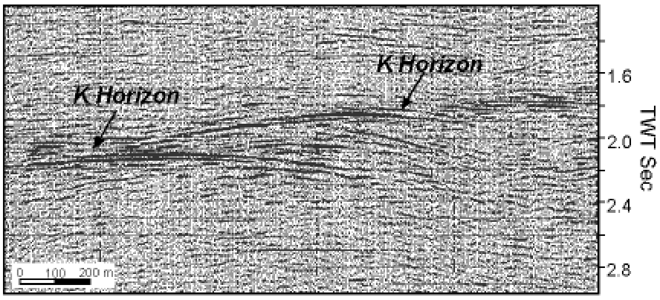

Figure 11 The K-horizon in the Larderello-Travale geothermal area. Vertical axis: TWT seconds.

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

Geothermal reservoirs and fluids

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

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

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

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

Concluding remarks

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

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

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

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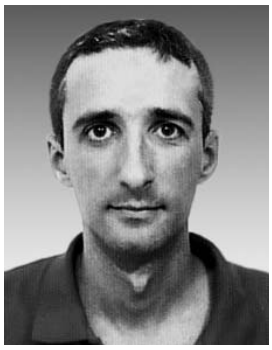
Fausto Batini *is manager of geothermal exploration of the Enel Green Power Italian electric company. He has published numerous research papers on the Italian geothermal fields.*

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

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

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



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

by Gianluca Valensise, Alessandro Amato, Paola Montone, and Daniela Pantosti

Earthquakes in Italy: past, present and future

Istituto Nazionale di Geofisica e Vulcanologia - Via di Vigna Murata, 605 - Rome, Italy. *valensise@ingv.it* *http://www.ingv.it/*

Italy has a long-standing tradition of earthquake investigations. Seismologists can rely on one of the longest and most detailed records of historical seismicity, 20 years of homogeneous and reliable instrumental data, systematic and widespread active stress data and a comprehensive database of potential seismogenic sources. Here we describe these datasets and discuss how they may help us anticipate the large earthquakes of the future.

1. Introduction

According to UNESCO, densely populated Italy hosts about 50% of the known cultural heritages of the entire globe. Italian cities are thus especially vulnerable to earthquakes that would not be as dangerous in other earthquake-prone countries. Damage to the frescoed vault of the San Francesco basilica in Assisi during the 1997 Umbria-Marche earthquake (M_w 6.0) exemplifies Italy's special vulnerability.

Saving lives and protecting Italy's immense cultural patrimony are important responsibilities of Italian seismologists. But can the location and size of future large Italian earthquakes be anticipated with some confidence? Seismologists certainly know Sir Charles Lyell's famous motto "the present is the key to the past". They also know by everyday experience that the corollary to that statement, "the past is the key to the present", is just as true. But, most importantly for those concerned with the future of our planet, they are also learning that "the past and the present are the key to the future". The seismological translation of this motto is that "successful forecasting means projecting into the future what we learned from the past and witness in the present".

There are few countries where the earthquake record can be exploited at time scales from 10,000 years to seconds, but Italy is one of them. This record, properly organized into a knowledge continuum, allows several facets of the earthquake process to be investigated simultaneously. This paper describes how seismological information (instrumental and historical) and field evidence at different scales can be integrated to make realistic long-term earthquake forecasts in Italy. In the last section of this paper we propose a viable approach to this problem by introducing the concept of *potential seismogenic source* (Valensise and Pantosti, 2001a) a simplified three-dimensional scheme of a physical fault system.

2. The past: Italy, the cradle of earthquake catalogues

Italian seismicity today does not appear as frightful as it did a few centuries ago, at least in comparison with much larger earthquakes that take place in other parts of the globe. Nevertheless, Italian earthquakes have left a strong imprint on the country's landscape, heritage and traditions, and even place names and the distribution of dialects often reflect a large earthquake. Italians also learned early that to fight earthquakes you ought to know them: where they occur repeatedly, what damage they cause, what to expect after the main shock has struck. Major advancements in understanding earthquakes were spurred by catastrophic events that attracted the interest of contemporary scholars. Such is the case of the central-southern Italy earthquakes of 1456, which devastated a large portion of the peninsula between Abrutii and Basilicata. Giannozzo Manetti's *De terraemotu libri tres*, published in 1457, is the first known systematic study of Italian seismicity after Antiquity (see Guidoboni, 2000 for a review).

For over two centuries the cataloguing efforts continued, although the compilers often emphasised more the "portentous" nature of earthquakes rather than the associated risk. Marcello Bonito's *Terra Tremante*, written after the catastrophic 1688 Benevento earthquake (Campania, southern Italy), is the principal expression of seismology of that time and the herald of modern earthquake catalogues. Around the middle of the XVIII century, however, these efforts were somewhat frustrated by the popular theory of electricity that effectively detached earthquakes from geology and landscape to make them totally random.

The XIX century marked the onset of modern catalogues, which had by then become a real scientific tool. Earthquakes were again seen as a *systematic* and *recurrent* phenomenon stemming from "seismic centers", just as volcanic eruptions were known to take place always at the same spots. Modern catalogues thus became the basis for demonstrating the regularity of earthquakes, the main pillar of modern earthquake forecasts. Towards the end of the century, Italian seismologists Michele Stefano de Rossi and Giuseppe Mercalli devised the first systematic scales of earthquake intensity. By allowing objective assessment of earthquake effects at many localities, this step favored the transition from descriptive to modern parametric catalogues. It also allowed different events to be compared through their epicentral location and severity.

The great catalogue by Mario Baratta (1901), a compilation of nearly 1,400 Italian earthquakes between Antiquity and 1898, is the most mature outcome of these new trends. Perhaps due to the enormous advancement represented by this work and the concurrent rise of instrumental seismology, not much happened until 1935 when Alfonso Cavasino (1935) published an update of Baratta's catalogue. Ironically, his 35-year update had to account for four of the most dreadful earthquakes of Italian history: 1905, central Calabria; 1908, Messina Straits; 1915, Avezzano; 1930, Irpinia: all having M~7.0. Data collection became then systematic, leading to the first computerized catalogue in 1973. In the booming '70s and early '80s, also as a result of catastrophic earthquakes (1976, Friuli; 1980, Irpinia), research into historical seismicity had become a national priority. It is hence not surprising that a "consensus" catalogue published in 1985 contained almost 42,000 events (Postpischl, 1985).

The 1990s brought about yet another revolution in the task of retrieving, analyzing and storing historical earthquake data. In 1995, INGV released the first version of the *Catalogue of Strong Earthquakes* in Italy, a "new-generation" compilation that summarizes research conducted by ENEL (Italy's national electricity company) and the former ING (Istituto Nazionale di Geofisica) (Boschi et al., 2000). In addition to the standard source parameters, for each earthquake this catalogue supplied a set of specifically prepared summaries, details on the effects suffered in each locality involved, and a list of references. Two years later, this work was paralleled by the Internet-based compilation prepared within Italy's CNR (National Research Council) (Monachesi and Stucchi, 1997). Both compilations served as a basis for the *Catalogo Parametrico dei Terremoti Italiani* (CPTI Working Group, 1999), an updated version of the work completed by Postpischl in 1985. Individual datapoints allowed earthquake source parameters to be determined through an analytical technique which gave a statistical basis to subjective historical data (Gasperini et al., 1999). For the best documented earthquakes this technique also returns an estimate of source orientation and length, thus allowing the determination of a "virtual fault" for earthquakes that occurred centuries ago.

After investigating surface ruptures produced by the 1980 Irpinia earthquake (M_S 6.9; e.g. Pantosti and Valensise, 1990), scientists started looking at historical earthquakes from the point of view of their impact on the environment (see Valensise and Guidoboni, 2000 for a review). Identifying major historical earthquake ruptures thus became a challenging but promising activity for many Quaternary geologists.

A further revolution in investigating Italy's past earthquakes is the onset of paleoseismological studies in the late '80s. Trenching of the fault responsible for the 1980 earthquake supplied evidence for at least four of its predecessors occurring every 2,000 yr on average (Pantosti et al., 1993). Subsequent trenching of a number of large faults confirmed these findings and returned fundamental constraints on the frequency of large Apennines earthquakes (see Valensise and Pantosti, 2001b for a review).

A recent attempt to extend the historical record back in time is the development of Archeoseismology, a discipline that explores evidence for destructive earthquakes emerging from the archeological record. For instance, a thorough reappraisal of the history of settlements and land-use in the Messina Straits during Antiquity supplied evidence for a large event of social and territorial disruption in the IV century (Guidoboni et al., 2000). This event was interpreted as a predecessor of the devastating 1908 earthquake, agreeing with geological estimates of 700–1,500 yr for the average repeat time of 1908-like events (Valensise and Pantosti, 1992). Additional promising work is progressing at different sites throughout the peninsula.

Combining paleoseismological and archeoseismological evidence suggests that (1) the typical repeat time for individual major Italian earthquake sources is 1,000–3,000 yr, and (2) apparently shorter recurrence intervals result from multiplicity of sources or stress triggering of events on adjacent sources. These findings are supported by missing "twins" of the very same earthquake throughout history. Although the infrequency of Italian quakes is clearly favorable, the mismatch between typical repeat times and the length of the reliable historical record (~700 yr) implies that several important potential earthquake sources are not represented in current historical catalogues. One of them could go off in the next large earthquake, turning it into yet another "unexpected event".

Should we then suspect historical catalogues? Of course not. We will see later how this condition may be handled by carefully combining all available historical, geological, instrumental evidence. Meanwhile, Italian seismologists, planners, civil defense authorities and ordinary people are learning from the historical record. The earthquake distribution revealed by the CPTI (Figure 1) is surprising, because seismicity is widespread and moderate shocks seem to occur nearly everywhere; but unsurprising also, because most large earthquakes occur where they would be expected, given the current understanding of Italy's geodynamics.

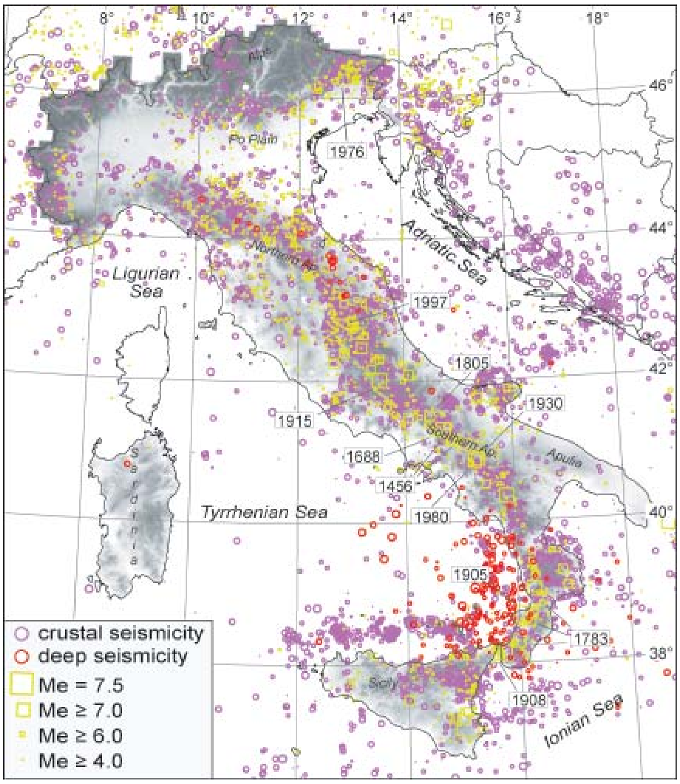


***Figure 1** Historical seismicity from CPTI catalogue (CPTI Working Group, 1999) and instrumental seismicity from INGV bulletin (1983–2002, M≥2.8), both scaled by magnitude. Location of large earthquakes cited in text is also shown.*

3. The present: understanding Italy's stress, strain and seismicity

Present seismicity is the most direct evidence for ongoing tectonic activity. Unlike the causative faults of historical earthquakes, for which we normally recover only an approximate estimate of size and location, modern seismological instrumentation allows the most significant parameters to be assessed confidently. These include the hypocentral location and depth, source size, coseismic displacement, faulting mechanism and details of the rupture process.

In Italy, instrumental seismology started more than a century ago, and all large earthquakes of the XX century have been recorded by several seismic observatories. Recovering early historical seismograms and investigating these earthquakes have already proved very promising (e.g., Pino et al., 2000) under SISMOS, a major project that started recently at INGV.

All magnitude 5.5^+ earthquakes of the past 30 years have been investigated by combining sparse data from "historical" Italian observatories and global networks that became operational in the early '70s (see Gasparini et al., 1985; Anderson and Jackson, 1987; INGV RCMT, 2003; among many others). Further details on seismic rupture, such as source duration, directivity and complexity, have also been retrieved thanks to the strong motion instruments formerly operated by ENEL and presently by Servizio Sismico Nazionale (SSN).

Systematic collection of short-period earthquake data began after the 1980 Irpinia earthquake, when the centralized Italian National Seismic Network began. Since then, the network has been steadily extended and improved in instrumentation and data reliability. Tens of thousands of earthquakes having magnitudes between ~1.5 (the detection threshold in the best instrumented parts of the country) and 6.0 (M_W of the 1997 Umbria-Marche event) were recorded. Several regional and local networks also contributed to highlight details of Italy's "earthquake structure". After twenty years of uninterrupted surveying, the Italian dataset of instrumental seismicity indeed represents a powerful tool for investigating the region's active tectonics (Figure 1). The resolution offered by this dataset, however, is insufficient to illuminate details of the seismic process in fault zones. Most information on Italian seismogenic faults at depth comes from dense, state-of-the-art portable networks temporarily deployed for aftershock studies. Important details on the geometry, kinematics and complexity of fault systems were identified with this tool. For example, detailed analyses of the 1997 Umbria-Marche sequence show that the fault zone changes within a very short distance from a simple linear, low-angle rupture plane to a complex flat-ramp geometry (Chiaraluce et al., 2003). Overall, the data highlighted the role of shallow structural complexities, both parallel and perpendicular to the main fault, in determining rupture behavior (Chiarabba and Amato, 2003).

Besides aftershock studies, advancements in the understanding of active tectonics and fault zones were spurred by field experiments based on modern seismological arrays and large mass storage capabilities. The first experiment (2001–2002) was a six-month campaign that shed light on the geometry and kinematics of major active fault systems in northern Umbria (central Apennines). The results confirmed that these experiments are extremely promising, provided that the array is well designed and the earthquake rate is significant in comparison with the detection threshold (the largest recorded quake was only 3.0).

A recent but valuable source of information on Italy's active tectonics is the systematic analysis of in-situ stress data, essentially borehole breakouts in deep wells. It is now widely accepted that computing their average azimuthal distribution yields the local orientation of the horizontal stress field (Bell and Gough, 1983; Zoback et al., 1985). In the early nineties INGV started systematic activity in this field by cooperating with ENI-AGIP and other oil and geothermal companies in Italy. Over 300 wells have been analyzed so far (Montone et al., 1999). Along with earthquake fault plane solutions and fault slip data, breakout analyses yielded an image of present-day stress orientation in Italy (Figure 2) which now contains over 500 minimum horizontal stress directions (S_{hmin}) (Montone et al., 2003, and references therein). Stress directions obtained from different data have been found mutually consistent, despite the fact that they often refer to different depth intervals (0–7 km for breakouts, 0–20 km for crustal earthquakes) and different tectonic units. The inferred pattern confirms that the stress field in Italy's brittle crust is due to the interaction of "large scale" processes with "local sources". Stress data can be used to understand active tectonic processes, assess seismic hazard, and anticipate the behavior of seismogenic faults. Such data also represent a benchmark for geodynamic modeling of the central Mediterranean.

Information on the present strain field is commonly obtained from geodetic observations spanning at least a decade of ongoing tectonic activity. Unfortunately, Italy's strain field is still poorly known due to (1) the lack of a homogeneous and dense GPS network, and (2) the low ratio between strain rates and accuracy of the measurements. Nevertheless, first order estimates have already been obtained based on IGM95 (a network established and maintained by Istituto Geografico Militare) (D'Agostino et al., 2001), periodic resurveys of regional networks (Anzidei et al., 2001; Calais et al., 2002), and GPS resurveying of the national triangulation network set up by IGM in 1869–1881 (Hunstad et al., 2003). Owing to the exceptionally long time-span in comparison with the estimated uncertainties, this latter technique recently revealed that (a) peninsular Italy extends at 2.5–5.0 mm/y, (b) deformation is concentrated in a region a few tens of km wide straddling the crest of the Apennines, and (c) 70–80% of this strain has not been released by earthquakes and must occur aseismically.

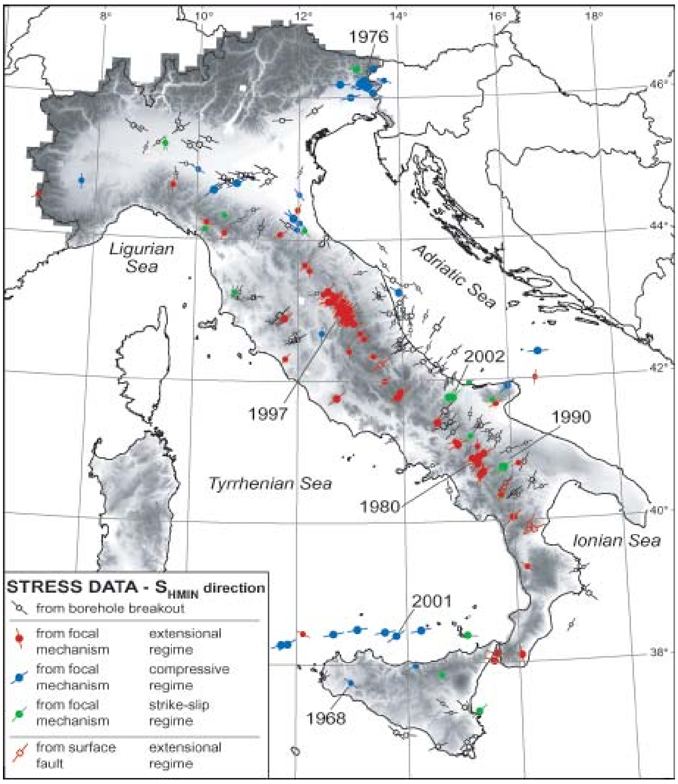


Figure 2 Active stress map of Italy with minimum horizontal stress orientations (from Montone et al. , 1999, slightly modified and updated), and location of selected significant recent earthquakes.

By comparing recent and historical seismicity, the stress and strain fields gathered from all available data and the deep structure recovered from seismic tomography, seismologists were able to set firm constraints on Italy's active tectonics (refer to Figures 1, 2 together):

- most of Italy is seismically active, except for a few stable areas such as Sardinia, southern Apulia and limited parts of northern Italy and Sicily;
- this almost continuous seismic belt follows the buried margin of the Adriatic microplate beneath the Apennines and the Alps; the boundary between the Ionian lithosphere and the Calabrian arc; the boundary between the "African" foreland and the Sicilian nappes in Sicily;
- earthquakes are sparse along the Alpine arc, from the Ligurian Sea to Friuli, with a lesser density in the central Alps. Alpine earthquakes directly testify to the ~N-S push of the Adriatic microplate, as shown by the thrust faulting mechanism of the 1976 Friuli events;
- widespread seismicity characterizes the northern Apennines arc, from northern Tuscany to Latium. Spots of low-magnitude seismicity punctuate Quaternary volcanoes and geothermal fields in the inner portion of the arc. The actively deforming belt extends from the inner Apennines basins, where seismicity results from crustal extension perpendicular to the belt (S_{hmin} oriented NE-SW), to the most external arcs buried beneath the Po Plain and the Adriatic offshore, where earthquakes reveal ongoing convergence (S_{hmin} oriented NW-SE). Although the mechanism driving this arc evolution is still debated, the presence of earthquakes down to 90 km depth (Selvaggi and Amato, 1992) and a strong high-velocity anomaly in the upper mantle beneath the northern Apennines (e.g. Lucente et al., 1999) support an actively retreating slab. This view is challenged by investigators who explain the same evidence in a context of prevailingly passive intra-continental rift (e.g., Lavecchia et al., 2003). In the northern Apennines, seismological evidence for extension and compression consistently agrees with the Neogene evolution of the region, although some investigators contend that compression died out before the end of Middle Pleistocene (Di Bucci and Mazzoli, 2002).
- a relatively narrow belt of seismicity characterizes the southern Apennines, where earthquakes are larger (M~7) than in the northern arc and individual faults are up to 30–50 km long. Earthquake and active stress data consistently reflect dominant ongoing NE-SW extension. Conversely, the role of E-W right-lateral strike-slip faults revealed by recent moderate-size earthquakes (e.g. Potenza, 1990; Molise, 2002) and affecting the Apulian foreland is still unclear. Such quakes may result from lateral complexities, accommodate different rates of extension, or represent the reactivation of pre-existing major faults dating back to the construction of the Apennines edifice;
- contrary to what is revealed by seismic tomography in the northern Apennines, the deep structure of the southern Apennines is dominated by a detached (or possibly less dense) lithospheric slab, which could be the engine of uplift and extension of the orogen;
- a narrow Wadati-Benioff zone is beautifully imaged in the Calabrian arc. Shallow earthquakes are recorded inland, whereas intermediate and deep events are recorded beneath the southern Tyrrhenian. The subduction zone is only 200 km wide but extends much more down-dip, reaching and possibly penetrating the 660 km discontinuity (Lucente et al., 1999; Amato and Cimini, 2001; Piromallo and Morelli, 2003). Intriguing features of this deep activity are the geometry of the Ionian slab, which

appears to sink passively in the mantle, and the slab's interior earthquake mechanism, a consistently down-dip compression;

- seismicity of Sicily results from processes interacting at very short distance and giving rise to an especially complex setting: roughly E-W extension in the Messina Straits, consistent with what is seen in the adjacent Calabrian arc; dense spots of largely extensional earthquakes around Mt. Etna; strike-slip earthquakes along eastern Sicily at the boundary between the Ionian lithosphere and the Sicilian foreland; a transverse structure bounding the Calabrian subduction to the west (Tindari-Letoianni lineament auct.); N-S compressional earthquakes in northern (2001 Palermo, M_w 5.4) and central-western Sicily (1968 Belice, M_w 5.6).

4. The future: long-term earthquake forecast Italian-style

In 1999, the Italian Dipartimento per la Protezione Civile (Civil Defense) launched a plan for seismic retrofitting of the country's historical centers and sought help from the scientific community to establish intervention priorities. This was not a standard or easy scientific problem for Italy, where seismic hazard is traditionally assessed based on the statistical distribution of historical and instrumental earthquakes with subordinate geologic and tectonic constraints (e.g. Albarello et al., 1999). This approach rather accurately determines rates of non-destructive earthquakes, constrains the maximum credible magnitude over extended regions (~100 km) and supplies reasonable probabilistic estimates of the expected ground shaking. However, little information on the physical properties of major seismogenic sources can be directly deduced from historical catalogues, and the contribution of instrumental data is clearly limited to earthquakes that occurred after the inception of modern seismological networks. Identifying large active faults geologically is normally a viable alternative, but Italy's complex tectonic history makes this approach quite problematical. In summary, none of these basic contributions (historical and instrumental seismicity, geodesy, geologic and tectonic data) may be used alone either to anticipate the location and size of future large earthquakes or to assess the characteristics of the associated ground shaking.

In 2000, a team of INGV scientists gathered to submit a large research project within the 2001–2003 Framework Program of GNDT (Gruppo Nazionale per la Difesa dai Terremoti, funded by the Italian Civile Defense: *http://gndt.ingv.it/*). The project, entitled "*Probable earthquakes in Italy 2000–2030: guidelines for determining priorities in seismic risk mitigation*", was intended as a direct response to the government request for identifying priority areas for seismic risk mitigation. This project was to blend several recent developments in Italian seismology and earthquake geology. These include fault identification and characterization with different tools; evaluation of seismic and geodetic strain to constrain hazard estimates; studies of regional seismic wave propagation and attenuation; studies of local site response; and development of time-dependent hazard evaluation tools incorporating innovative information on fault behavior and earthquake recurrence.

The starting point of the project's strategy is a fault segmentation model of Italy that was developed by INGV scientists starting in 1996. The model rests on the assumption that seismicity may be approximated by a finite number of *potential seismogenic sources*. It was based on available good historical data (see § 2), important findings arising from instrumental seismicity (see § 3) and the awareness of the specificity of active faulting in Italy, and was eventually structured as a GIS-supported database ("Database of Potential Sources for Earthquakes Larger than Magnitude 5.5 in Italy": Valensise and Pantosti, 2001a).

A *potential seismogenic source* is the surface projection of an inferred fault which is likely to experience a significant earthquake in the future. Not all potential sources will have earthquakes, and not all earthquakes will necessarily occur on identified potential sources, but in general the list of seismic sources should serve well as a basis for hazard estimation. Some potential sources have been identified from faulting or shaking in past earthquakes, while others have been inferred from more indirect observations. Each source is assumed to represent a discrete segment of a longer seismogenic zone. In its turn, each segment is assumed to generate its largest allowed or "characteristic" earthquake, not necessarily periodic. Although seemingly crude, these assumptions are being proved realistic in regions dominated by dip-slip faulting away from the main active plate boundaries, such as Italy.

The current release of the segmentation model (Figure 3) lists about 250 potential earthquake sources grouped according to their identification criteria and parameter assessment. New sources were added or improved during the GNDT project, particularly in poorly documented and strategic areas of the country. Most sources align along well-established tectonic trends coincident with those identified by instrumental and historical seismicity (see § 2, 3). Overall the Italian peninsula can be seen as a segmented belt of discrete seismogenic faults, at least to a first approximation. The best-identified segments are often neatly juxtaposed; their boundaries often coincide with "transverse lineaments" inherited from past tectonic phases and currently imprinted in the highly heterogeneous Italian crust.

Careful inspection of Italy's seismogenic sources reveals that some segments or longer sections of the seismogenic belt are not associated with known historical events. Can they be simply regarded as "seismic gaps"? Should we assume these areas have been accumulating stress for several centuries or millennia, and therefore are much more likely to rupture in the near future than any other Italian seismogenic area? Probably, but we must be very cautious. These faults certainly accommodate part of the deformation accumulated by ongoing tectonic activity. In the long-term (10^4–10^5 y) the process of recharge and sudden release of tectonic stress is relatively periodic, but in the short-term (10^2–10^3 y) it is certainly modulated by such effects as stress transfer after large earthquakes (e.g.

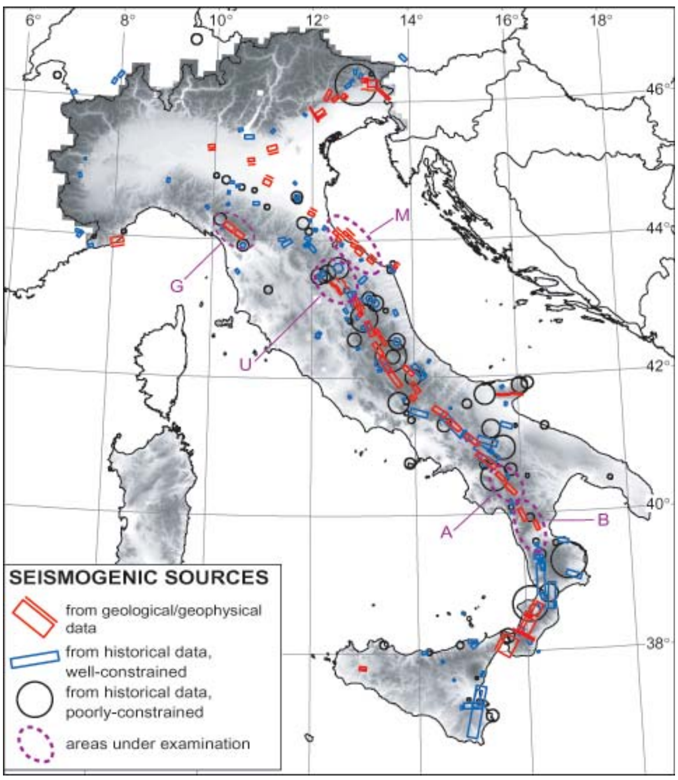


Figure 3 ***Segmentation model of Italy, showing main trends of identified seismogenic sources (derived from Valensise and Pantosti, 2001a). Areas outlined by a dashed line contain at least one silent seismogenic source and are the object of investigation by the GNDT project "Probable earthquakes in Italy 2000–2030" (from north to south: Garfagnana, northern Tuscany; northern Marche coastal belt; northern Umbria; Agri-Melandro, southern Apennines; Basilicata-Calabria border). See project's reports and Valensise and Pantosti (2001a) for further details.***

King and Cocco, 2000), lateral heterogeneities inducing creeping behavior of sections of the belt, or others yet to be explored.

Some potential gap areas (highlighted in Figure 3 and listed in caption) are currently being investigated by several of the GNDT project participants. Understanding the nature and role of structural complexities, the rate of tectonic stress accumulation vs. stress release in individual earthquakes, the ratio of seismic vs. non-seismic strain, the variations of stress and faulting style with depth are some of the challenges faced by Italy's modern seismology and earthquake geology. Nevertheless, positive identification and characterization of these the potential gaps and the hypothesized structural control have obvious implications for seismic hazard. Several forecasts concerning the locus of rupture initiation/termination and the approximate size of forthcoming large earthquakes will be issued at the end of the project. Success of even one would prove that modern seismology is ready to win against the apparently random nature of earthquakes, and that Sir Charles Lyell was right once again.

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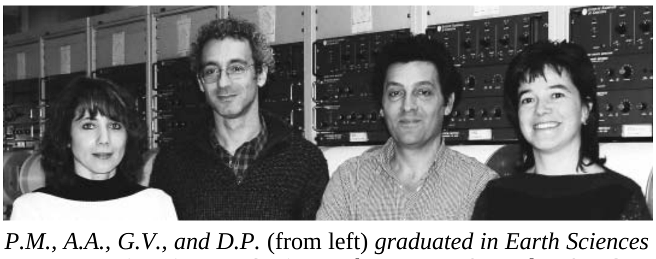

P.M., A.A., G.V., and D.P. (from left) *graduated in Earth Sciences at Rome University "La Sapienza" between 1982 and 1986. G.V. and A.A. have been with Istituto Nazionale di Geofisica since 1983. P.M and D.P. joined it in 1992 and 1987, respectively. They share a great interest in the active tectonics and seismogenic processes of the Italian region, and have given special emphasis to the geometrical and mechanical characterization of seismogenic sources in their respective fields of expertise. G.V. and D.P. focused specifically on historical and prehistorical seismicity, on the quantification of the main earthquake sequences of the XX century and on developing tools for investigating hidden and blind faults. A.A. and P.M. concentrated on Italy's present stress field from earthquake and borehole data, on intermediate and deep earthquakes of the Italian subduction-collision zones and on the structure of the lithosphere beneath the peninsula.*

by Piergiorgio Malesani[1], Elena Pecchioni[1], Emma Cantisani[1], and Fabio Fratini[2]

Geolithology and provenance of materials of some historical buildings and monuments in the centre of Florence (Italy)

1 Earth Sciences Department - University of Florence, Via G. La Pira, 4, 50121 Florence, Italy.
2 C.N.R.- I.C.V.C.B., Via degli Alfani, 74, 50121 Florence, Italy.

Two very important historical areas have been considered in this work. Piazza della Signoria with the Palazzo Vecchio and many other outstanding buildings, is a typical example of the use of sandstone materials characteristic of the Florentine area; Piazza del Duomo with the Cathedral of Santa Maria del Fiore standing in the middle is, on the other hand, an example of other equally typical lithotypes such as the "white" marble and the "red" and "green" "marbles" that decorate the Cathedral. A detailed study as well as the relief and the mapping of the historical buildings in Piazza della Signoria and of the Cathedral of Santa Maria del Fiore, have led to the description of the materials that make them up as well as to the establishment of their quarry provenance.

Historical introduction to the Florentine centre

Buildings in the centre of Florence range in time from the Middle Ages to the Renaissance up to the nineteenth century and record the history of this town.

Florence started expanding and increasing in importance at the end of the 12th century with the construction of new city walls, reaching the level of other large European cities in the 13th century . As its population increased, the Florentines began more and more to inhabit larger public buildings which were often built leaning one against the other due to a lack of space. The wealthy higher social classes exhibited their economic power by having tall towers built where they could seek protection from the violence of private revenges and from the struggle between factions. Churches were generally built in isolated places.

In Dante's times a new political order led to other changes: first of all, towers were demolished and new public buildings were built worthy of a own centre that had become one of the largest and richest in Europe. The *Palazzo Vecchio* was built as well as the *Loggia dei Priori* (today known as the *Loggia dei Lanzi*) and many other mansions belonging to the emerging artisan middle class. On September 8, 1296, the first stone was set for the *Cathedral of Santa Maria del Fiore* which was not dedicated and structurally completed until 1436. The *Orsanmichele* rose (entirely built in stone) destined to the wheat market became one of the most beautiful churches in the historical centre after 1350.

Although in the late 14th century many buildings had bestowed a rather uniform, anonymous appearance on the city with their flat even façades. During the 15th century, the construction of privately owned buildings and original mansions started to give the town a more diversified aspect. The façades of the latter generally reached the third floor, were symmetrical and concentrated on a single entrance, leading to the inside; their covering was stone carved such as ashlar or plaster; these two types of styles were often used together to point out the differences between the storeys of the building. These mansions stood along the narrow streets of the medieval town and almost none of them faced the already existing squares.

At the beginning of the 16th century, the concentration of buildings in some streets had increased to the point of changing the morphology of the quarter.

From the 12th century the main stone materials used in Florentine architecture were two sandstones: *Pietraforte* and *Pietra Seren*a; while various types of marbles, serpentinites, bricks and plasters were also used to reach special chromatic effects. *Pietra Serena* was mainly used for decorative purposes, while *Pietraforte* was mostly used during the Middle Ages in the construction of the bearing structures of buildings and as a sheathing during the Renaissance (AA. VV., 1993, Bargellini et al., 1970).

Piazza della Signoria

Piazza della Signoria, one of the most beautiful squares in Florence, hosts the *Palazzo Vecchio*, the seat of the present City Hall. The area covered by the square has very ancient roots, in fact excavations made in 1974 and more recently, when the square was repaved between 1982 and l989, have even revealed important original remains of the former imperial Roman city that lay in its place. The original square was modified during the medieval period and later between the 11th and 14th centuries, when many towers and houses were demolished after the Guelf victory against the Ghibelline faction; at the end of the 13th century, the *Palazzo dei Priori* (now the *Palazzo Vecchio*) was built by Arnolfo di Cambio. In 1342, the Duke of Athens gave a new arrangement to the square ordering the demolition of some buildings to create an open space in front of the *Palazzo Vecchio*. During the florentine republic, the *Marzocco* and the *Giuditta* sculptures were introduced by Donatello and the *David* by Michelangiolo (Allegri and Cecchi, 1980). During the Medici dynasty, important decorative changes were made to the square such as the *Fonte di Nettuno* by Bartolommeo Ammannati and the *Monumento Equestre a Cosimo I* by Giambologna.

Materials making up the buildings and the pavement of Piazza della Signoria

The materials mainly used in Florentine architecture are those that can most easily be found in the surroundings of the city such as the Pietra Serena sandstone, cropping out in the hills near Fiesole, north of the city and Pietraforte which are found in abundance in the hills south of the city.

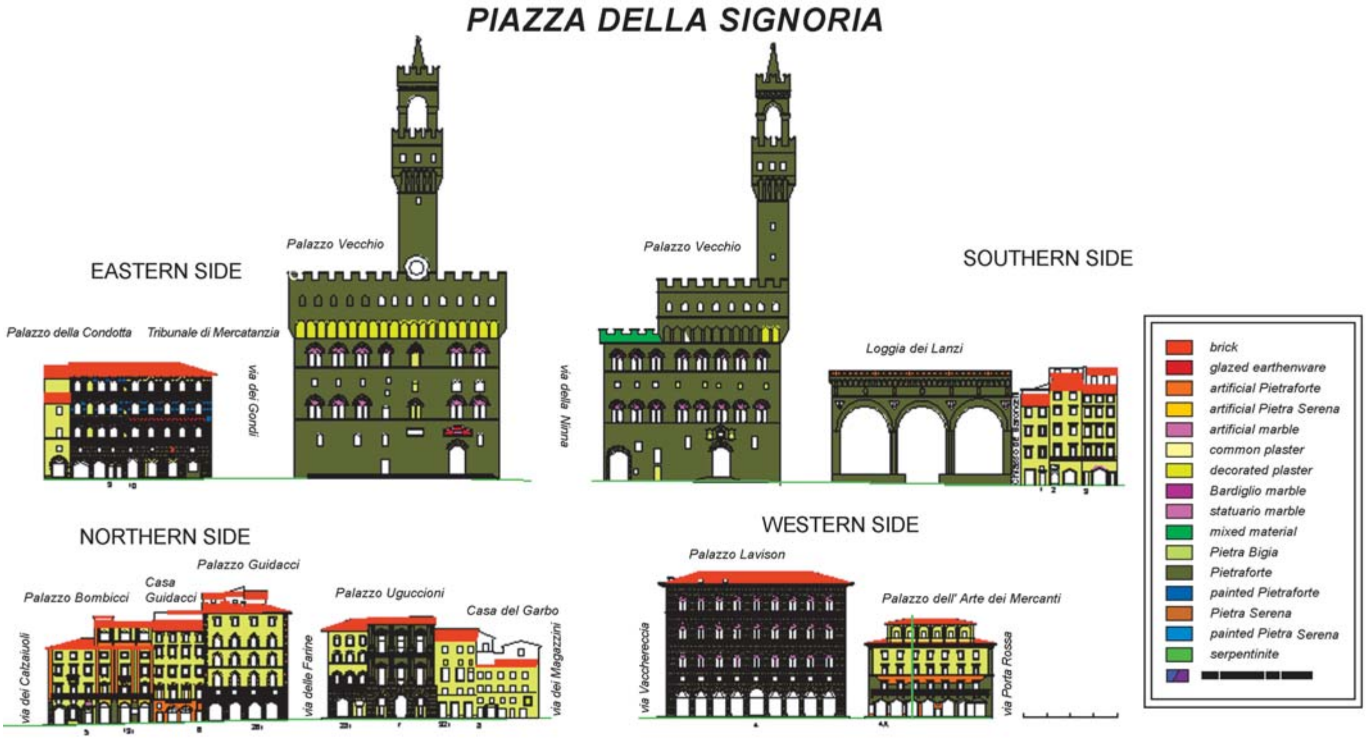

Figure 1 Prospect of the Piazza Signoria: view of the materials making up the buildings.

The mansions standing in the *Piazza della Signoria* (Bargellini and Guarnieri, 1985–87) and especially those of medieval age consist mainly of Pietraforte with portions decorated with Pietra Serena. The most deteriorated parts have either been replaced during the centuries or rebuilt using mortars that imitate the stone (artificial Pietraforte or artificial Pietra Serena) or using decorated plasters. The square also shows good examples of these types of operations. The buildings of lesser importance in the square are covered instead with common plasters.

Starting from the eastern part of the square in a clockwise direction, the first building one sees is the *Palazzo della Condotta* (13th–14th century), built in Pietraforte, painted Pietraforte in the upper part, with common plaster all over the left side and brick on the roof and in the coats of arms. Close to it is the *Palazzo del Tribunale di Mercatanzia* (dating from the second half of 14th century), built exactly as the former one in Pietraforte, painted Pietraforte, common plaster and brick. Next, the *Palazzo Vecchio* (13th–14th century) (Cecchi, 1995) present City Hall and the seat of the *Palazzo Vecchio* Museum and a laboratory for the restoration of tapestries of the *Opificio delle Pietre Dure*. It was built by Arnolfo di Cambio with a façade consisting almost entirely of Pietraforte. Then one sees the *Loggia dei Lanzi* (14th century), an open air gallery completed by the architects Benci di Cioni, Simone Talenti and Taddeo Ristori and entirely built in Pietraforte, except for the balustrade on the roof made of Pietra Serena and the coats of arms which are in glazed earthenware and marble. The houses standing in a row (14th century) presently used as restaurants, dwellings or offices are mostly covered with common plaster and artificial Pietra Serena; Pietraforte is present along the balconies and windows and marble decorations are on the arch of the main door. Following is the *Palazzo Lavison* (1869–1871), the present seat of the Assicurazioni Generali, shops and coffee shops, built by the architect Giuseppe Landi, is entirely in Pietraforte with statuario marble decorations around the mullioned windows. The building of the *Arte dei Mercanti* (14th–19th century), nowadays hosting shops and offices, has undergone many adaptations throughout the centuries. It was covered with common plaster in the higher part of its façade and with Pietraforte and Pietra Serena in the lower part; artificial Pietraforte and artificial Pietra Serena are present in the decorative elements. *Palazzo Bombicci* (18th century), nowadays hosting the Cassa di Risparmio di Firenze and the Alberto della Ragione Art Collection as well as a café, attributed to Bernardo Fallani, shows common plaster and Pietraforte on the façade, artificial Pietraforte, artificial Pietra Serena, Pietra Serena and statuario marble in its decorative elements. The *Casa Guidacci* (built in the 13th century and later remodelled) nowadays hosts the seat of the Banca Toscana and private homes. It has common plaster on the façade and artificial Pietraforte as well as parts in Pietra Serena and artificial Pietra Serena. The *Palazzo Guidacci* (13th–19th century), nowadays the location of the Banco Ambrosiano Veneto and private homes, has common plaster and Pietraforte, with elements in artificial Pietra Serena and artificial Pietraforte. The building *Canto dei Giugni*, now used for dwellings and restaurants, consists of common plaster in the upper part of the façade and Pietraforte in the lower part, with ornamental elements in Pietra Serena and Pietra Bigia. *Palazzo Uguccioni* (16th century), attributed to Mariotto di Zanobi Folfi, was built entirely in Pietraforte with Pietra Serena and common plaster parts. *Casa del Garbo* (12th–19th century), at present used as homes and restaurants, shows a façade covered with common plaster with decorations in Pietra Serena and plaster (Bellucci, 2001; Jacorossi, 1972; Profeti, l999) (Figures 1, 2).

Figure 2 Piazza della Signoria: a particular of a building decoration in Pietra Serena and ashlars in Pietraforte.

The pavement of the square has consisted of different materials and colours throughout the centuries, starting from the original earthen soil to brickwork, to the Pietraforte and Pietra Serena of the 18th century, to slabs in great part replaced with Pietra di Firenzuola in the eighties (20th century).

Palazzo Vecchio

The building planned by Arnolfo di Cambio was built in 1299 as a new seat for the *Priori delle Arti* of the Florentine Republic (Bargellini, 1968).

The building stands on the remains of the preceding structures dating back to the Roman age (theatre whose cellars were used as prisons) as well as to medieval times: the houses of the Uberti family, defeated in 1258, used to stand here. The original core, in the shape of a cube, was initially completed in 1302 and fully accomplished in 1315. The building was constructed as a fortress to resist and repel enemie assaults as can be seen by the thick ashlar walls and the high ironed windows of the ground floors. The building was divided into three storeys by thin frames. It presents a double row of mullioned round arched windows in gothic style. The crowning has a jutting-out gallery with covered and uncovered communicating corridors protected by strong merlons and held by brackets linked to round arches. The polychrome coats of arms of the Florentine Republic were painted under small arches. On the right stands the 94 metres high tower, built on top of the pre-existing Foraboschi or della Vacca tower, where Cosimo il Vecchio and Girolamo Savonarola were imprisoned. In the 15th century, it became the seat of the Signoria. In 1540, it was the mansion of the dukes until Cosimo I moved his court to the *Palazzo Pitti* and the building was given the name of *Palazzo Vecchio*. From 1865 to 1871, it housed the Italian Parliament when Florence became the capital of the newly united Italian Kingdom. Now it is the seat of the City Hall.

The main façade of the western side of the palace is mostly built in Pietraforte carved as little ashlars with elements in common plaster and corbels in Pietra Serena. The covered communicating corridor of the gallery is made of decorated plaster, the mullioned windows are of white statuario marble, and the decorating elements above the main entrance door are of glazed earthenware. The façade on the northern side standing on Via dei Gondi is mainly built in Pietraforte with elements in common and decorated plaster; the mullioned windows are in some cases in statuario marble or in plaster which imitates marble. The battlement placed below is built in a mixed material consisting of limestone, earthenware, Pietraforte and Pietra Serena.

Pietraforte is a sandstone belonging to the turbiditic formation present in the allocthonous complex of the External Liguridi superposed on the Tuscan Series. The formation dates back to the Upper Cretaceous (Abbate and Bruni, 1987; Bortolotti, 1962). Between Florence and Civitavecchia, it crops out in three areas (Florence, Amiata and Tolfa mountains), presenting different thicknesses. In the neighbourhood of Florence, Pietraforte consists of great lenses of turbiditic materials mostly at the base of the Sillano Formation. Thickness varies from 100 m to about 800 m. From a lithological point of view it is a regular alternation of grey sandstone layers and argillites (hemipelagites) with rare intercalations of marls and marly limestones.

Petrographically, Pietraforte is a fine-grained 150 μm lithic sandstone made in the same proportion by silicatic grains (quartz, feldspars and magmatic fragments) and carbonatic grains (dolostones). The grains are bounded by a mainly calcitic matrix that makes the rock particularly strong. Its porosity is 4–6%. The mechanical parameters are higher than those of Pietra Serena thanks to the greater compactness established in the stone by the recrystallization caused by diagenesis; the maximum stress determined perpendicular to bedding is of about 140 MPa (Cipriani and Malesani, 1966; Bruni et al., 1994).

In fresh cut the sandstone has a grey colour, but is easily oxidised acquiring an ochre brown colour on buildings. Such change in colour, due to the oxidation of iron, proceeds very quickly from the surface to the inside, without cohesion decrease.

Pietraforte is often characterized by laminations of a convolute type and by the presence of calcite veins which may stand for preferential areas of block detachment.

The most ancient quarries of this material were on the left side of the Arno river (e.g. Piazza Santa Felicita, Costa San Giorgio, Boboli, etc.). These quarries were later incorporated by the urban expansion and new ones were opened further south (Monteripaldi, Florence).

Nowadays the stone is quarried in the surroundings of Greve and in the quarry of Riscaggio, east of Florence.

Pietra Serena belongs to the sandstones of the Macigno Formation. It consists of beds of turbiditic sandstones separated by pelitic levels which are the finest components of single turbidity current. The Macigno which constitutes the upper part of the Tuscan Nappe, can be dated to the late Oligocene and stratigraphically overlies the Scisti Policromi. It has the greatest thickness along the alignment of Mt. Orsaro-Chianti Mts. where it can reach 3,000 m.

Petrographically, this sandstone can be defined as a medium-coarse-grained greywacke made by quartz, feldspars, micas, fragments of metamorphic and magmatic rocks. The matrix is quite abundant and is made by illite, kaolinite and chlorite-vermiculite (present only in some outcropping areas) (Cipriani and Malesani, 1963).

The grain size of the material used in architecture varies from medium fine to medium, with values around 250–300 μm. It shows a porosity varying from 4 to 6%. It has a very high imbibition capacity and its saturation index can reach 80%.

The mechanical parameters are relatively poor, the maximum stress in the best qualities, determined perpendicularly to bedding, is of about 70 MPa.

Pietra Serena has a bluish-grey colour in fresh cut, while under strong alteration, owing to the iron oxidation, it takes on an ochraceous colour different however from that of Pietraforte.

Pietra Serena seldom shows sedimentary structures such as parallel laminations. In addition, the veins of spathic calcite, due to the filling of pre-existing fractures, making up preferential detachment surfaces, are rather rare. It is owing to the rarity of these sedimentary structures and to the high workability that this stone is one of the materials most widely used in Florentine architecture for decoration purposes.

The most ancient quarries of Pietra Serena were probably located even in Roman-Etruscan times in the Fiesole ridge. Periods of greater development of quarrying of this stone can be dated between the 13th–15th centuries, both bound to the urban growth of the city. In the 13th century, the quarrying activity was carried out in areas around Fiesole (Valle del Mugnone, Monte Ceceri, Vincigliata, Settignano); while in the 15th century new quarries were opened, for example, to the west (Gonfolina and Carmignano) and to the south of Florence (Tavarnuzze), to meet the great demand for this material during the Renaissance. Nowadays the quarrying of this stone around Florence has almost ceased, except for the quarry at Caprolo near Greve, 25 km south of Florence and at Ponte di Mezzo near Lastra a Signa, about 20 km west of Florence (Banchelli et al., 1997).

Historical introduction to the Cathedral of Santa Maria del Fiore

From the laying of the first stone of the Cathedral of *Santa Maria del Fiore* on September 8, 1296 to its dedication and structural completion in 1436, famous architects such as Arnolfo di Cambio, Giotto, Francesco Talenti and Filippo Brunelleschi, just to mention a few, were involved in its construction one after another leaving enduring marks of their architectural expressions. The external decoration of the Cathedral was entrusted to Arnolfo di Cambio, who created the first project. Arnolfo carried out a decoration with a trichromy obtained using "white", "red" and "green" Tuscan "marbles", which recalled on one side the mosaics of the Roman cloisters of *San Paolo in Laterano* and on the other the Florentine tradition that had significant examples in the Battistero and in the façade of *San Miniato al Monte*. When Arnolfo died (March, 1301), all the walls had already been put up and the marble decoration, so well appreciated by those who carried on his work, was started and extended to the whole building. The setting up of the marble faces

Figure 3 A panoramic view of the Florence Cathedral.

went on with the construction of the walls. This aspect is indirectly confirmed by the resolution of November 19, 1367 in which the setting up of the marble faces was to be postponed to give priority to the building of the walls (Figure 3).

The dedication of the Cathedral, completed in its structure, took place in 1436 when many parts were still unfinished, such as the lantern, completed in 1470, after Brunelleschi's death. The façade planned by Emilio De Fabris was inaugurated in 1887 while the sheathing of the tambour of the Dome was still unfinished.

Materials used in the original construction and for the façade

Information on the materials used in the originally setting up of the Cathedral can be drawn from many archive documents kept at the Library of *Opera del Duomo* in Florence.

Cesare Guasti (1887) in his "*Santa Maria del Fiore—La costruzione della Chiesa e del Campanile secondo i documenti tratti dall'archivio dell'Opera Secolare e da quelli di Stato*" relates the resolution of the Opera dated January 5, 1350 confirming the lithotypes making up the trichromy of the Cathedral.

The entire bibliography leads to the conclusion that the "red", marly limestones of San Giusto di Monterantoli (Cintoia in Chianti mountain) were used for the original construction of the Cathedral, as well as marly limestones from Monsummano in Valdinievole (Pistoia), Carrara marbles for the "white", and serpentinite from Figline di Prato (Prato) for the "green".

There is a difference in the case of the façade which was built later: the "green" of serpentinites coming from Figline di Prato was used along with the "red" coming from several quarries, above all in the Grosseto area, at Villa Collemandina (Garfagnana-Lucca), at Monsummano and in the Perugia area; while for the "white" after several visits to Carrara and Serravezza an agreement was reached with G.B. Sancholle Henraux who provided the marble from his quarries at Serravezza (Lucca) at cost price (AA.VV., 1987) (Figures 4a, b).

The "red"

A preliminary microscopic exam of the "red" used in the side faces of the original construction of the Cathedral showed the presence of two different typologies. In the literature and archive documents, there wasn't unanimous agreement on the provenance of the material, but several works of research carried out in recent years (Sartori, 1996; Vannucci et al., 1997; Fazzuoli et al., 1998; Sartori, 1998) have led to an exact lithological identification. The "reds" have been therefore attributed to the Marls Formation of Sugame of the Tuscan Series cropping out at S. Giusto a Monterantoli (Cintoia in Chianti mountain) and to the Marls Formation of Sugame cropping out at Monsummano (Pistoia). Research carried out both on the "reds" present in the side façades of *Santa Maria del Fiore* and on *Santa Maria Novella*, as well as on the columns of the *Grotta Grande* in the *Boboli* Gardens, have led to the determination of a series of mineralogical and geochemical parameters able to identify the lithotype employed and, in some cases, the sites of supplying.

It must be also specified that the main lithotypes of the Marls Formation of Sugame, cropping out at Cintoia, consist of violet-reddish, liver-red, greyish pink, light olive grey marly limestones, sometimes not stratified, with a scaly fracturing. The same litofacies of the Marls Formation of Sugame cropping out at Monsummano, having a higher thickness than that of Cintoia, is almost entirely non-stratified.

Moreover, the "reds" from Monsummano show two different lithologies. The first consists of liver-reddish marly calcilutites with frequent bioturbations and fractures filled with calcite. Their appearance is very close to that of the marly limestones from Cintoia and is in agreement with Agostino Del Riccio's description (Del Riccio, 1597). In the second case, they look like nodular limestones as they have more or less amygdaloid forms made by lighter micritic limestone surrounded by a redder marly limestone. This has a consequence of boundinage phenomena due to plastic imperfections in the originary deposit (Fazzuoli and Maestrelli Manetti, 1973).

These lithotypes are characterized by a content in calcite that goes from 79 to 87% and a content in quartz from 6 to 13% often prevailing on the phyllosilicates; the feldspars are represented only by plagioclases.

Among geochemical parameters the content in barium appears, at least in most of the samples, quite discriminating according to the provenance. At Cintoia the average content is around 1100 ppm against an average content of about 140 ppm in Monsummano.

The association of clay minerals that characterize the two lithotypes (both at Cintoia and at Monsummano) consists, in order of abundance, of illite, kaolinite, chlorite and chlorite-vermiculite.

As concerns physical characteristics, porosity ranges between 2 and 3.5%. The absorption in water is low with a saturation index generally lower than 35%.

As to the façade several types of "reds" have been noticed: nodular limestones rich in ammonites coming from several Tuscan areas (Gerfalco, near Grosseto and Monsummano near Pistoia) and marly micritic limestones coming from Montieri (Grosseto), Villa Collemandina (Garfagnana-Lucca) and Monte Malbe-Perugia (Red Scaglia Formation in the Umbria Series).

The nodular limestone, rich in ammonites, appears as micrite whose nodules are delimited by the presence of discontinuities of irregular course filled by iron oxides. Micritic limestones turn out to be rich in calcite veins characterized by a greater content in oxides present both in the mass and concentrated in thin levels.

The "white"

The white of the original construction of the cathedral sides is marble coming mostly from the Apuan Alps as well as the Montagnola Senese.

Apuan marbles formed during the Oligocene-lower Miocene (27–10 m.y.) at temperatures of 350˚–450˚C and pressures of 5–6 kbar (Di Pisa et al., 1985; Schultz, 1996; Franceschelli et al., 1997). Several varieties have been identified among the marbles, some deriving directly from the original sedimentary lithologies preserving some of their texture patterns (common white marble used in the Cathedral's faces), others originating through pliable interdigitation between two different lithologies, such as Arabescato and Nuvolato varieties.

Common white marble can present several typologies; from white (statuary type) to more or less veined. Compositionally it consists of 99% cal-

a

b

Figures 4a,b A detail of the trichromy of "marbles" in the façade.

cite and traces of quartz, albite, muscovite and pyrite. Common marble can also show microstructures varying from polygonal granoblastic with rectilinear contacts between grains, to xenoblastic with suturated contacts between grains (Coli, 1989; Molli et al., 1997; Meccheri and Molli, 1996). The size of the grains generally ranges between 200 and 500 µm, while their porosity varies from a minimum of about 1% for the xenoblastic structure to a maximum of 2.5% for the polygonal granoblastic structure (Barsottelli et al., 1998).

Marbles from the Montagnola Senese have colours that go from white at the base of the sequence (the variety present in the Cathedral) to greyish, while at the roof they take over a pink to yellow shade (Giannini and Lazzarotto, 1970; Micheluccini et al., 1981). These marbles formed in the same period as the Apuan marbles, but under much lower temperature conditions. The recrystallization processes were milder and the average size of the grains turns out to be smaller than that of Apuan marbles (50–150 µm).

The marbles of the façade, even if coming from the same quarries as those of the side faces, show a mainly granoblastic structure characterized by the presence of rectilinear contacts between the grains. This has caused a greater decay in time that leads to the need of many replacements.

Instead the marbles of the sides show structural characteristics mostly referable to a xenoblastic microstructure, characterized by contacts between calcite granules, lobated or suturated that can give the blocks a considerable durability.

The "green"

The "green" used in the original construction of the Cathedral is a serpentinite, a lithotype originated from oceanic metamorphism of peridotite (composed of olivine and pyroxenes) with a consequent formation of serpentine (Wick and Wittaker, 1997). These rocks are present, even if in outcroppings of limited extensions, mostly in the Northern Apennine and make up the so called Ofiolitic Complex in which gabbros and basalts are always found associated. Serpentinite is characterized by a considerable mineralogical homogeneity, but the rocks show varieties with very different colours and macro and microscopical structures. Many shades of green can be found on the Cathedral, from light to dark till almost black, sometimes with bluish reflections often spotted with stains.

The macroscopical structure of serpentinite can be relatively uniform, interrupted by small whitish veins of fibrous serpentine (chrysotile) or characterized by a thick interlacement of light veins as in the "ranocchiaia" variety.

Two varieties are distinguished at a microscopical level: the first is characterized by the presence of variable percentages of bastitic texture, mesh texture and hourglass texture. Bastites are the relics of the serpentinization of amphiboles and pyroxenes, whereas mesh texture and hourglass texture are the relics of the serpentinization of olivine. All these textures are constituted by lizardite type serpentine; while the rare veins are made of chrysotile type serpentine. The presence of magnetite in small crystals is widespread; chromiferous spinel and pyrite are also present.

Hourglass texture prevails in the second variety with a small percentage of bastite and the absence of mesh texture. Magnetite is present in grains, which are larger in size than those of the former variety. This determines a light green colour in the rock. The tiny chrysotile veins can be so plentiful as to reach a thick interlacement that characterizes the "ranocchiaia".

Concerning petrophysical characteristics, it has been noticed that the Apennine serpentinites at lizardite+chrysotile, having a porosity ranging between 3 and 10%, are characterized by a saturation index often greater than 100%, which causes strong problems of durability (Fratini et al., 1987, 1991; De Vecchi et al., 1991; Bralia et al., 1995).

As to the façade, the "green" is always the Appeninic serpentinites.

Materials employed in replacements

Most of the replacements and/or restorations were performed on marble used in greater quantity and making up most of the modelling, subject to a stronger decay, unlike the faces that always turn out to be better preserved. Replacements on the "red" were fewer and still fewer on the "green".

It can reasonably be inferred that the quarries furnishing the white marble, the red marly limestones and the serpentinite, when necessary, have remained the same for economical and practical reasons, at least up to the first years of the twentieth century: the marble quarries were in the Apuan Alps, those of marly limestones at San Giusto a Monterantoli and Monsummano and that of serpentinite at Figline di Prato. In recent years, except for the Apuan marble, the "red" and the "green" quarries were closed both to safeguard and protect the environment and to avoid depletion of the materials; other quarries were identified: the "red" has changed with the micritic limestone of Villa Collemandina and the "green" comes from the Alpine area (Val Malenco) (Bianchini, 1999). The "Alp green" of Val Malenco is a serpentinite where the high pressure metamorphism changed the chrysotile and lizardite in antigorite.

The use of marble from the Montagnola Senese has no longer been reported from information gathered at the *Opera del Duomo.*

Conclusions

In this paper geolithology and provenance of the materials of some historical buildings of *Piazza della Signoria* and of the monuments of *Piazza del Duomo* have been considered.

In *Piazza della Signoria* two sandstones are mostly used: Pietraforte, as building material and Pietra Serena in the architectonic elements. Moreover, the use of other materials such as marbles, decorated plasters, glazed earthenware can be seen in the decorative elements. As to provenance of the sandstones, the ancient quarries of Pietraforte were on the left side of the Arno river (south of the city), while the ancient quarries of Pietra Serena were in the Fiesole ridge (north of Florence).

Concerning the Cathedral of *Santa Maria del Fiore* it is necessary to distinguish the lithotypes of the original construction and the more recent façade. The original setting up is constituted by a trichromy: a red limestone, a white marble and a green serpentinite. The "red" is a marly limestone quarried at San Giusto di Monterantoli (Cintoia in Chianti mountain) and at Monsummano in Valdinievole (Pistoia), the "white" is a common Carrara marble quarried in the Apuan Alps and the "green" is a serpentinite from Figline di Prato (Prato).

The same colours are present in the façade, but the "red" is a marly limestone coming from several quarries, above all in the Grosseto area, at Villa Collemandina (Garfagnana-Lucca), at Monsummano (Pistoia) and in the Perugia area; the "white" is a common marble coming from Serravezza (Lucca); while the "green" of serpentinites is the same as the original construction.

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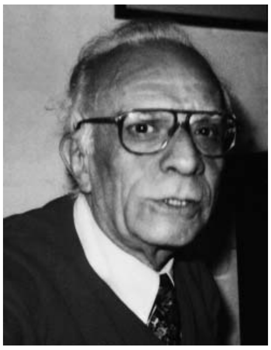
Piergiorgio Malesani, *Full professor of Petrography at the Faculty of Mathematical, Physical and Natural Sciences of the University of Florence since 1976. Member of the Scientific Council of the "Centro di Studio sulle Cause di Deperimento e sui Metodi di Conservazione delle Opere d' Arte-CNR-Firenze". His scientific interest includes Applied Petrology, Conservation of Cultural Heritage, Sedimentary and Magmatic Petrology and Engineering and Regional Geology.*



Elena Pecchioni, *Researcher in Petrology applied to Conservation of Cultural Heritage at Earth Science Department of University of Florence. Specialized in Science applied to Conservation of Cultural Heritage. Her scientific interest are weathering and conservation of natural and artificial stone. She teaches Applied Petrology and Mineralogy at the Restoration School of Opificio delle Pietre Dure of Florence.*

Fabio Fratini, *Researcher in Petrology applied to Conservation of Cultural Heritage at the National Council of Researches (C.N.R.) of Florence. He carries out researches on natural and artificial stone materials, archeometry and conservation.*

Emma Cantisani *Graduated in Applied Petrology in 1999 at the Earth Science Department of Pisa University, attending a PhD Course in Science for the Conservation of Cultural Heritage at Florence University.*

by Giovanni Ferraris and Marco Ciriotti

From anorthite to vesuvianite: an excursion through the minerals first discovered in Italy

Dipartimento di Scienze Mineralogiche e Petrologiche, Università di Torino, and Istituto di Geoscienze e Georisorse, CNR - Via Valperga Caluso 35, 10125 Torino, Italy. <*giovanni.ferraris@unito.it*> <*marco.ciriotti@libero.it*>

With regard to the number of mineral species first discovered in a particular country, Italy ranks fourth, being preceded only by USA, Russia and Germany. Important rock-forming minerals like analcime, anorthite, anorthoclase, antigorite, celadonite, diopside, dolomite, forsterite, humite, leucite, lime, lizardite, magnesiochloritoid, magnesite, nepheline, and sepiolite are among the 240 species first discovered in Italy. Other significant first Italian findings are some important index minerals of the UHP metamorphism (ellenbergerite, phosphoellenbergerite, magnesiodumortierite, and magnesiostaurolite; second occurrence of coesite) which were discovered in the well-known and first reported UHP outcrop of the Dora-Máira massif. Italy is also the native land of about 30 zeolites and of the two most abundant modern species: balangeroite and carlosturanite, which are rock-forming asbestos-like minerals of serpentinites in the western Alps.

Introduction

About 4500 reasonably well described geological chemical compounds and were recognized as species by the scientific community. It should be noted that in the mineralogical literature, thousands of "mineral names" cannot be related to well-defined compounds, but are just synonyms or refer to varieties (de Fourestier, 1999). These names should not be used according to the rules established by the Commission on New Minerals and Mineral Names (CNMMN) of the International Mineralogical Association (IMA) [cf. Nickel and Grice (1998) and the web site of the CNMMN *http://www.geo.vu.nl/users/ima-cnmmn/*]. At the same time, an undefined number (not less than 1000?) of geological chemical compounds are sufficiently described in literature, but they have not been assigned a name; consequently they do not have the status of species. More than 50% (i.e. about 2500) of the mineral species have been officially accepted by the CNMMN since its establishment in 1962 (cf. the quoted web site).

National and regional records

If countries are classified according to the number of mineral species first discovered in their territory, Italy, with its 240 species (at the end of 2002), ranks fourth with about 5.5% of species being preceded by USA (~ 15%), Russia (~ 14%), and Germany (~ 7%). Limiting the list to the countries with at least 100 first discovered species, Italy is followed by Canada (~ 4%), Sweden, UK, Australia, China, France, and Japan. This classification certainly reflects well the activity, both past and present, of these particular countries in the search for new mineral species (sometimes, as amateur rather than a professional activity) than those countries' geochemical potentiality. It should be noted that in recent years well crystallized (or just rare) samples of minerals have became a good business and the leading countries are well known for hosting famous international fairs and trading mineral samples. Of course, the long established participation of a country in the science research is also important. Thus Italy boasts one species (sal ammoniac) described by Agricola in 1546, five other species (analcime, chabazite-Ca, dolomite, leucite, and vesuvianite) described in the XVIII century, and 74 in the XIX century. This group includes mainly species found by Scacchi (cf. Mottana, 2001) in the Vesuvius area. The group of the volcanic species has been further increased by Zambonini (1935). The historical record of Italy could certainly be improved if Plinius had quoted localities for the mineral species he mentioned. About 120 species have been described or redefined in the CNMMN era.

Guinness' records have always a touch of parochialism, consequently let us arrange the mineral species first discovered in Italy according to the Italian historical (now also administrative) regions. Of course, the regions are listed below with their Italian name, according to the number of species first discovered in each region. To give their Guinness' classification its proper scientific value, the chemical nature of minerals is shown as follows: **silicates in heavy type**; OXIDES IN SMALL CAPITALS; sulfides and sulfosalts underlined; *organic compounds in italic*; ***carbonates in heavy italic***; other groups in normal type.

Campania (71): Anorthite, aphthitalite, avogadrite, bassanite, calciobetafite, carobbiite, **chabazite-K**, chlorothionite chalcocyanite, chloraluminite, chlormanganokalite, chlorocalcite, chloromagnesite, **clinohumite**, cotunnite, covellite, cryptohalite, **cuprorivaite, cuspidine**, cyanochroite, **davyne**, dimorphite I, dimorphite II, dolerophanite, eriochalcite, erythrosiderite, euchlorine, ferrohexahydrite, ferruccite, **forsterite, haüyne, humite, kaliophilite**, kremersite, **leucite-*high*, leucite-*low***, LIME, **lititionite**, MAGNESIOFERRITE, malladrite, manganolangbeinite, **marialite**, mascagnite, matteuccite, **meionite**, melanothallite, mercallite, **microsommite**, misenite, mitscherlichite, molysite, **montesommaite, monticellite, *nahcolite*, nepheline**, palmierite, **panunzite**, PERICLASE, picromerite, **potassic-fluorrichterite**, pseudocotunnite, **quadridavyne**, sal ammoniac, **sarcolite**, scacchite, sylvite, TENORITE, **vesuvianite**, voltaite, ZIRCONOLITE-3*O*, ZIRCONOLITE-3*T*.

Toscana (54): Ammonioborite, APUANITE, biringuccite, bonattite, BOTTINOITE, boussingaultite, BRIZZIITE, campigliaite, carraraite, cetineite, CLINOCERVANTITE, coquandite, **dachiardite-Ca**, DESSAUITE-(Y), *dinite*, **elbaite**, fiedlerite-1*A*, **franzinite**, garavellite, ginorite, grattarolaite, grumiplucite, **ilvaite-*M***, **ilvaite-*O***, larderellite, **liottite,** meneghinite, minguzzite, moëloite, mohrite, nasinite, onoratoite, pellouxite, peretaite, pillaite, **pitiglianoite, pollucite**, riomarinaite, rodolicoite, rosenbergite, ROSIAITE, **rubicline**, santabarbaraite, santite, sassolite, sborgite, scainiite, *simonellite*, STIBIVANITE-2*O*, **tuscanite**, URANOPOLYCRASE, VERSILIAITE, zaccagnaite, zincalstibite.

Piemonte (29): Antigorite, balangeroite, bavenite, bazzite, ***calcioancylite-(Nd)***, canavesite, **carlosturanite, cascandite, cervandonite-(Ce), diopside, ellenbergerite**, FETIASITE, gasparite-(Ce), GRAMACCIOLIITE-(Y), **jervisite**, **magnesiodumortierite, magnesiostaurolite,** ***magnesite,*** monazite-(Nd), **paracelsian**, paraniite-(Y), phosphoellenbergerite, **roggianite, scandiobabingtonite, sepiolite**, STRÜVERITE, **taramellite**, VIGEZZITE, **wenkite.**
Sicilia (19): Analcime-1*C*, analcime-1*M*, analcime-1*O*, analcime-1*Q*, anorthoclase-*high*, anorthoclase-*low*, barberiite, cannizzarite, **chabazite-Na, fluoro-edenite**, hieratite, magnesio-aubertite, MELANOPHLOGITE-*C*, MELANOPHLOGITE-*Q*, millosevichite, mozgovaite, **phillipsite-Na**, siderazot, sulfur-β.
Lazio (18): Alunite, cesanite, **ciprianiite, gismondine-Ca, giuseppettite, hellandite-(Ce), katoite, latiumite, merlinoite, mottanaite-(Ce)**, peprossiite-(Ce), **perrierite-(Ce)**, **phillipsite-K, sacrofanite, spadaite, stoppaniite, vertumnite, vicanite-(Ce)**.
Liguria (14): Brewsterite-Ba, *caoxite*, cavoite, **cerchiaraite**, GRAVEGLIAITE, **heulandite-Sr, medaite, mozartite**, palenzonaite, reppiaite, **saneroite, strontiopiemontite, tiragalloite, vanadomalayaite**.
Veneto (9): Celadonite, ***cerussite*****, gmelinite-Ca, gmelinite-K, gmelinite-Na, heulandite-K, johannsenite, lizardite-2*H*$_1$, pectolite.**
Sardegna (6): Barrerite, MONTEPONITE, ORLANDIITE, ***rosasite***, sabelliite, **stilbite-Na.**
Trentino-Alto Adige (5): Chabazite-Ca, dachiardite-Na, ***dolomite*****, gehlenite, pectolite.**
Emilia-Romagna (4): Aliettite, jamborite, **lizardite-2*H*$_2$**, troilite.
Lombardia (4): ***Artinite, brugnatellite*****, chiavennite**, sigismundite.
Val d'Aosta (4): Magnesiochloritoid, piemontite, ROMÉITE, STRONTIOMELANE.
Abruzzi (1): *Refikite.*
Friuli-Venezia Giulia (1): Bianchite.
Puglia (1): Francoanellite.
Umbria (1): Willhendersonite.

From anorthite to vesuvianite

To mention both anorthite and vesuvianite in the title of this article seems to involve both a journalistic touch (geological relevance of anorthite ... ; nearly first and last Italian species in alphabetical order ...) and a recognition of Somma-Vesuvius volcanic complex, which is the most prolific Italian area for mineral species. It is unnecessary, of course, at this point to restate the role of plagioclases, including anorthite, in Earth Sciences, not to mention the difficulties students and researchers meet in characterising them exactly for the presence of twinning, exsolutions, etc. Presumably, the etymology of anorthite (from the Greek word for "oblique") applies not only to its morphology but also to its possibility of slanting from a correct interpretation.

Geological records

All mineral species have a scientific meaning in the sense that their occurrence, often in very limited amounts and in very restricted areas, is related to specific genetic and geochemical situations. Clearly, however, only a few of the mineral species reported above (but this can be extended to all known species) have wide prominence, not necessarily economic but at least relevant to the Earth Sciences.

Dolomite—A mountain-forming mineral

Thanks also to an earlier geological exploration of its territory, Italy is recognised as the motherland of some important rock-forming minerals like analcime, anorthite, anorthoclase, antigorite, celadonite, diopside, dolomite, forsterite, humite, leucite, lime, lizardite, magnesiochloritoid, magnesite, nepheline, and sepiolite. In particular, dolomite is a mountain-forming mineral!

Index minerals of the UHP metamorphism

Most of, if not all, the evidence for the attainment of UHP conditions by metamorphic rocks is a mineralogical one (cf. Chopin and Ferraris, 2003). In fact, some minerals by their nature, composition, texture or reactions, may be specific of such conditions. The coesite reported by Chopin (1984) from an outcrop of the Dora-Máira massif (western Alps) was already the second occurrence in the world but, correctly interpreted, opened a new era for modern Petrology and Earth Sciences in general. In fact, the presence of relics of a silica phase which is stable only at a pressure higher than about 3 GPa showed that Dora-Máira outcrop had been uplifted from a depth of about 100 km.

Coesite is not the only mineral which has contribution to telling yet unknown geological history of what is now among the most famous outcrops in the world. The first discovery of ellenbergerite, phosphoellenbergerite, magnesiodumortierite, and magnesiostaurolite; the occurrence of an almost Mg-pure magnesiochloritoid, the most OH-rich wagnerite, and giant practically pure pyrope crystals; the second occurrence of bearthite, together with that of coesite. All these facts represent a proud record for the Italian Mineralogical Sciences, in general, and for the Dora-Máira massif in particular.

Abundant minerals

With very few exceptions, all new mineral species are found in small to very small quantities and they can be characterized only because powerful microanalytical methods are available to modern mineralogy. The adjective "abundant" can only be applied to four of the approximate total of 2500 modern species without risk of an overly subjective evaluation. Two of them are Italian minerals: carlosturanite and balangeroite (the other two are minrecordite and defernite, both from Namibia). These species were originally found in localities about 80 km apart in Piemonte. Both minerals are fibrous and always mistaken for chrysotile (Compagnoni et al., 1983, 1985).

In addition to the type locality of the Auriol asbestos mine, Sampeyre, Val Varaita, carlosturanite has been found in Taberg, Sweden, and abundantly in several fractured serpentinite outcrops in the western Alps. It is certainly the most abundant mineral approved by the CNMMN. Balangeroite has been found at the type locality of the Balangero asbestos mine (the largest one in western Europe) and in a few other localities in the same Lanzo Valley. The mineral occurs as tufts of brown xyloid fibers, more or less rigid, up to several cm in length. It was known as "metaxite" or "xilotile" by the miners from the Balangero mine. Neither massive serpentinites (where balangeroite is formed during the metamorphism of the ultramafic body) nor schistose serpentinites with short-fibre asbestos veins, were industrially exploited. They generally constituted the gangue material normally utilized for the mine constructions, and as granulates for road- and rail-beds, with the surplus dumped as waste. On the base of old miners' stories and data from the Amiantifera di Balangero S.p.A. historical archives, and considering the localisation in the deposit as well as the ubiquitous mingling with long-fibre chrysotile (50%), it is probable that thousands of tons of balangeroite have been exploited together with chrysotile.

Scandium minerals

Baveno (Piemonte) is a well-known mineralogical locality of REE and rare mineral species such as: gadolinite-(Ce), gadolinite-(Y), hingganite-(Y), gugiaite, kaynosite-(Y), and calcioancylite-(Nd). In particular, Baveno is the type locality of bazzite the first discovered scandium mineral (Artini, 1915). Recently three other scandium minerals (jervisite, cascandite and scandiobabingtonite) were first reported from Baveno (Mellini et al., 1982; Orlandi et al., 1998). Thus, this locality accounts for 4 of the 14 known scandium minerals.

Mineral species and modern technology

Zeolites are important minerals for modern technology because of their microporous activity (molecular sieves; cf. Gottardi and Galli, 1985). 30 (analcime-1*C*, analcime-1*M*, analcime-1*O*, analcime-1*Q*, barrerite, brewsterite-Ba, brewsterite-Ca, chabazite-Ca, chabazite-K, chabazite-Na, chiavennite, dachiardite-Ca, gismondine-Ca, gmelinite-Ca, gmelinite-K, gmelinite-Na, heulandite-Ca, heulandite-K, heulandite-Sr, leucite-*high*, leucite-*low*, montesommaite, merlinoite, pectolite, phillipsite-K, phillipsite-Na, roggianite, stilbite-Na, wenkite, and willhendersonite) out of about 90 known zeolites (Coombs et al., 1998) have their type locality in Italy.

Elbaite (tourmaline group) is worth mentioning for the exploitation of its piezoelectric effect. The sulfate palmierite gives its name to a structure-type module which alternates with perovskite-type modules in synthetic compounds with important physical properties, like superconductivity. Cotunnite is a type structure which is very popular in high-pressure solid-state research; recently a TiO_2 phase with cotunnite-type structure and hardness comparable with that of diamond has been synthesized at about 60 GPa and 1000 K. Palmierite and cotunnite (both from Vesuvius), are not the only cases which demonstrate the precursor role of Mineralogical Crystallography in solid state science. Practically all type crystalline structures refer to a mineral species: from perovskite and spinel to diamond, rutile, and corundum, ...

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Giovanni Ferraris *has been professor of Crystallography at the Faculty of Sciences of Torino University since 1975. Currently he is Vice-President of the Commission on New Minerals and Mineral Names of IMA and President of the Commission on Inorganic and Mineral Structures of IUCr. Born (1937) in Prarolo (Italy), he graduated in Physics (Torino University, 1960) and received the Libera Docenza in Crystallography (1969). Author of some 190 scientific papers and contributions to books, collaborated in the description of 29 new mineral species. Plinius medallist of the Italian Mineralogical and Petrologic Society and recipient of the Accademia dei Lincei Tartufari prize.*

Marco Ciriotti *is an amateur mineralogist, a "grouper", and a systematic collector. Born in Calosso (Italy, 1945), he graduated in Natural Sciences while working in industry . He pursued his career in industry until 2000 when, as General Manager, he retired. He then devoted himself to his primary interest and passion: mineral collecting and studies. He is the founder and the coordinator of the recently constituted AMI (Italian Micromineralogical Association), fellow of ICDD, SIMP, EcaSig5 and mineralogical associations. Author of articles on topological and structural mineralogy and of an unpublished mineral classification.*

by Domenico Rio[1], Isabella Premoli Silva[2], and Luca Capraro[1]

The Geologic Time Scale and the Italian stratigraphic record

1 Dipartimento di Geologia, Paleontologia e Geofisica, Università di Padova, Via Giotto 1, I-35137 Padova, Italy.
2 Dipartimento di Scienze della Terra, Università di Milano, Via Mangiagalli 34, I-20133 Milano, Italy.

The construction of the Geologic Time Scale (GTS) is a titanic scientific challenge that has been under way for two centuries and will require much dedicated effort in the future. Italy preserves a paramount stratigraphic record of Mesozoic and Cenozoic marine sediments that have been significant in the development of the modern GTS. The Italian stratigraphic record has been historically important in introducing and defining the standard Chronostratigraphic Units (CUs) of the Neogene and Quaternary. Pelagic successions from Northern Apennines and Southern Alps have been used in the seventies for integrating the late Cretaceous-Paleogene Geomagnetic Polarity Time Scale (GPTS) with planktonic microfossil biostratigraphy and standard CUs. This was a major contribution to the construction of a new generation of GTS based on integrated magnetobiochronology. The middle Miocene to early Pleistocene marine record from Sicily and southern Italy has been fundamental for establishing the recently developed Astronomical Time Scale (ATS). In prospect, there are many potentials still to be exploited in the Italian marine stratigraphic record for implementing the GTS by defining GSSPs of various CUs, improving magnetobiochronology and extending downwards the ATS.

Introduction

The central role of time in Earth sciences can hardly be overstated. We need time as a relative age vernier (Chronostratigraphy) for unraveling and chronicling the Earth's and life's evolution. We need absolute time (Geochronology) if we want to understand and model processes like rates of tectonic deformation or organic evolution or the loops of the global climatic system. Indeed, geoscientists have been confronting with the problem of time since the foundations of the discipline, and we like to think that the achievement of the concept of deep geologic time, well beyond the biblical chronology, is a major contribution to the modern *weltanschaung* (Gould, 1990). Actually, the "discovery" of deep geologic time, although badly needed and perceived by geologists and paleobiologists like Darwin, has been definitively achieved only in the 20th century with the discover of radioactivity. The birth of Geology in the first half of the 19th century and successive developments are strictly interwoven with the construction of a Geologic Time Scale (GTS), which represents the standard and common language for Earth scientists for achieving most of their goals. The modern GTS in essence is composed of two distinct scales: the relative time scale (Chronostratigraphic Units, CUs) and a chronometrical or so-called absolute time scale (Geochronologic Units, GUs). Its construction has been and still is fraught with many obstacles, which include the limits of the stratigraphic data base; the accuracy, precision and reliability of correlation and dating tools; and, last but not least, problems of nomenclature, procedures and stratigraphic philosophy. However, the GTS is an ongoing enterprise because our need of accuracy and precision in dating and correlation is ever-increasing and, because, even the apparently easily achievable goal of an internationally agreed upon set of standard CUs has not yet been accomplished.

In retrospect, the development of the modern GTS is punctuated by five major breakthroughs that resulted in quantum improvements in our ability to read geologic time:

The origins: the Paleontologic Time Scale. In the first half of the 19th century, the application of two simple paradigms, the principles of superposition of Steno and of faunal succession of Smith, led to construction of the first GTS, establishing the large-scale basic chronostratigraphic subdivisions of the GTS that are still in use. Finer subdivisions (what we now call "stages"), based on megafossil biostratigraphy, were added in the second half of the 19th century, although within a much confused conceptual and terminological framework.

Radiometry: the Geochronologic Time Scale. With the discovery of radioactivity at the end of the 19th century, the way was open to the dating game of the traditional CUs and the foundation of the modern GTS (Holmes, 1960).

Magnetobiochronology: the Geomagnetic Polarity Time Scale (GPTS). Up to the early 1970s, the chronology of the GTS was based only on sparse radiometric dates of CUs. In the 1960s, a major breakthrough was represented by the integration of the GTS with the history of polarity reversals of the geomagnetic field (GPTS), initially based on lava flows and marine magnetic anomalies. The drilling of oceanic sediments, development of detailed multiple biostratigraphies based on planktonic microfossils, and technologic advances in measuring magnetic properties of sediments led to a new generation of GTS with improved chronology based on the calibration of biostratigraphic events via magnetostratigraphy (magnetobiochronology) (e.g. Harland et al., 1982; Snelling, 1985; Berggren et al., 1995).

Cyclostratigraphy: the Astronomical Time Scale (ATS). In the early 1990s, advancements made it possible to accomplish a century-old dream of geologists: to utilize the pervasive cyclicity of the stratigraphic record for deriving a chronology. In particular, in the most recent part of the GTS (the icehouse world of the Pleistocene), it has been proved beyond any reasonable doubt that sedimentary, geochemical and paleontologic cycles are related to Milankovich climate change from variations of Earth's orbital parameters, the chronology of which is well established in the most recent geologic past. An unprecedented tool for dating events in the geologic record became available with the construction of an Astronomical Time Scale (ATS; Shackleton et al., 1990; Hilgen, 1991). The chronology of the ATS is independent of radiometry (Hilgen et al., 1997).

The GSSP concept: procedures, nomenclature and philosophical approach to the GTS have always been contradictory and confused. The need of a uniform terminology and philosophical approach was felt early on and first attempted at the International Geological Congress of Bologna in 1881. However, it has not been until the early 1970s that, thanks to the dedicated effort of Hollis Hedberg, an international agreement has been achieved, although not universally accepted, with the publication of the "stratigraphic guide" (Hedberg, 1976). With reference to the GTS, of utmost importance has been the introduction to the practice of defining each CUs in the rock record by a boundary stratototype section, later refined as Global Stratotype Section and Point (GSSP; Cowie et al., 1986). The GSSP concept is intended to provide a formal and unequivocal definition in the rock stratigraphic record of the CUs of the GTS by means of an international agreement.

Our aim here is to review the role of the Italian stratigraphic record in various accomplishments outlined above and comment on ongoing work and prospects for the future.

The Italian stratigraphic record

Italy, in spite of its small size, is characterized by a varied, geodynamically very active geological landscape that has been the cradle of many geologic concepts from stratigraphy to geodynamics (Krijgsman, 2002). In particular, two marine stratigraphic records in Italy have been of cardinal importance for the GTS: the Mesozoic Tethys pelagic record and the Neogene-Quaternary Mediterranean record.

The Tethys pelagic record

During late Triassic and early Jurassic times, the fragmentation of Pangea resulted in the opening of a large ocean known as Tethys, that separated Eurasia from Africa and the southern continents. Tethyan sediments deposited on the margins of southern Europe and northern Africa make up the bulk of the Alps and the Apennine orogenic belts. A large part of these sediments have been disrupted by tectonic processes but the pelagic record of the African margin (Adria promontory) is well preserved in the Apennines and Southern Alps. In particular, an almost complete pelagic record that spans from the early Jurassic to the early Miocene is preserved in the northern Apennines of central Italy (Marche and Umbria regions; Figure 1). This record, containing plenty of various marker fossils, good magnetic properties, widespread volcanic material and lithologic cyclicity, has been critical in fostering the development of magneto-biochronology (Alvarez et al., 1977; Lowrie et al., 1980), and it is suitable for defining GSSPs.

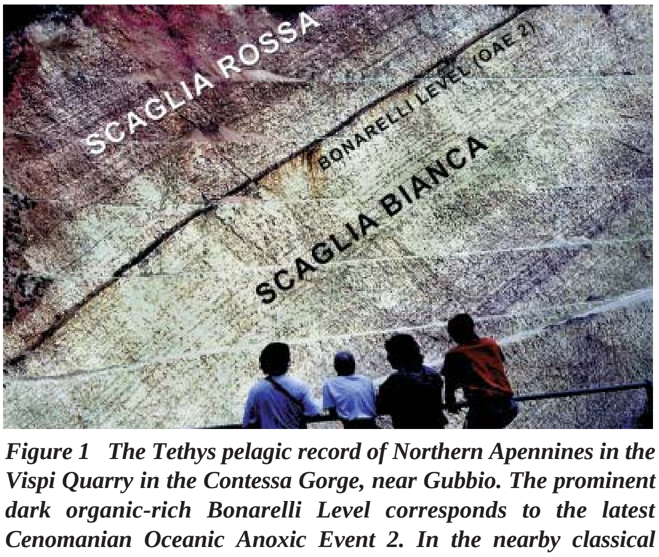

***Figure 1** The Tethys pelagic record of Northern Apennines in the Vispi Quarry in the Contessa Gorge, near Gubbio. The prominent dark organic-rich Bonarelli Level corresponds to the latest Cenomanian Oceanic Anoxic Event 2. In the nearby classical Gubbio section iridum anomaly at the K/T boundary was first recognized.*

The Mediterranean record

The Mediterranean Sea is a relatively recent geologic vestige of the collision of Europe and Africa-Arabia plates, which led to the consumption of most of the Tethys oceanic crust and to a progressively more severe connection with the global ocean. The Africa-Europe collision resulted in thrust-belts, large extensional basins, diverse tectonic arcs. The disruption of the connection with the Indo-Pacific Ocean during the Miocene and the severe progressive restriction of the connection with the Atlantic Ocean since late Miocene, resulted in a semi-enclosed marginal basin that acts as an amplifier of the climatic system. The Mediterranean marine stratigraphic record has thus become an unique archive of the evolution of the Earth climatic system. As far as Italy is concerned, its tectonic shaping is strongly related to the opening of the Algero-Balearic basin during the early Miocene and, overall, to the opening of the Tyrrhenian Sea from the late Miocene to the Pleistocene. Most of the Italian peninsula and Sicily have been uplifted during the Pleistocene providing incredible exposures of late Neogene and Quaternary marine sediments of varied facies that are richly fossiliferous and often cyclically organized (Figures 2 and 3). The Italian late Neogene marine record has been important in introducing and defining CUs of the GTS and the construction of the ATS.

***Figure 2** Orbitally-driven cyclicity in the Trubi Formation (Lower-Middle Pliocene) in Calabria (above) and Southern Sicily. (below).*

The Origins

The first GTS (the Paleontological Time Scale) was established basically in England and in France. However, it has been mainly in Italy, during the Renaissance and the 17th and 18th centuries, that the true nature of fossils was strongly debated and understood by several scholars, such as Leonardo da Vinci (1452–1519), Agostino Scilla (1629–1700), Antonio Vallisneri (1661–1730), and many others. It was in Italy that the Danish scholar Niels Steensen (Steno) formulated in 1669 his principle of superposition of strata. Actually, the first rudimentary Geologic Time Scale might be considered to be that of Giovanni Arduino (1760), who subdivided the geologic record into the Primary, Secondary and Tertiary taking as reference the geology of northeastern and central Italy. In addition, Lyell (1833) and subsequent stratigraphers, in subdividing the Tertiary and Quaternary, made explicit reference to Italian localities as typical of their units. In most published GTSs, the subdivision of the Pleistocene into stages is entirely based on the Italian record (e.g. Calabrian, Sicilian, Tyrrhenian for quoting only the best known). In fact, most of standard stages of the Neogene, except those of the early Miocene, were also introduced with reference to the Italian record (Langhian, Serravallian, Tortonian, Messinian, Zanclean, Piacenzian, and the now obsolete Astian and Tabianian).

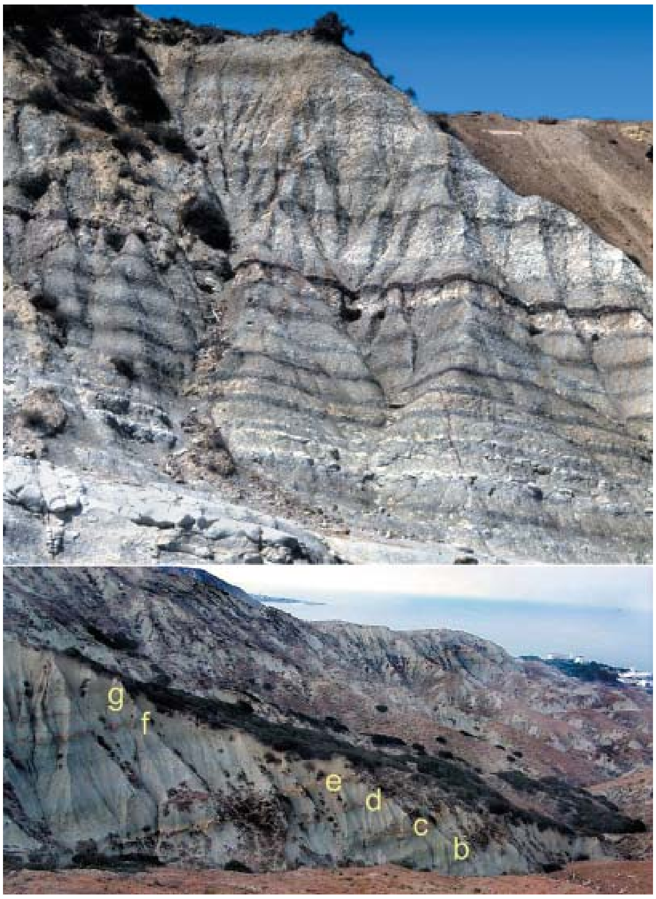

Figure 3 "Sapropels" in the Punta Piccola section, Sicily (Piacenzian GSSP, above) and in the Vrica section, Calabria (Pleistocene GSSP, below).

Magnetobiochronology

The globally synchronous reversals of geomagnetic field polarity were first revealed in dated basaltic lavas a few million years old. Volcanic events, however, occur irregularly and radiometric dates on discrete lavas becomes insufficiently precise to delineate polarity structure beyond ~5 Ma; moreover, volcanic sequences can not easily be tied into the paleontologic scale. Therefore, major advances were achieved after the discovery in the 1960s that marine biogenic oozes and indurated pelagic sediments yield reliable magnetic signals. Combined investigations of paleomagnetic signals and calcareous plankton content from the same continuous pelagic successions allowed the construction of a magnetostratigraphy directly calibrated to multiple paleontologic scales. In this respect, Italian sections played a key role. The pelagic succession from the Umbria-Marche region provided in the 1970s–early 1980s an almost complete magnetobiochronology based on calcareous plankton for the entire Cretaceous up to the Paleogene/Neogene boundary. In particular, the Gubbio section in Umbria was proposed as the magnetostratigraphic type section for the Campanian to Upper Eocene interval (Alvarez et al., 1977). More recently, Cretaceous and Paleogene magnetobiochronology has been improved and updated based on successions from both the Umbria-Marche region and the Southern Alps (Premoli Silva et al., 1988). In the last few years the Italian stratigraphic record has proved important also for the magnetobiochronology of the Triassic System. By calibrating biohorizons based on conodonts and ammonites with magnetostratigraphy in Southern Alps and in Sicily, it has been possible to link the classical continental Newark succession with the marine-based GTS (Muttoni et al., 2001).

Astrochronology

The detailed paleoclimatic record of the Mediterranean has allowed the establishment of an accurate late Neogene ATS that has been based largely on the Italian marine records. Specifically, in the early 1990s, a composite reference section for the Pliocene-early Pleistocene ATS was established by splicing together the Rossello section in southern Sicily (Figure 2) and the Singa and Vrica (Figure 3) sections in Calabria (Hilgen et al., 1997). The derived astrochronology led to significant adjustments to the radiometry-based chronology of the most recent part of the GPTS (Cande and Kent, 1995). Today, the Mediterranean ATS has been extended to ca. 13 Ma (Krijgsman, 2002). The main feature utilized for astronomical tuning is the cyclicity of distinctive dark, organic-rich, often laminated layers known as sapropels (Figure 3). Mediterranean sapropels occur in clusters, which correspond to eccentricity maxima where individual sapropels correspond to insolation maxima associated with precession minima.

GSSPs

The formal definition of CUs by GSSPs has been a slow process, even if it has apparently accelerated in the last years. Even higher rank CUs are still undefined and, hence, often contradictorily used in the literature. According to Remane (in press) only one third of the Phanerozoic stages have been formally defined by GSSPs. It is worth mentioning that out of the 38 GSSPs so far defined, 7 are located in Italy, specifically the bases of the Oligocene, Miocene, Pliocene and Pleistocene Series and the three Stages of the Pliocene (Figure 4). All these GSSPs, except for the base of the Miocene, are located in sections calibrated with astrochronology thus providing unprecedented precision in definition. We refrain here to further comment on the defined Italian GSSPs since exhaustive information is available at the web site of the International Commission on Stratigraphy (ICS; *www.micropress.org/stratigraphy/*). We only comment below on work in progress for defining other GSSPs in the Italian stratigraphic record.

The prospect

We are sure that the Italian stratigraphic record preserves other important clues for implementing the GTS in terms of formal definitions of CUs (GSSPs), improving magnetobiochronology and extending downwards the ATS.

Quaternary GSSPs

Much work is in progress in southern Italy, where thick sections of early to middle Pleistocene marine sediments occur that can be firmly constrained in time with biomagnetostratigraphy (Rio et al., 1996) and stable oxygen isotope stratigraphy. In addition, these sections are characterized often by rich pollen contents, thus allowing a direct link between marine and continental stratigraphies and climates (Capraro et al., 2003). Some of these sections might prove useful for proposing GSSPs for the formal subdivision of the Quaternary (Italian Commission on Stratigraphy, 2002). The formal subdivision of the Quaternary is badly needed because the existing practice and informal proposals are so contradictory that the most recent GTS omits any subdivision (see the web site of ICS). There has been some discussion of defining Quaternary CUs in deep-sea sediment cores. We strongly oppose these proposals (except perhaps for the base of the Holocene) since there are so many impressive, well exposed and high quality marine records available on land in Italy, New Zealand, Japan and California. Others propose we should abandon attempts to subdivide the Quaternary into stages. Again we strongly disagree and concur with Hedberg (1976, p. 82) that "*the basic principles to be used in dividing the Quaternary into chronostratigraphic units should be the same as for the other Phanerozoic strata (...)*".

Other potential GSSPs

In a recent overview by James Ogg (in Gradstein & Ogg, 2002) some Italian sections are reported as candidates for the GSSPs of various CUs. Specifically, candidates GSSPs for the Chattian and Aptian have been indicated in the Umbria-Marche region, a candidate section for the Norian has been proposed in the Sicani Mountains of Sicily and candidate sections for the Carnian and Ladinian have been proposed in Southern Alps (Figure 4). Not reported in this compilation is the recent proposal of the Tortonian GSSP in Monte dei Corvi section (central Italy) made by Hilgen and coworkers. Our ongoing work in the Veneto region indicates potentially useful sections for defining the Bartonian and Priabonian.

Magnetobiochronology and extending ATS

Much work is going on in ODP and expected to continue in IODP for better resolving various microfossil biostratigraphic scales and their calibration to the GPTS, and for extending downwards the ATS. The latter task is particularly difficult because below ca. 25 Ma we lack detailed knowledge of the correct astronomical solutions and new tuning strategies are needed (Pälike et al., 2001). However, we deem that the Italian stratigraphic record will continue to provide precious data. The marine Triassic of the Dolomites, Southern Alps and Sicily is being intensively studied (biomagnetostratigraphy and cyclostratigraphy) for a better definition of this geologic time interval unavailable in the deep sea. Work is continuing for improving magnetobiochronology and deciphering cyclostratigraphy in the Cretaceous and Paleogene records of Umbria-Marche regions. Our group has recently undertaken a systematic study of the Paleogene record of the Veneto region, where a continuous pelagic succession from the basal Paleocene to the late Priabonian is present with a spectacular cyclicity (Figure 5) that has been largely unexplored. The integration of the Italian records with others in the world will certainly contribute to our goal of better founding and better resolving the Geologic Time Scale.

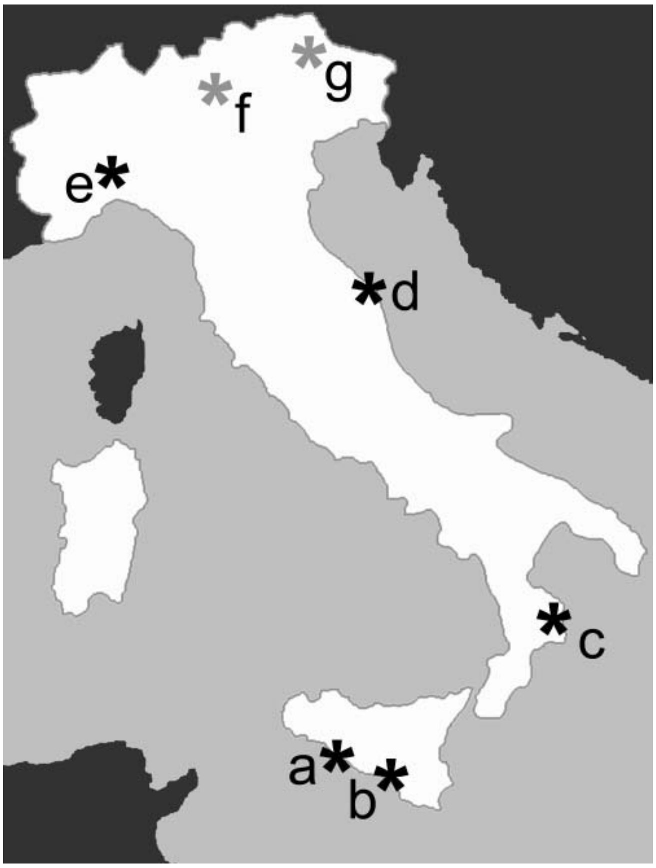

Figure 4 Location map of approved and candidate GSSPs in Italy.
a–Capo Rossello Composite Section (Sicily): Pliocene Series and Zanclean and Piacenzian Stages GSSPs;
b–Monte S. Nicola (Sicily): Gelasian Stage GSSP;
c–Vrica (Calabria): Pleistocene Series GSSP;
d–Massignano (Marche): Oligocene Series and Rupelian Stage GSSP;
e–Lemme-Carrosio Section (Piedmont): Miocene Series and Aquitanian Stage GSSP;
f–Bagolino (Central Southern Alps): candidate for the Ladinian GSSP;
g–Prati di Stuores (Dolomites): candidate for the Carnian GSSP.

Figure 5 Early Eocene pelagic sediments cropping out in the Venetian Alps (Northeastern Italy). Integrated biomagnetostratigraphy is in progress, nevertheless the very meaning of this paramount cyclicity is still unexplored.

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***Domenico Rio** is Professor of Paleontology at the University of Padova (Italy) and past Chairman of the Subcommission of Neogene Stratigraphy. His research includes calcareous nannofossil paleontology, Mediterranean Neogene stratigraphy and Northern Apennines geology. Presently he is working on Pleistocene climatic variability in the Mediterranean marine stratigraphic record and on the extreme climatic events in the Early Paleogene of the Southern Alps.*

***Isabella Premoli Silva** is Professor of Micropaleontology at the University of Milan (Italy) and past Chairman of the Subcommission of Paleogene Stratigraphy. Her research includes detailed biostratigraphic studies of Cretaceous to Miocene planktonic foraminifers from land and deep-sea deposits. She has participated in seven cruises, in two as Co-Chief scientist, of the Deep Sea Drilling Project and Ocean Drilling Program. Her present research topics deal with the response and recovery of planktonic foraminifers to the extreme climatic warmth events of the Cretaceous and early Paleogene.*

***Luca Capraro** is Assistant Researcher at the University of Padova (Italy). His research includes Pliocene and Pleistocene stratigraphy of the Mediterranean with a special emphasis in reconstructing the vegetational history and climatic variability of the Central Mediterranean during the middle Pleistocene climatic transition.*

by Antonio Brambati[1], Laura Carbognin[2], Tullio Quaia[1], Pietro Teatini[3], and Luigi Tosi[2]

The Lagoon of Venice: geological setting, evolution and land subsidence

1 Dipartimento di Scienze Geologiche, Ambientali e Marine, Università di Trieste, Via E. Weiss 2, I-34127 Trieste, Italy.
2 Istituto di Scienze Marine, Consiglio Nazionale delle Ricerche (CNR), San Polo 1364, I-30125 Venezia, Italy.
3 Dipartimento di Metodi e Modelli Matematici per le Scienze Applicate, Università di Padova, Via G.B. Belzoni 7, I-35100 Padova, Italy.

The paper deals with the geological setting, history and subsidence of the Venetian Plain. Major attention is paid to the Pleistocene-Holocene stratigraphic sequence in the Lagoon of Venice, in relation to its origin that dates back to 6–7 kyr BP. Geological land subsidence, which played an important role in the origin and the evolution of the lagoon, and anthropogenic subsidence, that has recently assumed a major importance for the Venetian environment, are discussed. Considering also the sea level rise, 23 cm loss in land elevation has occurred in the last century, leading to increased flooding events and environmental problems that require protective works.

Introduction

The lagoon of Venice is the largest lagoon in Italy, and the most important survivor of the system of lagoons which in Roman times characterized the upper Adriatic coast from Ravenna to Trieste. Bounded by the Sile River to the North and the Brenta River to the South, the Venice Lagoon is oblong and arched in shape. It covers an area of about 550 km^2, being 50 km long and 8–14 km wide. Its morphology consists of shallows, tidal flats, salt marshes, islands and a net of channels. The lagoon boundaries also include fish ponds, reclaimed areas and the coast that is presently interrupted by three inlets, namely Lido, Malamocco and Chioggia, which permit water exchange with the Adriatic Sea (Figure 1). The present setting of the Venice Lagoon is mainly the result of a number of human interventions.

Origin and evolution of the Lagoon

It has been demonstrated (Gatto and Carbognin, 1981) that the lagoon of Venice originated nearly 6–7 kyr BP during the Flandrian transgression, when the rising sea flooded the Upper Adriatic Würmian paleoplain and outlined the coast in approximately the present position. The early lagoon was smaller than the present one and the exchange of its waters with the sea occurred through eight inlets, against the three it has now. Originally, two main factors affecting the lagoon basin: i) the continuous sediment supply from Adige, Bacchiglione, Brenta, Sile and Piave rivers flowing into the lagoon, so that the filling was greater than the natural subsidence and the eustatic sea level rising; ii) the noticeable coastal nourishment, also coming from the Po river to the South, that led to a gradual silting-up of the tidal inlets.

These two processes steered unavoidably to the disappearing of the lagoon basin. Venetians, considering the lagoon a source of security against enemies and power with its channels and port, began to carry out several hydraulic works to preserve it. Mostly, the diversion of the major tributaries into the sea was started in 1400 AD and concluded in 1600 AD. They avoided making the lagoon a marshland, but also induced an abrupt reversal in its natural evolution. With the passing of time, sea properties began to prevail, enhanced further by human interventions. In fact, since 1800 man has newly altered the lagoon setting very intensely. He dug new deep canals and modified the sea openings, both in the number and the setting, to serve the industrial harbour; occupied intertidal flats to provide the necessary areas for new industrial and urban centers, and permanently closed areas for fish farms. These changes were decisive in spurring the lagoon hydrodynamics, accelerating erosion and modifying the flora and fauna habitat. Furthermore, in the last decades the industrial water supply was provided by rash exploitation of artesian aquifers causing a serious land subsidence. An induced consequence is a weakening of the littoral system (Brambati, 1987) with a deepening of the sea bottom that also contributes to the instability of the lagoon itself (Carbognin et. al., 1995).

Geological and sedimentological setting of the Gulf of Venice

Referring to the whole Venetian Plain, the geological setting down to about 5,000 m consists of Prepliocene, Pliocene and Quaternary deposits (Figure 2). The pre-Quaternary substratum southwards is characterized by fold and faulted overfolds, which are parallel to the main tectonic trend of the Apennines and include several gas-bearing traps at depths on the order of 2,000 m. Quaternary sediments range between 3,000 m (southern zone) and hundreds of meters (northern zone). They mostly consist of sandy and silty-clayey layers of alluvial and marine origin. The bottom follows the structure of the substratum showing a little tectonic disturbance only in the northern sector where the lagoon of Venice is found. The thickness of the Neozoic formations and, consequently, the subsidence rate exhibits a non-uniform space distribution.

The area of Venice is located in a complex foreland setting near the pinchout of both the Southalpine and the Apenninic wedges, that developed during Serravallian-Messinian and early Pliocene-Pleistocene times, respectively. The post-Messinian depositional sequence is represented by deep-sea hemipelagic deposits that drape the eroded Messinian surface: the *Santerno Clays* (Figure 3). This formation comprises Pliocene to middle Pleistocene sequences that crop out along the northern edge of the Apennines and is represented by isolated outcrops on the Alpine edge of the Po Valley. Its depositional environment is quite deep, from outer neritic to bathyal. The Santerno Formation is confined between the Messinian unconformity and the overlying Pleistocene *Asti Sands*. The Asti Formation displays an overall shallowing trend from turbiditic to deltaic/shallow-marine and finally continental settings. The present-day Venetian area was reached by the north-Adriatic turbidite system during the early to middle Pleistocene, due to the north-eastward shifting of the Apenninic foredeep. The *Asti Sands* turbiditic sequences show

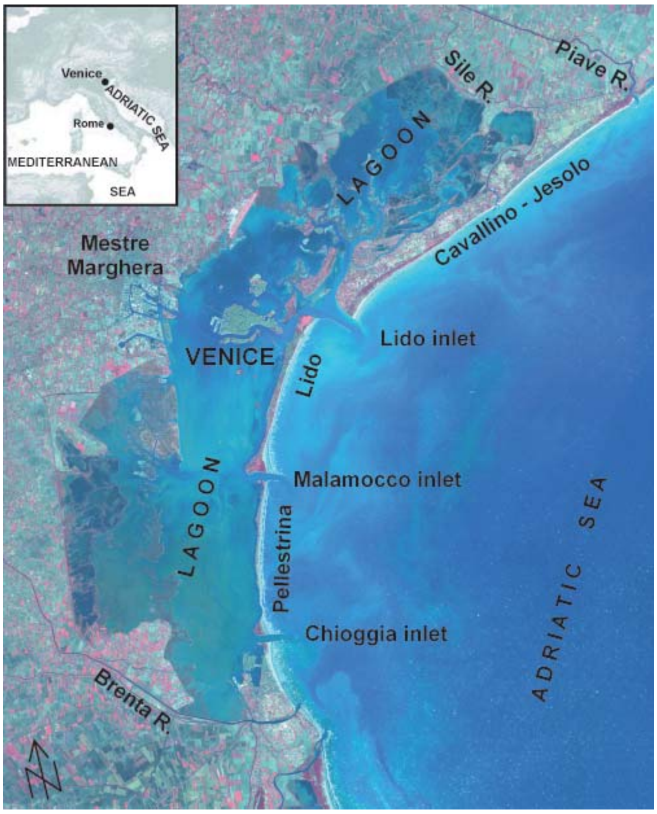

Figure 1 ASTER image of the Venice Lagoon and its surrounding mainland. Main localities are indicated.

flat parallel surfaces that onlap against the clayey substratum of the *Santerno Clays*. Peat-rich deposits close to the top of the *Asti Sands* indicate a floodplain environment. The thickness of the Asti formation is fairly uniform and averages 1,000 m. Close to the Po River delta, its thickness increases to up to 2,000 m.

Geological and palaeogeographical feature of the Venice area

Knowledge on geological and lithostratigraphical settings of the Venice area results from thousands of different analyses based on hundreds of cores drilled on purpose.

Main information sources about the Plio-Pleistocene subsoil are two boreholes: the VE-1 CNR, that extends down to 947 m with continuous coring, and the VE-2 CNR, down to 400 m, drilled with discontinuous coring. Recently, using an integrated magneto-bio-cyclo-stratigraphy of lithofacies and palynofloral analyses in the VE-1 core, Kent at al. (2002) could infer the following history: in the late Pliocene the depositional area was a strongly subsiding shelf which shoaled to near sea level; following a hiatus of at least 0.2 Myr the shelf rapidly drowned to bathyal depths over the early Pleistocene, and hemipelagic muds with sapropel layers were deposited; these are followed by a thick package of basinal turbidites, fed from the eastern Southern Alps; then, in the middle part of Brunhes, Po-related deltaic sedimentation led to the progressive infill of the basin; this episode ended with the first appearance of continental deposits; the upper part of the succession shows a cyclothemic organization with submergence of Venice area during glacio-eustatic highstands and emergence during glacial lowstands.

From the bottom of the borehole up to about 513 m in depth, the sands are characterized by a high amount of carbonate rock fragments (eastern South Alpine provenance). Going up to about 438 m in depth the composition abruptly changes into quartzolithic with

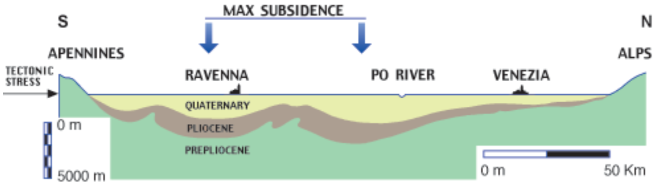

Figure 2 Schematic geological section across the eastern Po Plain (modified after AGIP Mineraria, 1969).

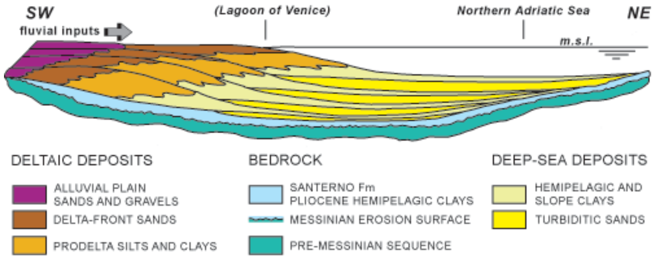

Figure 3 Sedimentological model of the post-Messinian basin across the Lagoon of Venice area and the northwestern Adriatic Sea (modified after AGIP Mineraria, 1999, unpublished report).

large amounts of quartz and metamorphic rock fragments (from Alps/Northern Apennine). This change is indicative of a significant middle-Pleistocene tectonic movement of the region related to the dynamic of the Apennine foredeep. From about –438 to –19 m in depth, the sand composition is a mixture of the two types of sediments, reflecting the simultaneous activity of the two sources (Stefani, 2002). The uppermost part of the succession contains numerous peat layers indicative of floodplain to marsh environmental settings; in this part of the core a chronology was established on the basis of radiocarbon data.

Kent et al. (2002) distinguished six prominent sea-level transgressions *tr1* throughout *tr6*, that occur at depths 10.5, 79, 136, 152, 202 and 262 m respectively, where continental sediments are overlain by shoreface and shelf marine deposits.

A good correlation with these transgressions results from recent analyses performed on thousands of cores. As examples, *tr6* and *tr3* may correspond to the top of regional aquifers 5 and 2 (Figure 4) and *tr1* reflects the base of the Flandrian transgression (Figure 5). The limit Holocene-Pleistocene is marked by a clay layer called *caranto* (Figure 6) which overcompacted because of the dry climate during the last phase of the low stand sea level. In spite of its discontinuity, it is an optimum marker horizon that ends the alluvial Pleistocene sequence (Gatto and Carbognin, 1981; Bortolami et. al, 1985). The *caranto* tends to emerge on the mainland, and varying between –5 and –23 m, gradually deepens towards the littoral (see Figure 5). A hiatus covering a period between 7 and 10 kyr from the last Pleistocene to the first Holocene deposition has been found (Bortolami et al., 1985; Tosi, 1994). The following marine Flandrian transgression progressively submerged the Würmian paleoplain and the *caranto*. The first presence of marine-lagoonal Holocene deposits founded in layers underlying the present southern littoral (Figure 7) have been dated 10–11 kyr BP (Bortolami et al., 1985). The early Holocene sediments are represented by a discontinuous level of silt and sand often in chaotic structure mixed with shelly marine-lagoon sands. The maximum flooding position was reached at 6–7 kyr BP when the primeval lagoon established. The middle-upper part of the series is a typical alternation of marine-lagoon and flood-plain sediments. During the following high-stand sea level, alluvial sedimentation re-established at the southern and northern outermost belts corresponding to the fluvial mouths. Moreover, some areas have been affected by episodes of emersion and submersion related to changes in climate, sediment source, subsidence rate and, finally, human intervention (Tosi, 1994; Bonardi et al., 1998; Carbognin and Tosi, 2002).

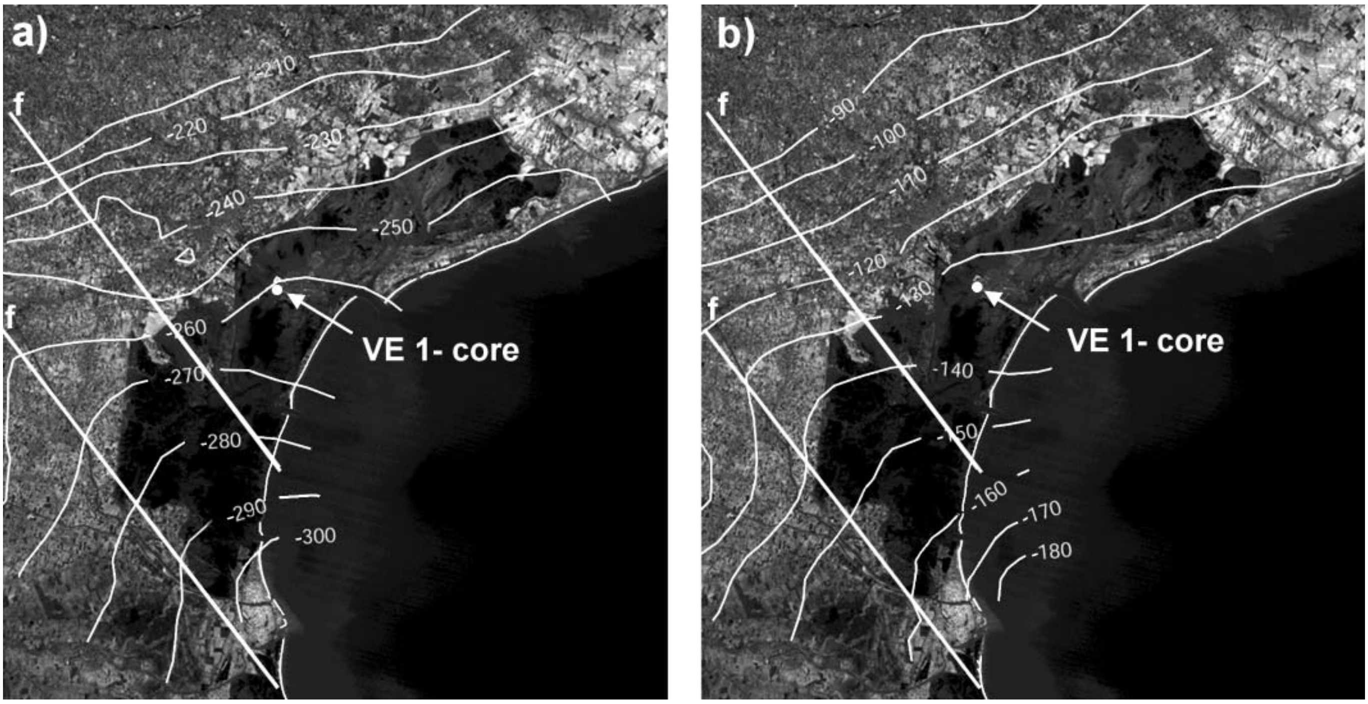

***Figure 4** Maps of the top (m a.s.l.) of: a) aquifer 5 and b) aquifer 2, which may correspond to tr6 and tr3, respectively, described by Kent et al. (2002). Straight lines (f) are drawn in correspondence to regional faults.*

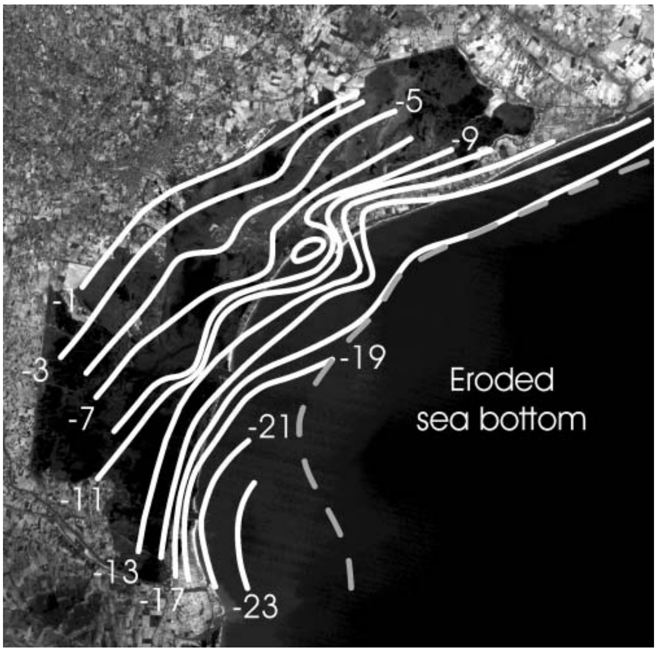

***Figure 5** Depth of Pleistocene-Holocene boundary (m a.s.l.) drawn on the basis of cores analyses and high-resolution seismic survey (courtesy of Aliotta, Carbognin, Stefanon). This surface agrees with the tr1 by Kent et al. (2002). Dashed line delimits the marine area where Holocene deposition has been eroding.*

Geological Subsidence

Estimates of natural subsidence rates result from both analysis on the VE 1-core, and from radiocarbon dating of latest Pleistocene and Holocene sediments collected in the lagoon and littorals (Bortolami et al., 1985).

According to Kent et al. (2002) the average long-term subsidence rate (less than 0.5 mm/yr) reflects mainly tectonic processes; it is rather lower than that occurring in the late Pleistocene-Holocene period (average rate of 1.3 mm/yr) which likely reflects the consolidation of sediments, and it is decidedly lower than the recent man-induced one (≥ 2.5 mm/yr). The average rate of 1.3 mm/yr dropped over recent centuries, reaching the current figure of 0–0.5 mm/yr (Gatto and Carbognin, 1981; Carbognin et al., 1995, 2003). Evidence of past tectonic influence could be the anomalous shape of the *tr3* and *tr6* contour dips close to the faults (see Figure 4).

Land subsidence due to natural consolidation played a major role during initial evolutionary phases of the modern lagoon. Later, the increased salt concentration in the clayey sediments of the substratum, inducing an electrochemical compaction process, caused a further lowering of the lagoon floor (Gatto and Carbognin, 1981).

Anthropogenic subsidence

Subsidence induced by groundwater withdrawals became a problem with the industrial boom after the 2nd World War. This process has been deeply studied, the cause-and-effect relationship quantified, and a 2-D and a 3-D simulation models were developed.

The exploited aquifer-aquitard system is located in the upper 350 m of the 1000 m thick unconsolidated Quaternary formation. Groundwater withdrawal, which progressively led to a noticeable drawdown of aquifer pressure, began in the 1930s reaching a peak between 1950–1970 together with the subsidence it caused (the maximum rate of 17 mm/yr was recorded between 1968–69 over the industrial zone). After this critical 20-year phase, a general improvement occurred quickly because of the closure of artesian wells and the diversification of water supply. The 1973 levelling clearly showed a reversed trend of subsidence and a slight but significant rebound was measured in 1975, with a maximum of 2 cm in the historical city, as predicted by simulation models (i.e. Gambolati et al., 1974). The 1993 and 2000 regional surveys confirmed the arrest of anthropogenic subsidence as a widespread phenomenon. Subsidence maps for the lagoon territory and the hinterland (Figure 8) show an ideal *line of demarcation* between the ground stability zones in the mainland, Venice and its surroundings, and the subsiding zones at the northern and southern extremities of the lagoon edge and along the littoral. This fact may be related to the greater thickness of the

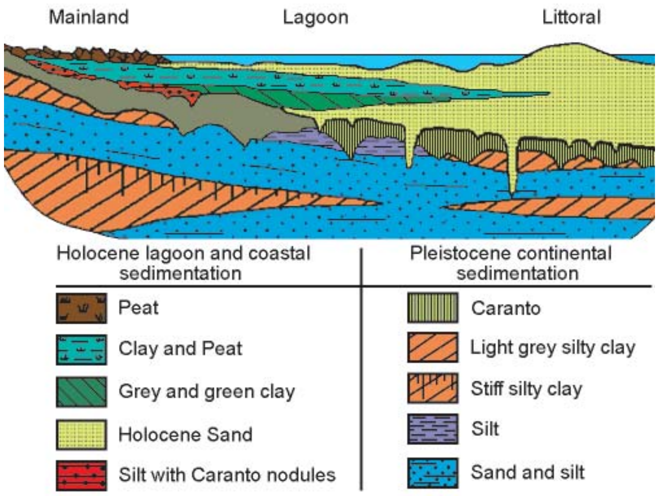

Figure 6 Holocene-Pleistocene stratigraphical sequence across the central Lagoon of Venice (after Gatto and Previatello, 1974).

sandy layers present along the Mestre-Lido axis with respect to those found in north and south lagoon areas where compressible clayey layers predominate. On the other hand, this *line of demarcation* could be correlated with maximum Flandrian transgression of about 6 kyr BP (see Figure 7). Concerning the subsidence recorded along the coastline (1–3 mm/yr), and at the furthermost northern and southern boundaries (2–4 mm/yr), it can be attributed to the greater natural consolidation of more recent formation in these areas, and to different local situations. Finally, both the uplift and the highest subsidence rate found in the central-southern lagoon edges may partly reflect tectonic activity. Integrated researches, still in progress, based on hydrogeological, geomorphological, geophysical and geoelectrical investigations, indicate the presence of the two faults (Figures 4 and 8). Satellite Radar Interferometry (InSAR) maps confirm this regional subsidence behaviour (Figure 8) and supply useful details of urban areas (Tosi et al., 2002).

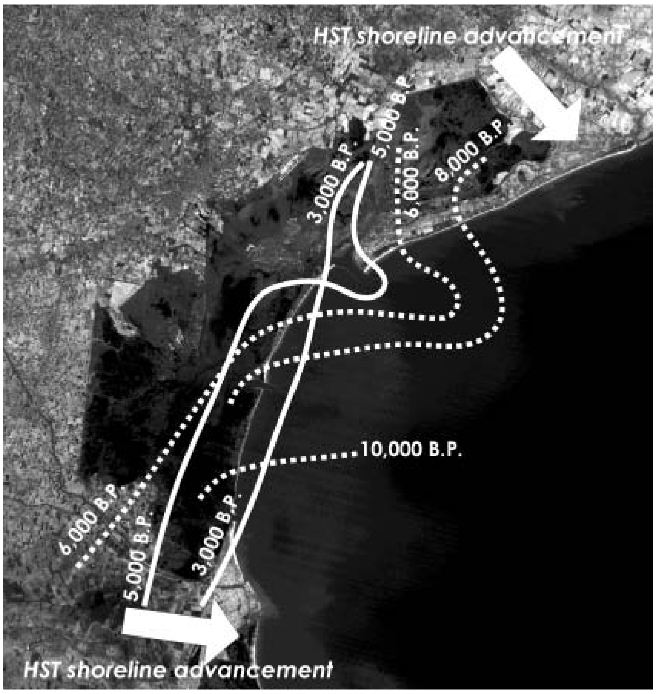

Figure 7 Schematic model of the Venetian littoral evolution during the Flandrian transgression. Arrows indicate the shoreline advancement during the high-stand sea level because of the progradation of the river mouths (modified after Tosi, 1994).

The overall relative land subsidence

The relative land subsidence of the city is associated with sea level rise. During the last century the elevation loss of 23 cm, consisting of about 12 cm of land subsidence, both natural (3 cm) and anthropogenic (9 cm), and 11 cm of sea level rise, has occurred. Referring to the latter issue, the most reliable estimate of the sea level rise in the Upper Adriatic over the last 100 years is given by Carbognin and Taroni (1996): excluding subsidence, a rising rate of 1.13 mm/yr was computed. The extent of the period is sufficient to average alternating

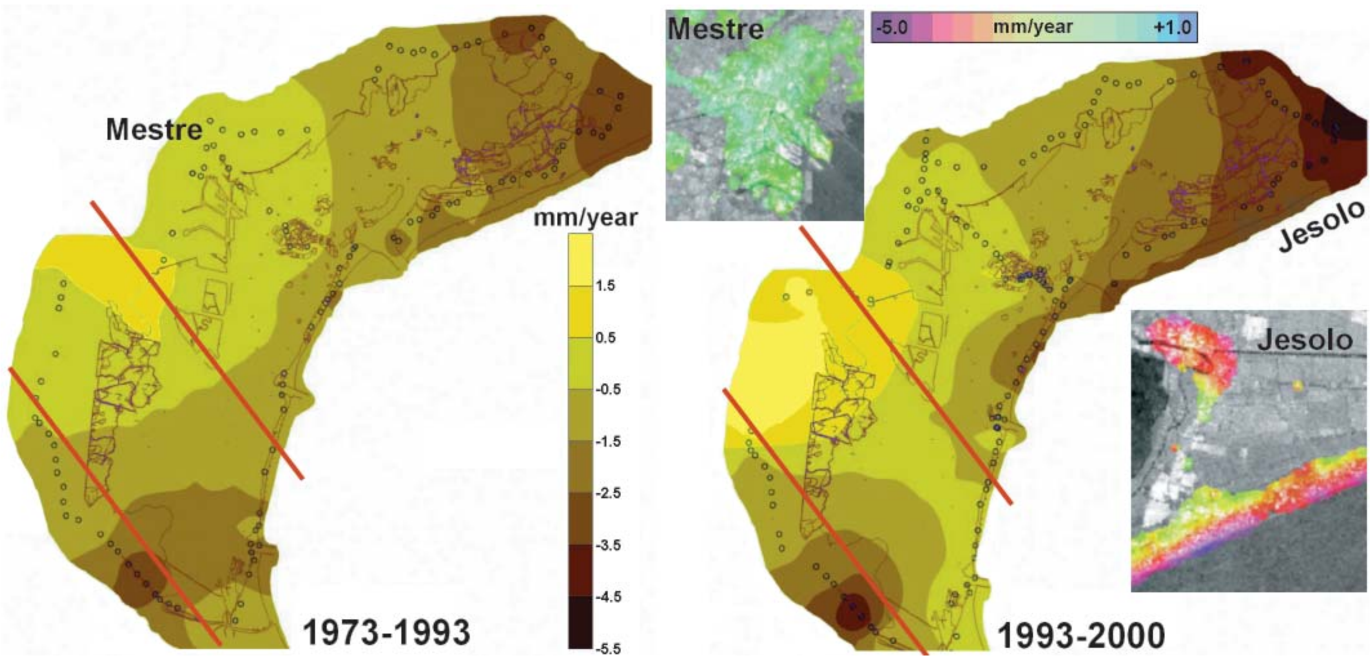

Figure 8 Maps of annual vertical displacement (mm/yr) occurred in the periods 1973–1993 and 1993–2000 (after Carbognin et al., 2003). For the latter period, inserted smaller InSar deformation maps (after Strozzi et al., 2003) show in detail the two different situations of land stability at Mestre and subsidence at the northern littoral (Jesolo). Red lines correspond to regional faults.

trends, corresponding to alternating climate changes. This rate is consistent with the data provided by others tide gauges in the Mediterranean Sea. Worth mentioning is that in the last decades the rise of sea level is slowing down slightly both in the Mediterranean Sea and in the Indian Ocean (Tsimplis and Baker, 2000; Mörner et al., 2003).

The relative 23 cm rise in the sea level in the Lagoon has created a great concern since it has contributed to the increasing of: i) the flooding phenomenon, both in frequency and degree, with immediate and indirect damages to population and monumental patrimony; ii) the hydrodynamics inside the lagoon, leading to erosion of its floor, channel silting up, and changes in the internal eco-morphology; and iii) the fragility of littorals, enhancing the risk of destructive sea storms and flooding from overtopping.

Conclusions

The Lagoon of Venice is the largest one of the Mediterranean Sea and is located on a foreland between the Alps and the Apennines. The lagoon originated 6–7 kyr BP, establishing its domain over the Würmian paleoplain, which has influenced its original morphology. Subsequent evolution of the basin has occurred through different phases. Initially, during the high-stand sea level, a significant role was played by the alluvial yield, which was not counterbalanced by both sea level rise and subsidence causing the filling in of the basin and progradation of the river mouths. In historical times, human intervention, going from the diversion of tributaries to more recent rash groundwater exploitation, has reversed the natural evolutionary trend, favouring the deepening of the lagoon and seriously modified the morphological setting of the environment.

Summarizing the results, it can be said that natural subsidence ranged between 0.5 and 1.3 mm/yr during the Quaternary, and that man-induced processes more than doubled the rate in the period 1950–1970. The land survey of 2000 shows that subsidence is no longer occurring in the central part of Venetian area, which includes the city, the Mestre-industrial zone and surrounding territories, but it is still going on at the northern and southern lagoon areas and bordering lands. Anyway, since subsidence is mostly irreversible and a contribution to the land lowering with respect to mean sea level is given by eustasy, 23 cm of relative elevation loss has occurred during the last century, inducing consequences that require restoration and safeguarding measures.

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Antonio Brambati *is Head of the Department of Geological, Marine and Environmental Sciences at the University of Trieste, and Full Professor of Sedimentology since 1975. Former President of the CNR Oceanographic Commission and of the Osservatorio Geofisico Sperimentale (OGS) of Trieste, he has been coordinator of many projects on the Mediterranean, Ross Sea (Antarctica) and the Magellan Strait. His interests deal with marine Quaternary geology, sedimentology of the coastal zone, marine pollution and paleoclimatology. He is Director of the UNESCO 'International Course on Coastal Management' in Venice.*

Laura Carbognin *is Research Director at the CNR in Venice. As geostatistical senior scientist, her main interests concern the understanding of environmental problems related to land subsidence, coastal processes and sea level. For her innovating statistical approaches to environmental problems, she has received international recognition in the book "Italian Contributors to the Methodology of Statistics", 1987. She has been co-ordinator of many international and national research projects, and Italian member of the UNESCO Working Group on Land Subsidence since 1977.*