

Crustal structure, tectonic evolution and seismogenesis in the Northern Apennines (Italy)

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ABSTRACT The aim of this paper is to illustrate how the time and space relationships between compression and extension in the Northern Apennines can be expressed through three main rules: i) at any time, compression and extension are contemporaneously active, in the foreland and in the hinterland, respectively; ii) at any position, extension follows compression; iii) extension can occur in the uppermost crust at the same time and position of compression at greater depth (lower crust and upper mantle). These rules are always verified, by considering both the long-term evolution of the Apennines, illustrated by the geological data about the age of syntectonic basins, and the short-term, present-day image, highlighted by the instrumental seismicity pattern. A comprehensive review of the geological and geophysical evidence supporting the contemporaneous activity and eastward migration of compression and extension in the Northern Apennines is reported. The main results of the CROP03 project, integrated with new geophysical data sets, are discussed together with the space distribution and kinematics of instrumental seismicity recorded in the last 20 years, that has furnished a present-day, instantaneous image of the evolution of the Northern Apennines. The last part of the paper presents and discusses some relevant and debated topics on the setting and evolution of the Northern Apennines.

1. Introduction

Since Lower Miocene (Late Burdigalian), the tectonic evolution of the Northern Apennines and of their hinterland, the northern Tyrrhenian Sea, is characterised by the contemporaneous activity and eastward migration of coupled compression (in the foreland) and extension (in the hinterland). Merla (1951) first recognised that the foredeep basins of the Northern Apennines become younger towards the east, following the migration of the compressional front. The model was completed by Elter *et al.* (1975), who described the contemporaneous (Tortonian to present) eastward migration of extension in the Tuscan region. In the northern Tyrrhenian Sea, the onset of extension was set at Late Burdigalian time, about 17 Ma (Bartole *et al.*, 1991). A similar evolution is described by Ghisetti and Vezzani (2002) mainly referred to the Central Apennines.

The eastward migration of contemporaneous compression and extension was later reprised by many Authors and used as a key for understanding the crustal structure, the distribution and kinematics of seismicity and the geodynamic evolution of the Northern Apennines (e.g. Lavecchia *et al.*, 1984, 1987, 1994; Royden *et al.*, 1987; Piali *et al.*, 1995; Jolivet *et al.*, 1998).

These ideas were further developed and confirmed by the results of the CROP03 project, imaging the present-day crustal structure of the region composed of two structural domains, an internal Tyrrhenian domain, dominated by extension and an external Adriatic domain, dominated by compression (Barchi *et al.*, 1998a; Decandia *et al.*, 1998; Lavecchia *et al.*, 2004). However, the same data were also used to propose an alternative interpretation, denying any substantial Tertiary extension of the Tyrrhenian domain (Finetti *et al.*, 2001).

Many different geodynamic models have been proposed to explain this peculiar tectonic setting and evolution, whose discussion is not the object of this paper. Beyond significant differences, most models refer to the west-dipping subduction and sinking of the Adriatic continental lithosphere, delaminated at a more or less shallow level, whose eastward retreat make possible the positioning of a shallow asthenospheric wedge, below a new formed, severely extended and relatively thin Tyrrhenian crust [Reutter *et al.* (1980) and Channel and Marshall (1989) among many others]. Alternatively, the evolution of the Tyrrhenian-Apennine system has been explained by applying either a passive or active rifting; e.g., Lavecchia *et al.* (2003a) recently interpreted the evolution of this area as the result of a combination of large-scale, plume-induced lithospheric stretching (Tyrrhenian Sea) and of a local scale rift push, inducing crustal shortening towards the east (Apennines), expressed by deep thrusts, involving the whole crust and the upper mantle (Lavecchia *et al.*, 2003b, 2004).

The aim of this paper is to make a comprehensive review of the geological and geophysical evidence supporting the contemporaneous activity and eastward migration of compression and extension in the Northern Apennines, illustrating how the time and space relationships between compression and extension can be expressed through three main rules, more or less explicitly enounced in the previous literature:

- a. at any time, compression and extension are contemporaneously active, in the foreland and in the hinterland, respectively;
- b. in any position, extension follows compression;
- c. extension can occur in the uppermost crust at the same time and in the same position of compression at greater depth (lower crust and upper mantle).

We first describe the crustal structure of the Northern Apennines, considering the results of the CROP03 project (Pialli *et al.*, 1998), integrated by new geophysical data sets, made available after the conclusion of the project (e.g. stress maps of the Apennines, seismological data, strain vectors inferred by GPS data, palaeomagnetic data, structure of the lower crust and upper mantle).

We also synthesize and discuss two complex data sets, describing the tectonic setting of the Northern Apennines over two very different time spans:

- the age of the syntectonic basins (both hinterland and foreland), constraining the tectono-sedimentary history of the region since Late Burdigalian (17 Ma);
- the space distribution and kinematics of instrumental seismicity recorded in the last 20 years, recently reviewed by Chiarabba *et al.* (2005), that is an instantaneous image of the present-day evolution of the Northern Apennines.

Using this approach, we try to explain the tectono-sedimentary evolution of the Northern Apennines, its present-day crustal structure and seismotectonic setting as the result of a coherent and quasi-continuous evolution, started at about 17 Ma.

2. The crustal structure of the Northern Apennines

The Northern Apennines can be divided into two different structural domains, whose distinction is underlined by their peculiar geophysical and geological features, as well imaged by the results of the CROP03 project (Barchi *et al.*, 1998a). A western, Tyrrhenian domain, where extensional deformation destroyed the pre-existing compressional belt, and an eastern, Adriatic domain, where the compressional structures are present today (Fig. 1a). The former is characterized by a strong reduction of lithospheric thickness detected by passive seismology (Calcagnile and Panza, 1981; Suhadolc and Panza, 1989; Amato and Selvaggi, 1991, Piana Agostinetti *et al.*, 2002; Di Stefano *et al.*, 2005). This interpretation is supported by the study of

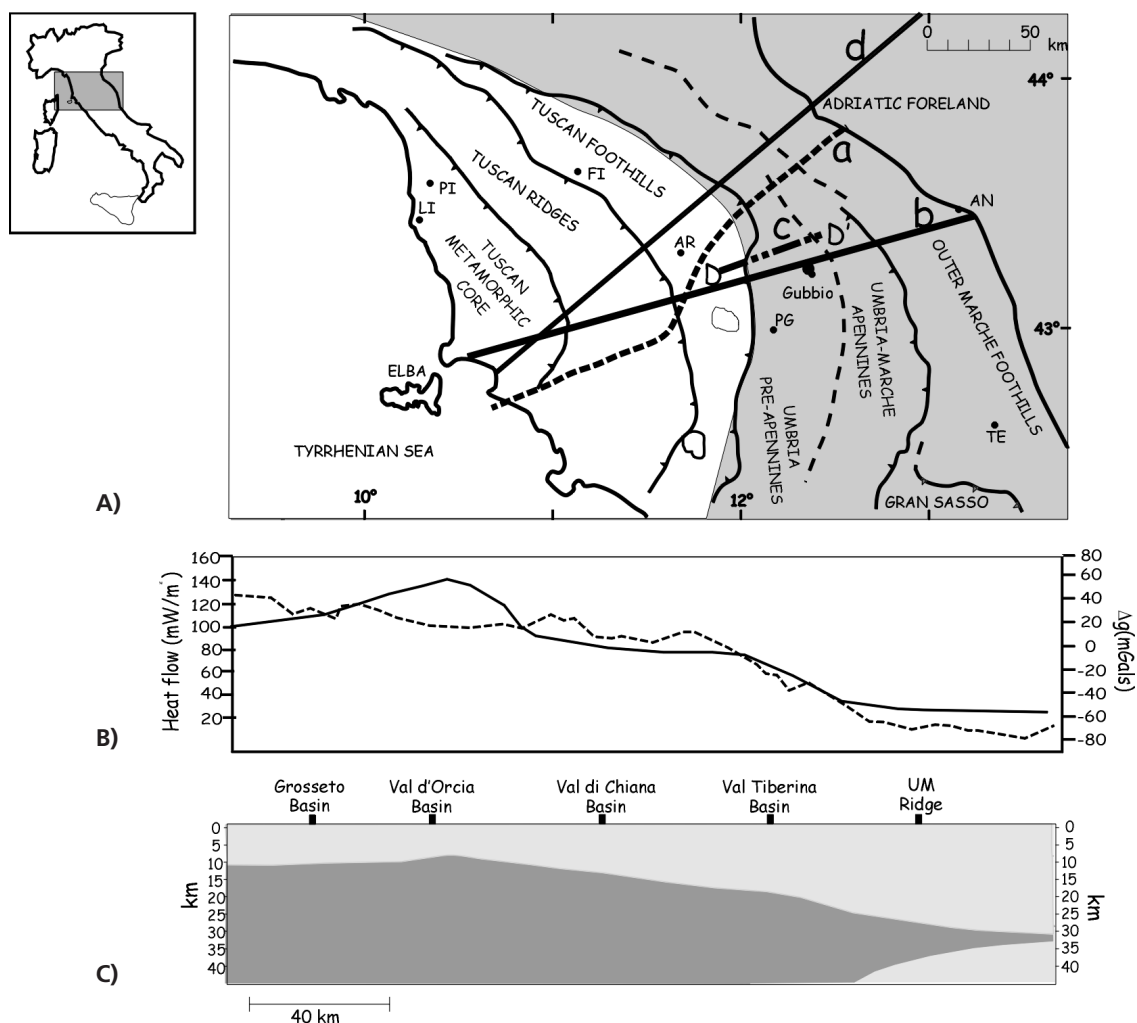


Fig. 1 - Location map of the study area. A) Simplified geological map of the Northern Apennines. Tyrrhenian domain (white); Adriatic domain (gray). a) CROP03 NVR seismic line; b) DSS (Deep Seismic Sounding) seismic refraction; profile. c) D-D' seismic profile of Fig. 5; d) local earthquake travel-times tomographic profile of Fig. 2b. B) Heat-flow observations at the surface (continuous line) (Mongelli *et al.*, 1998) and Bouguer anomaly (broken line) (Marson *et al.*, 1998) along the CROP03 line. C) Brittle (light gray) and ductile (dark gray) field along the CROP03 profile (Pauselli and Federico, 2002).

the thermal data in this region, which locates the lithosphere-asthenosphere boundary (1600 K isotherm) at a depth of about 30 km (Pauselli and Federico, 2002). The overall thinning of the lithosphere (Fig. 1b) is reflected in the high-surface heat flow [$> 150 \text{ mW/m}^2$; Della Vedova *et al.* (2001)] and provides an explanation of the positive Bouguer anomalies found in this domain (Marson *et al.*, 1998). In addition, the “elastic LID” (the uppermost mantle between the Moho and the underlying low-velocity layer) is less than 15 km thick (Peccerillo and Panza, 1999; Pontevivo and Panza, 2002) indicating a considerable amount of partial melting. A relatively shallow Moho is detected by seismic refraction surveys at a depth of between 22 and 25 km (Ponziani *et al.*, 1995; De Franco *et al.*, 1998). The study of the thermal field (Fig. 1c) locates the brittle/ductile (B/D) transition at a depth of 10-12 km (Pauselli and Federico, 2002), in good agreement with the distribution of the earthquake hypocentres, which are concentrated in the uppermost 10 km of the crust (Amato and Selvaggi, 1991; Chiarabba *et al.*, 2005).

The “Adriatic domain” (Fig. 1a) is characterised by a lithospheric thickness of about 70-90 km and the Moho is located by seismic refraction surveys (Ponziani *et al.*, 1995; De Franco *et al.*, 1998) at a depth of about 35-40 km. The Bouguer anomalies are negative and there are relatively low heat flow values (between 70 and 40 mW/m^2). Shallow seismic events (depth less than 10 km), with strike-slip and thrust fault plane solutions, suggest still active compression in the external area. Below the main ridge of the Apennines, seismicity shows different patterns: moderate earthquakes occur on extensional faults in the uppermost crust (7-12 km deep) and a light to moderate seismicity, probably compressive in character, is located within the deep crust and at subcrustal depths [30-100 km: Selvaggi and Amato (1992) Chiarabba *et al.* (2005)].

Concerning the rheological behaviour, the B/D transition deepens from 12 km to about 25 km moving from west (Val di Chiana basin) to east (active front of compression). From the active compressional front to the Adriatic region, the rheological behaviour of the lithosphere (Fig. 1c) is brittle at a depth of less than 25 km and greater than 34 km (Pauselli and Federico, 2002).

The CROP 03 seismic reflection profile [Fig. 2a: Barchi *et al.* (1998a)] has shown that the Tyrrhenian and Adriatic domains, with distinctive characteristics of both geological and geophysical points of view, also exhibit a distinct reflectivity pattern at any crustal level. In the upper crust of Tuscany (Tyrrhenian domain), extensional tectonics, active since the middle Miocene, has largely obliterated the previous compressional structures so that they cannot be recognized on seismic profiles and local earthquake tomography (Chiarabba *et al.*, 1995). In contrast, compressional structures are still clearly preserved in the Umbria-Marche Apennines (Adriatic domain) where extension has been active for a much shorter time period, about 3 Ma, and has only affected the westernmost part of the region (Barchi *et al.*, 1998b). Both seismic reflection profile (Barchi *et al.*, 1998b) and local earthquake tomography (Chiarabba and Amato, 2003) show that the compressional structures are the most visible features of the upper crust that are still not overprinted by the active extension.

The Tyrrhenian lower crust is characterized by several, discrete sub-horizontal reflections, some of them laterally continuous for as much as 10 km, whilst the Adriatic lower crust possesses a weak and diffuse reflectivity, without any prominent reflections (Magnani, 2000).

The Moho in the Tyrrhenian domain is shallow [$< 25 \text{ km}$: Locardi and Nicolich (1988)], gently deepening towards the east. The Moho in the Adriatic domain is deeper [$> 30 \text{ km}$: Locardi and Nicolich (1988)], gently west-dipping. In addition, seismic refraction detects a relatively slow

Moho (7.7 km/s) in the Tyrrhenian domain and a faster Moho (about 8.0 km/s) in the Adriatic domain. Seismic refraction data (Suhadolc and Panza, 1989; Ponziani *et al.*, 1995; De Franco *et al.*, 1998) localize a doubling of the crust for about 30 km, below the Val di Chiana-Val Tiberina region. The Moho doubling is poorly imaged by the CROP03 NVR profile, which highlights the presence of a highly reflective window, which includes some diffraction hyperbola down to a depth of 10 s; in this region a major, west-dipping crustal shear zone can be hypothesized.

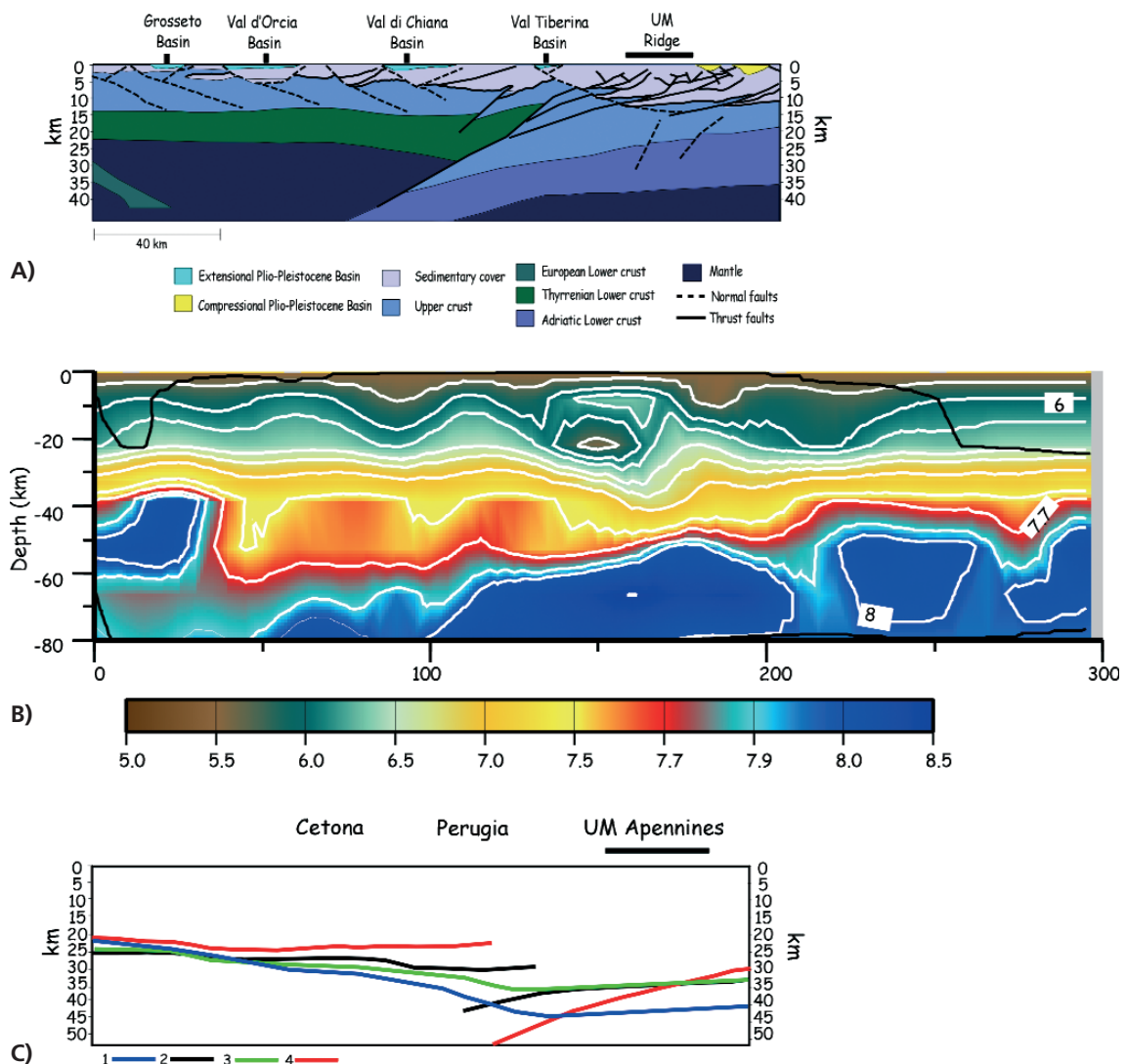


Fig. 2 - A) Geological interpretation of the seismic reflection line CROP03 [modified after Barchi *et al.* (1998a)]. B) Vertical section through the 3D P-wave regional velocity model (Di Stefano *et al.*, 2005) cut approximately along the CROP03 NVR profile. White lines = Vp contouring; black solid line = inverted volume; colour palette indicates absolute Vp values in km/s. C) Depth of the Moho discontinuity, along the profile b of Fig. 1, provided by: 1) Morelli *et al.* (1967); 2) Ponziani *et al.* (1995); 3) Piana Agostinetti *et al.* (2002); 4) Mele and Sandvol (2003).

The structure and geometry of the deep crust and of the upper mantle have been also imaged by teleseismic data (Piana Agostinetti *et al.*, 2002; Mele and Sandvol, 2003). The Moho signature from receiver function analyses is consistent with active seismic data for most of the profile (Fig. 2c) but a slightly different depth for the Adriatic Moho below the main ridge of the Apennines is obtained, e.g. > 50 km in Mele and Sandvol (2003) and about 35 km in Piana Agostinetti *et al.* (2002): the latter value is more consistent with the results of seismic refraction (Ponziani *et al.*, 1995; De Franco *et al.*, 1998).

Just below the crustal doubling detected by the refraction, tomographic images obtained by using different methods (Amato *et al.*, 1993; Spakman *et al.*, 1993; Castellarin *et al.*, 1994; Ciaccio *et al.*, 1998; Lucente *et al.*, 1999; Lucente and Speranza, 2001; Piromallo and Morelli, 2003) all show a pronounced high-velocity anomaly, interpreted as a submerged crustal slab, representing the westward continuation of the Adriatic Moho. The shape and location of the imaged high velocity slab are slightly different, reflecting difference in model resolution obtained by using local, regional and teleseismic data. In particular, slab continuity is still debated. Lucente *et al.* (1999) and Piromallo and Morelli (2003) define a continuous slab down to 670 km depth, while Spackman *et al.* (1993) detected relatively low velocities between 150 and 200 km and interpreted these values as related to a slab detachment.

The anomalous high S-wave velocity at depths exceeding 50 km confirmed the presence of lithospheric roots below the sector from Val di Chiana to Val Tiberina (Calcagnile and Panza, 1981; Du *et al.*, 1998; Pontevivo and Panza, 2002). These roots are also marked by deep crustal and subcrustal seismicity, extending to a depth of about 100 km, first described by Amato and Selvaggi (1991) and confirmed by the recent review of available instrumental data (Chiarabba *et al.*, 2005). However, seismicity data do not image a clear west-dipping Wadati-Benioff plane.

Both the seismicity and the high-velocity anomaly are strictly limited to the Northern Apennines, and do not appear below the Central and the Southern Apennines. A similar, more pronounced feature exists below the Calabrian arc, where the seismicity reaches a much greater depth [about 400 km: Selvaggi and Chiarabba (1995), Chiarabba *et al.* (2005) among many others].

A P-wave velocity structure of the Northern Apennines has been recently obtained from the tomographic inversion of the larger high quality data set available for the Italian region (see Di Stefano *et al.*, 2005). A section of Fig. 2b lays approximately along the CROP 03 NVR profile. The black solid line encloses the inverted area. The lithosphere of the Adriatic domain is imaged as a high-velocity body, plunging toward the south-west beneath the Tyrrhenian lithosphere. The Moho is located more or less in correspondence with the 7.7 km/s to 8.0 km/s velocity gradient of the tomographic image, as indicated by DSS studies (De Franco *et al.*, 1998). The base of the Adriatic lower crust is imaged deflected from an about 35 km depth to about 50 km supporting results from Barchi *et al.* (1998a), Piana Agostinetti *et al.* (2002), and Mele and Sandvol (2003). A large volume of the low-velocity anomaly (V_p between 7.0-7.7 km/s) is found in the Tyrrhenian domain between a 30 km and 60 km depth, extending down to the main ridge of the Apennines (Di Stefano *et al.*, 2005). This feature is interpreted as an intruding body of hot asthenosphere, caused by the detachment of the subducting lower Adriatic lithosphere from its buoyant crust around the Moho level, coherently with models proposed for the geodynamic setting of the Northern Apennines [Doglioni *et al.* (1999) among many others].

In this scenario, the intruding asthenosphere is probably the reason for strong physical and

chemical modifications at the Adriatic crust's bottom. Hence, a valuable hypothesis is that the Tyrrhenian Moho identified at 24-26 km and characterized by a lower seismic velocity contrast than the Adriatic Moho, might have been newly generated by the contact between the upwelling asthenosphere and the delaminated Adriatic crustal material partly modified by asthenospheric thermal effects (Barchi *et al.*, 1998a; Di Stefano *et al.* 2005).

3. The tectonic evolution of the Northern Apennines

The Northern Apennines are commonly interpreted as the result of the convergence between the already formed Alpine orogen and the continental crust of the Adriatic promontory of the African plate (e.g. Reutter *et al.*, 1980; Alvarez, 1991; Doglioni *et al.*, 1998). In this view, the Burdigalian to present tectonic evolution of the Northern Apennines, on which this paper focuses, is superimposed on previous compressional events, related to the formation of the Alpine orogen (Eocene), and to the consumption of the oceanic lithosphere of the Ligure-Piemontese Ocean (Upper Oligocene-Lower Miocene), simultaneous to the rotation of the Corsica-Sardinia micro-continent.

Compression moved eastwards throughout the Neogene-Quaternary times. The present-day activity along the easternmost compressive front is still widely debated [e.g. Lavecchia *et al.* (1994, 2004), Frepoli and Amato (1997), Di Bucci and Mazzoli (2002) this problem is furtherly discussed in section 6.3]. Contemporaneously, extensional deformation, related to the opening of the northern Tyrrhenian Sea, propagated to the western side of the Apenninic belt, starting in the Late Burdigalian in the Corsica basin (Bartole *et al.*, 1991), and is presently affecting the axial ridge of the Apennines (Lavecchia *et al.*, 1984, 1994; Montone *et al.*, 2004).

The best evidence of the contemporaneous, eastward migration of coupled compressional and extensional belts is provided by the eastward decreasing ages of the syntectonic basins, generated by compression in the foreland and by extension in the hinterland. Compression and extension are co-axial, i.e. the direction of maximum extension is W-E oriented in the northern Tyrrhenian Sea and WSW-ENE in the Tuscan mainland, nearly parallel to the maximum shortening induced by the previous compression.

3.1. The migrating foreland basins

As Merla (1951) early on recognized, the eastward moving Northern Apennines has generated progressively younger, foreland basins [foredeep and piggy back basins: Ori *et al.* (1986)], that were successively incorporated in the collision zone since Late Oligocene: at present, the Adriatic Sea represents the younger and easternmost foredeep active until at least the early Pleistocene. The eastward migration of the "flexural belt" is also effectively described in Casero (2004).

These basins completed the development of the Umbria-Marche thrust and fold belt, that represents the outer part of the Northern Apennines (Costa *et al.*, 1998; Barchi *et al.*, 2001). Since these basins were subsequently involved in the compressional deformation, they do not necessarily preserve their original shape and size, and their sediments mostly crop out in the syncline areas. Moreover, the original distance between the basins has been reduced by orogenic contraction.

The diagram in Fig. 3 summarizes available data about the ages of the turbidites infilling the

post-Burdigalian, syntectonic foreland basins, along a transect from Perugia to Ancona, grossly corresponding to the CROP03 profile, thus illustrating the overall eastward migration, through time, of the compressional deformation. The diagram also shows the time interval of activity of the major thrusts, which the foreland basins are related to.

The basins can be divided into three major groups, corresponding to, from west to east: i) the Marnoso-Arenacea foredeep, located in the Umbria pre-Apennines; ii) the Inner Marche thrust-top basins, located over the main ridges of the Umbria-Marche Apennines; iii) the Marche-Adriatic foredeep, located in the Outer Marche Foothills, east of the main mountain belt.

1. The Marnoso-Arenacea Fm. is deposited in a “complex foredeep” [sensu Ori *et al.* (1986)] presently cropping out between the eastern front of the Tuscan Units (near Perugia) and the western boundary of the main ridge of the Apennines, in the Late Burdigalian – Late Serravallian time interval. The onset of the foredeep is driven by the deformation of the innermost Tuscan units, that later overthrust the inner part of the Marnoso-Arenacea basin (Barchi *et al.*, 1998b), whilst the start of activity of the westernmost thrust, involving the Umbria-Marche units, occurred during the Late Serravallian (Brozzetti *et al.*, 2002). Tectonic activity during the Late Serravallian is also suggested by the presence, within the Marnoso-

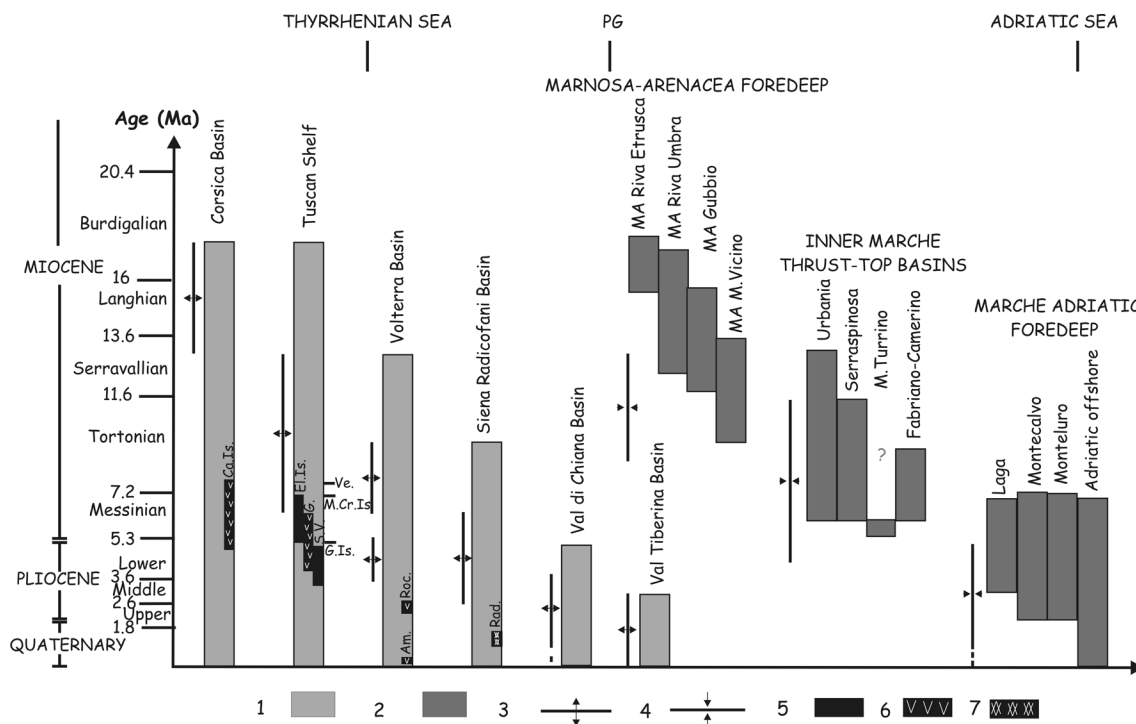


Fig. 3 - Age of syntectonic 1) hinterland and 2) foreland basins; 3) period of extensional phase; 4) period of main compressional phase (data from seismic reflection profiles: Barchi *et al.* (1998a, 1998b); Pascucci *et al.* (1999). 5) plutonites, 6) vulcanites and 7) subvolcanic rocks. Ca.Is.: Capraia island vulcanites (7.6-4.7 Ma); El.Is.: Elba island, (7-5 Ma); G.: Campiglia-Gavorrano (6-4 Ma); S.V.: San Vincenzo (4.9-3.2); Ve.: Vercelli plutons (7.3-6.2 Ma); M.Cr.Is.: Montecristo island (7 Ma); G. Is.: Giglio island (5 Ma); Am.: Amiata vulcanites (0.4-0.2 Ma); Roc.: Roccastrada vulcanites (2.6-2 Ma); Rad.: Radicofani sub-volcanites (1.3-0.9 Ma) [after Poli (2004), ages after compilation of Serri *et al.* (2001)].

Arenacea succession, of olistostromes and proximal fan deposits, as well as of coeval, slumping phenomena (Ricci Lucchi and Piali, 1973; Ridolfi *et al.*, 1995). At present this formation is involved in four main tectonic units, consisting of progressively younger successions, moving from west to east: the Riva Etrusca unit (west of the Tiber Valley, Burdigalian to Langhian), the Riva Umbra unit (from the Tiber Valley to the Gubbio anticline, Late Burdigalian to Middle Serravallian), the Gubbio unit (east of the Gubbio anticline, lower Langhian to Upper Serravallian), and the easternmost Mt. Vicino unit (Serravallian). The sedimentation is closed by the Lower-Middle Tortonian Mt. Vicino sandstones, a thrust-top basin, located over the easternmost part of the foredeep.

2. In the Tortonian - Messinian time interval, no proper foredeep basin exists, but only minor, relatively restricted, thrust-top basins [Cantalamezza *et al.* (1986), and references therein], whose genesis and evolution are related to the development of thrusting in the Umbria pre-Appennines, started in the late Serravallian (Brozzetti *et al.*, 2002), and to the early folding of the Umbria-Marche ridge (Alvarez, 1999). Most of these basins (Urbania, Serraspina, Mt. Turrino, Fabriano, Camerino) are located above the main ridge of the Umbria-Marche Apennines, or immediately to the west (Mt. Vicino) and to the east (Laga) of the mountain belt.
3. The Outer Marche-Adriatic syn-orogenic succession is deposited in a foredeep in the Messinian-Quaternary time interval. The onset of this basin is related to the major compressional phase of the Umbria-Marche Apennines (Late Messinian-Early Pliocene) and its subsequent tectono-sedimentary evolution is markedly influenced by the fact that in the late Messinian the peripheral bulge of the Adriatic foreland reached the outermost Dinarides and stopped migrating, so that the Marche-Adriatic foredeep was early transformed into a complex basin, comprising several syn-sedimentary folds and thrusts [complex foredeep, *sensu* Ori *et al.* (1986)] very early. The sedimentation is opened by the Messinian Laga Flysch, deposited in a quite large thrust-top basin, located in the westernmost part of the region, and continues through the Pliocene-Quaternary time interval. The syntectonic successions were later involved in the eastward prograding folds and thrusts, whose style and timing of deformation have been depicted in detail in many papers, combining surface geology, biostratigraphy and good quality seismic reflection profiles, calibrated by many boreholes (e.g. Argnani *et al.*, 1991; Ori *et al.*, 1986; Coward *et al.*, 1999).

3.2. The migrating hinterland basins

Simultaneously with the development of the above-described compression, at the back of the eastward-thrusting Northern Apennines, the Tyrrhenian domain was involved in extensional tectonics, continuously migrating through time from west to east, from eastern Corsica to the Tuscan mainland, and deeply dissecting the previously formed compressive structures. This process, firstly described by Elter *et al.* (1975), was successively confirmed by other research [e.g. Lavecchia *et al.* (1984, 1987), Carmignani *et al.* (1994) among many others].

In strict analogy with the compressional history, the best evidence of the migrating extension is furnished by the onset of the extensional hinterland basins (Fig. 3), whose age systematically decreases from the Corsica basin through the Northern Tyrrhenian Sea and Tuscan mainland, to reach the Umbria-Marche Apennines ridge in Lower Pleistocene, where extension is presently active (Bartole *et al.*, 1991; Jolivet *et al.*, 1998; Pascucci *et al.*, 1999; Collettini *et al.*, 2006).

The results of the CROP 03 project (Fig. 2a) produced significant advances in the knowledge of the extensional process, revealing:

- the strong asymmetry of the extensional deformation, which appears dominated by a set of east-dipping, low-angle normal faults;
- the depth of the low-angle detachment, which is located at approximately 5 s (i.e. about 12 km);
- the existence of a still active, east-dipping normal fault (Alto-Tiberina Fault, ATF), affecting the westernmost part of the Adriatic domain.

The upper crust of Tuscany and western Umbria was thinned by a set of six major, east-dipping low-angle normal faults (Pialli *et al.*, 1998; Decandia *et al.*, 1998) whose location is strictly connected with the position of the shallow marine and/or continental syn-tectonic basins, from west to east: the Punta Ala; Grosseto; Val d'Orcia; Siena-Radicofani; Val di Chiana and Tiber basins (Fig. 2a). The age of the syn-rift deposits, infilling these "hinterland" basins (Fig. 3), testifies the regular eastward migration of the extensional deformation (e.g. Pascucci *et al.*, 1999; Collettini and Barchi, 2004).

Further images of these basins and of the extensional faults driving their development have also been provided by many seismic reflection profiles, located on both the northern Tyrrhenian Sea and the Tuscan mainland (e.g. Bartole *et al.*, 1991; Pascucci *et al.*, 1999; Cornamusini *et al.*, 2002). Shallow extensional faults, producing the direct superposition of younger over older rocks, exhumed by a subsequent uplift, have been also mapped at the surface [e.g. Zuccale fault: Keller *et al.* (1994), Collettini and Barchi (2004)] and/or drilled by deep wells in the geothermal areas of Larderello (Batini *et al.*, 1985; Brogi *et al.*, 2003) and Mt. Amiata (Calamai *et al.*, 1970; Brogi, 2004).

Along the CROP 03 profile, the low-angle normal faults are detached in the upper crust, reaching a depth of about 5 s (approximately 12 km) (Fig. 2a). This is the same depth of both the B/D transition inferred from the thermal field (Pauselli and Federico, 2002) and the cut-off of seismicity (Chiarabba *et al.*, 2005). The underlying lower crust shows an anomalously reduced thickness (about 9 km), possibly produced by intense ductile flow. Summarising, the Tyrrhenian domain has probably been extended by brittle/semi-brittle fault zones in the upper crust, accomplished by prevailing ductile flow in the lower crust.

In the Northern Apennines, the younger and easternmost expression of extension (ATF) is exposed in the Umbria region, at the western border of the Adriatic domain (Boncio *et al.*, 1998, 2000; Collettini and Barchi, 2002). This low-angle normal fault cuts the thickened crust down to a depth of about 14 km, greater than that reached by the older, no longer active and previously mentioned faults of the Tyrrhenian domain, reflecting incipient extension. The relationships with the pre-existing compressional structures and the age of the syn-tectonic Val Tiberina basin pin the fault activity to the Upper Pliocene-present time interval. Strong evidence for the present-day activity of the ATF are furnished by microseismicity surveys, operated by local networks, along the fault trace. Two temporary microseismic networks operated in the Città di Castello-Perugia area in two distinct time periods (Deschamps *et al.*, 1989; Piccinini *et al.*, 2003). The two surveys show similar characters: most of the seismicity fits the trace of the ATF, as depicted by depth-converted seismic profiles, separating an active hangingwall from an almost aseismic footwall (Boncio *et al.*, 1998); the focal mechanisms show extensional kinematics with NNW-SSE trending planes, parallel to the ATF, and the resulting stress tensor (with a vertical σ_1 and a ENE

trending σ_3) is consistent with both the presently active stress field (Montone *et al.*, 2004) and the Quaternary long-term stress field (Lavecchia *et al.*, 1994; Boncio *et al.*, 2000) of the region.

In conclusion, the western part of the Northern Apennines has been dominated by extensional tectonics, producing hinterland basins since Miocene. The amount of lithosphere thinning distinctly recognized by the CROP 03 profile and testified by other geophysical evidence (such as high surface heat flow, magmatism, positive Bouguer anomalies) and the presence of metamorphic cores [e.g. Apuane Alps: Carmignani and Kligfield (1990)] both highlights the fact that the Tuscan region possesses all the typical characters of an extensional belt, as defined by Lister and Davis (1989).

3.3. The migration of magmatism

Magmatic activity also migrates eastwards, following the formation of the extensional basins. The oldest magmatic activity is found in eastern Corsica with the Sisco lamproitic sill (15-13.5 Ma) followed by the Montecristo Island (7 Ma), Elba Island (7-5 Ma), Vercelli plutons (7.3-6.2 Ma) and Capraia Island vulcanites (7.6-4.7 Ma). Emplacement of acid intrusive bodies is documented all over the Tuscan-Latium continental shelf [Porto Azzurro, Giglio Island (5 Ma), Campiglia-Gavorrano (6-4 Ma), Castel di Pietra, Monteverdi, etc.: Serri *et al.* (1993, 2001), Poli (2004), and references therein] followed by the Roccastrada vulcanites (2.6-2 Ma), Radicofani sub-volcanites (1.3-0.9 Ma) and Amiata vulcanites (0.4-0.2 Ma). The diagram of Fig. 3 shows that the magmatic bodies are emplaced late in the extensional history, after the major rift phase, possibly because the emplacement of the magmatic bodies requires that a significant crustal extension already occurred.

4. Seismicity and stress maps

Seismological data (distribution of seismicity and focal mechanisms) strongly support the contemporaneous activity of extension and compression. This concept is the base of the seismotectonic zoning proposed by Lavecchia *et al.* (1994) and Boncio and Lavecchia (2000): it was reprised by Frepoli and Amato (1997), presenting a systematic collection of the focal mechanisms documenting this phenomenon.

The pattern of instrumental seismicity, recorded in the Northern Apennines in the last 20 years (Chiarabba *et al.*, 2005) shows the presence of three distinct seismotectonic zones (Fig. 4). The western and the eastern zones correspond to the Tyrrhenian and to the Adriatic domains respectively, whilst the intermediate zone, located along the Apennine belt, where most of the seismicity is clustered, grossly corresponds to the border between the two crustal domains. Similar relationships between the character and magnitude of the observed seismicity, the crustal structures, and the Late Miocene to Present tectonic evolution are described by Ghisetti and Vezzani (2002).

In the Tyrrhenian domain, microseismicity and small magnitude earthquakes occur at shallow depths within the seismogenetic layer whose thickness is reduced mostly due to geothermal and post-magmatic processes active in the area (e.g., Liotta and Ranalli, 1999; Pauselli and Federico, 2002).

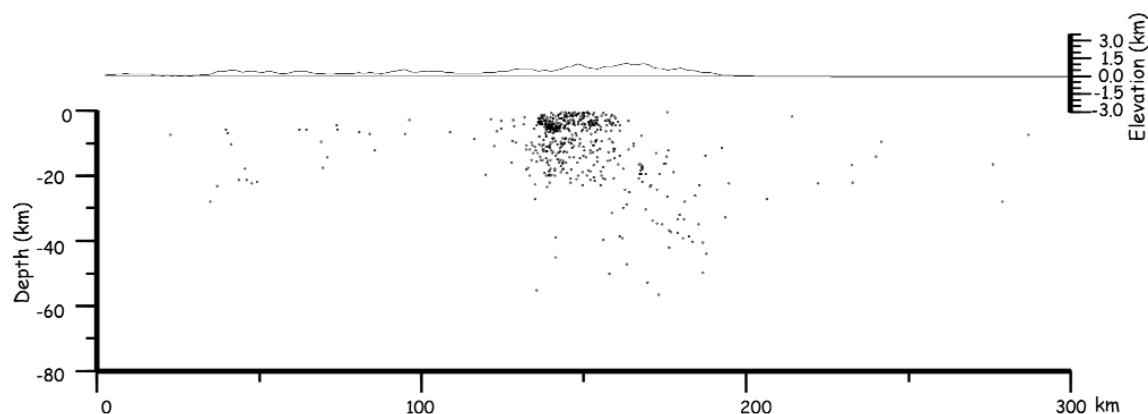


Fig. 4 - Hypocenter distribution along the tomographic profile of Fig. 2. The topography is also reported. Note different vertical scale is adopted for the topographic profile and for the seismicity section.

The Apennine belt is characterized by moderate to large normal-fault earthquakes ($M < 6.0$), mostly occurring on NW-trending SW-dipping, 5 to 10 km long, adjacent segments (Amato *et al.*, 1998, Chiaraluce *et al.*, 2003). The contiguity of these segments delineates an elongated extensional belt that crosses pre-existing compressional structures and is mostly confined in the upper 6-8 km of depth. Recent surveys reveal the existence of an east-dipping low-angle seismic detachment co-located with the formerly recognized ATF (Piccinini *et al.*, 2003). The low-angle detachment accommodates the extension and is presumed to drag the SW-dipping segments on its hangingwall.

To the east of the normal fault belt, a deep crustal microseismicity has been recognized (Chiarabba *et al.*, 2005). Earthquakes occur within the lithosphere which is probably related to a different amount of retreating, but the kinematics of this region is still unclear. Shallower, small earthquakes can occur in the Adriatic region (Ancona, 1972; Porto San Giorgio, 1987), but we lack information to constrain their kinematics and seismotectonics.

Focal mechanisms of the largest earthquakes (Chiaraluce *et al.*, 2003, Chiarabba *et al.*, 2005; Lavecchia *et al.*, 2003b) and borehole break-outs, confirm that a ENE extension is presently active in the western sector, whilst co-axial, active contraction is registered in the Po Plain-Adriatic foreland [Montone *et al.* (2004) and references therein; Chiarabba *et al.* (2005): Fig. 6].

5. The three rules of the Northern Apennines

The time and space relationships between compression and extension, highlighted by the geological data about the age of the syn-tectonic basins (long-term image) as well as by the instrumental seismicity pattern (short-term image) can be expressed through three main rules, which are illustrated in the following.

5.1. The first rule: at any time compression and extension are contemporaneously active, in the foreland and in the hinterland, respectively

The best evidence is the distribution and age of the syn-tectonic basins of the Apennines, already discussed in chapter 3 (Fig. 3). For example, in late Tortonian-early Messinian age (about

9 to 6 Ma) extensional deformation affected both the offshore Tuscan shelf and inland western Tuscany, where syn-rift sediments are preserved in the extensional basins of Grosseto and Volterra (Pascucci *et al.*, 1999). In the same time interval, active contraction affected the Umbria pre-Apennines, involving the already deposited successions of the Marnoso-Arenacea Fm., and produced early folding and thrusting of the axial region of the Umbria-Marche Apennines, where a set of minor, satellite basins [inner Marche thrust-top basins: Cantalamessa *et al.* (1986)] developed. Some million years later, in Early Pleistocene (about 2 Ma), the axial ridge was reached by extensional tectonics, generating the intramountain extensional basins of Gubbio, Colfiorito and Norcia, whilst contraction affected the coastal Adriatic region, where clastic foreland sedimentation occurred.

Considering the present-day tectonic activity, the Northern Apennines are characterised by extensional seismicity in the axial zone of the Apennines and compressional earthquakes at the Adriatic coast. This pattern is also consistent with other stress indicators, recently resumed in the new stress map of Italy (Montone *et al.*, 2004) and with the strain measures performed by GPS (Battaglia *et al.*, 2004).

5.2. The second rule says that at any position, extension postdates compression, and describes how the tectonic regime varies with time

This is a very common experience of Apennine geologists: in fact, as we move across the inner part of the Northern Apennines, from the Tyrrhenian coast to the axial zone of the main mountain range, in any place we find folds and thrusts, displaced by later normal faults. For example, we can consider the Gubbio region, located in northern Umbria [Collettini *et al.* (2003) and references therein]: here, in the Upper Miocene, the Marnoso-Arenacea foredeep basin was formed, closely followed by the onset and growth of the Gubbio anticline. Later (in the Quaternary) the same region was reached by the extensional tectonics (Fig. 5), generating the NW-SE trending Gubbio normal fault, possibly still active, disrupting the inner limb of the Gubbio anticline and bordering the hinterland (i.e. extensional) Gubbio extensional basin (Menichetti, 1992; Pucci *et al.*, 2003; Mirabella *et al.*, 2004).

The presence of an active extensional belt, grossly corresponding to the main ridge of the Apennines, is now generally recognised (e.g. Barchi *et al.*, 2000; Chiaraluce *et al.*, 2003), as confirmed by many different data sources (i.e. geological, seismological, geophysical, geomorphological, etc.). In particular, seismological data univocally indicate active extension at upper crustal levels (Chiarabba *et al.*, 2005). In this region, the presently active extensional faults systematically displace older (Upper Miocene) folds and thrust faults.

5.3. The third rule says that extension can occur in the uppermost crust at the same time and position of compression at greater depth (lower crust and upper mantle)

This effect is due to the fact that both compressional and extensional deformation are markedly asymmetric: in fact, the major compressional faults are west-dipping thrusts, and the major extensional detachments are east-dipping low-angle normal faults. The overall geometry is illustrated in the relevant section of Fig. 2a. In the same position, where the easternmost east-dipping normal master fault is now active at the surface (e.g. ATF), at greater depth (>20 km) we can have an active west-dipping thrust fault, that reaches the surface only in an outermost position (e.g. close to the Adriatic coast).

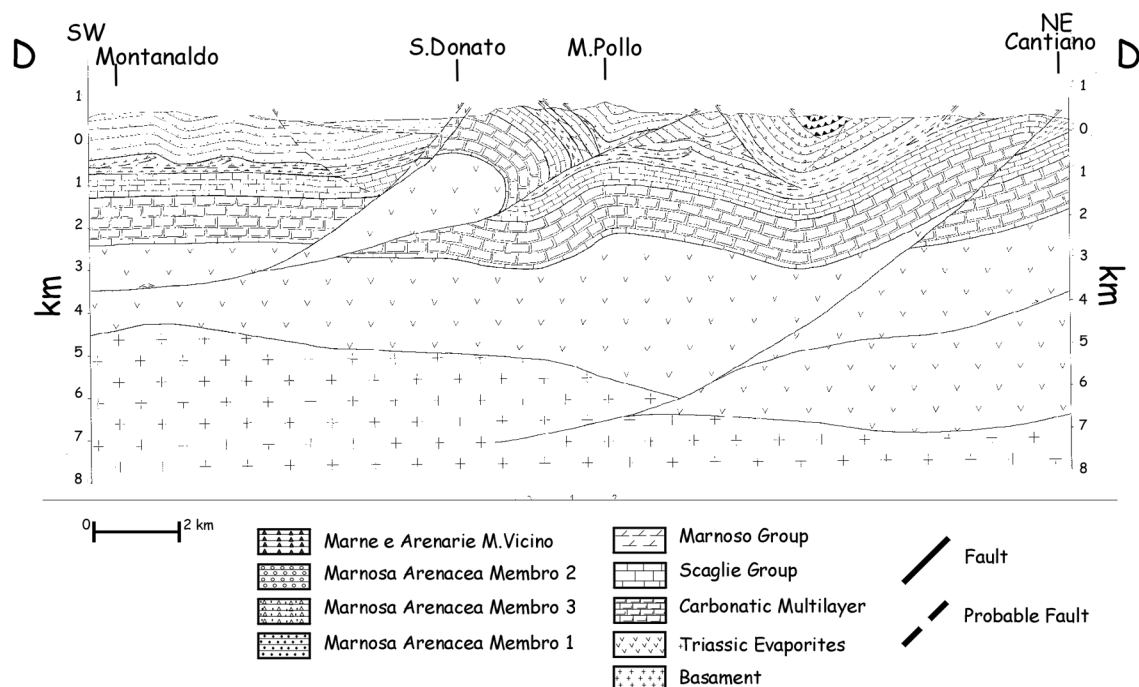


Fig. 5 - Geological section of the seismic profile D-D'. See location in Fig. 1A (after Ciculo, 2003).

Below the axial zone of the Apennines, we have shallow, extensional earthquakes, and a deeper, compressional (?) seismicity. The pattern of seismicity comprises two main clusters (Chiarabba *et al.*, 2005): a shallow, upper crustal extensional seismicity (depth < 12 km), generally deepening towards the east and a deeper cluster (depth > 20 km), where some focal mechanisms indicate compressional kinematics.

It is possible to argue that this mechanism was active also in the past, as postulated by Pialli *et al.* (1995): the extensional detachments are located at upper crustal levels, at the hanging wall of a deep thrust, which is active at lower crustal or deeper levels, at the hanging wall of the retreating Adriatic lithosphere. In this sense, the Apennine extension can be defined as “syn-orogenic” (Jolivet *et al.*, 1998; Rossetti *et al.*, 1999).

These rules are strictly geometrical and they are valid independently from the adopted geodynamic model: e.g., the major west dipping, compressional shear zones can be interpreted either as crustal scale thrust zones (e.g. Lavecchia *et al.*, 2004), or as related to subduction of the Adriatic continental crust [Doglioni *et al.* (1999) among many others].

6. Discussion

In the following, we briefly take into account the recent scientific debate on the tectonic evolution of the Northern Apennines, focusing on the topics, more directly related to the co-

existence and eastward migration of the extensional and compressional field. Three main topics are discussed here:

1. some authors deny the extensional nature of the hinterland basins, proposing a quasi-continuous, eastward thrusting of the Apenninic crust;
2. those Authors who accept the extensional nature of the hinterland basins, propose different ideas, on the mechanisms driving extension and the time-space relationships between low-angle and high-angle faults;
3. present-day activity of the compressional external front in the foreland region (Po-Plain Adriatic Sea) is not univocally constrained and not accepted by all the researchers.

6.1. Compressional hinterland basins

Finetti *et al.* (2001) have proposed a different interpretation of the CROP03 profile, suggesting a dominant tectonic style of compression throughout the whole Northern Apennines, minimizing the importance of extensional deformations in the western Tyrrhenian domain. This interpretation is supported by a set of regional and local studies (e.g. Boccaletti *et al.*, 1997; Bonini *et al.*, 1999; Bonini and Sani, 2002), where hinterland basins of Tuscan mainland and western Umbria are interpreted as “piggy back” basins, whose location, genesis and evolution are driven by thrust tectonics. In this interpretation, the normal faults are second-order accommodation structures, related to either collapse of the back-limb of the thrust anticlines or to local transtension along strike-slip, transversal fault zones. Extensional tectonics would have started only recently (about 2 Ma), affecting the already formed, syn-compressional hinterland basins with high-angle geometry faults.

This interpretation is in contrast with several data, collected and interpreted by different Authors who, aside different possible interpretations, all agree that extensional tectonics have dominated the evolution of the Tyrrhenian domain since the end of the Lower Miocene. In the following, we briefly summarize these evidences:

- the thinned crust existing below the Tyrrhenian domain;
- the very high heat flow measured in the Tuscan region, which requires an advective process to allow the rapid propagation of the thermal anomaly from the asthenosphere to the surface;
- the exhumed, low-angle normal faults, that have been observed and studied in detail on the Island of Elba, as well as below the Larderello geothermal fields;
- the onset and evolution of the hinterland basins, driven by major normal faults, possessing displacements in the order of thousands of meters, extensively documented by both regional studies, mainly based on seismic reflection profiles (e.g. Bartole *et al.*, 1991; Pascucci *et al.*, 1999) and by surface geology and borehole data (Liotta, 1996; Collettini *et al.*, 2006);
- the absence of regionally documented thrusts that unambiguously cut recent deposits, but do not control sedimentation; only minor (outcrop scale) structural compressional features have been observed in these deposits, which can be easily explained as minor features, related to the complex geometry of the normal faults (hanging-wall anticlines and ramp synclines) or as minor, compressional faults, located along transversal transfer faults (Decandia *et al.*, 1994), controlling the basin/normal faults segmentation.

6.2. Modes of extension and hinterland basins formation

Among the Authors who accept the extensional nature of the hinterland basins, different ideas

have been proposed, about the mechanisms driving extension and exhumation of metamorphic rocks in the Tyrrhenian domain, as well as the time and space relationships between low-angle and high-angle faults.

Different models have been proposed to explain the exhumation of the metamorphic cores in the Apuane Alps (Carmignani and Kligfield, 1990), Corsica (Jolivet *et al.*, 1998), Giglio Island (Rossetti *et al.*, 1999) and other areas in Tuscany (Jolivet *et al.*, 1998). Carmignani and Kligfield (1990) used field-data from the Apuane core complex to suggest that since the Middle Miocene, extension occurred during bulk coaxial thinning of the crust during post-orogenic collapse of the overthickened orogenic wedge. Extension was thought to be characterised by two episodes: in the first stage (Early-Middle Miocene), the development low-angle normal faults by reactivation of pre-existing thrusts and other structures led to core complex exhumation. In the second stage (Late Tortonian-Quaternary), high-angle normal faults cut and displaced the previously developed extensional structures, forming extensional basins. This exhumation style requires a high temperature regime. More recently, Jolivet *et al.* (1998) and Rossetti *et al.* (1999) have used peak pressure estimates (10-14 kbar and 350°C), derived from carpholite discovered in some metamorphic cores, to propose a two-stage process: a first exhumation during crustal thickening in a HP/LT regime ("syn-orogenic extension") followed by LP/HT phase of back-arc extension ("post-orogenic extension"). These two extensional phases were thought to be separated by minor, compressional pulses (Storti, 1995; Rossetti *et al.*, 1999).

Apart from the localized regions in where deeply exhumed rocks (30-40 km) have been discovered, extensional models do not have to account for a great deal of exhumation. This has led to the emergence of an alternative model, based on a continuous, where an eastward migration of extension occurs (e.g. Lavecchia 1988; Barchi *et al.*, 1998a; Doglioni *et al.*, 1999). Such models are consistent with the westerly-directed Apenninic subduction zone and result in a low elevation, single-vergence mountain chain, with deep foredeep deposits and a back-arc basin (Doglioni *et al.*, 1999).

Both Carmignani *et al.* (1994) and Jolivet *et al.* (1998) proposed polyphasic modes of extension, where east-dipping LANFs (low angle normal faults) are generated during an earlier phase and are successively displaced or reprinted by later, post-orogenic faults. These two phases are related to different causes and are separated by a tectonic pause or even by minor compressional pulses. This concept has been extensively applied to the evolution of the hinterland basins of Tuscany (e.g. Baldi *et al.*, 1994; Liotta, 1996), where two or even three different stages of extension have been recognised.

On the other hand, Collettini *et al.* (2006) have postulated a monophasic extensional deformation, where east-dipping LANFs are accomplished by high-angle, synthetic (east-dipping) and antithetic (west dipping) normal faults, generated at their hanging wall during a single, more or less continuous phase of crustal extension. This model is mainly based on the study of the easternmost LANF (ATF), where a complex data set of detailed surface geology data (Minelli and Menichetti, 1990; Brozzetti, 1995), seismic reflection profiles (Boncio *et al.*, 1998; Barchi *et al.*, 1999) and accurate microseismic survey (Piccinini *et al.*, 2003) show that the ATF is presently active, at the same time as the high-angle, synthetic (east-dipping) and antithetic (west-dipping) normal faults, generated at its hanging wall.

6.3. Activity of the external front of the Apennines

The present-day activity of the compressional front, in the foreland, Po Plain-Adriatic region is not univocally constrained and not accepted by all the researchers. Di Bucci and Mazzoli (2002) put in evidence that the focal mechanisms of the earthquakes do not univocally constrain the activity of the compressional front. Moreover, the interpretation of seismic reflection profiles, in the off-shore northern Adriatic, suggests that thrusting and folding ended in Early Pleistocene times. They conclude that the Adriatic lithosphere is no longer retreating beneath the chain and that the present-day tectonic setting is dominated by tectonic uplift, related to the slab detachment.

However, we note that the maps of the stress field (Montone *et al.*, 2004) and the strain GPS measurements (Battaglia *et al.*, 2004) both underline active contraction, perpendicular to the Northern Apennine Arc, also supported by other, independent geological, geophysical and geomorphological evidence (Burrato *et al.*, 2003; Benedetti *et al.*, 2003; Massoli, 2004). Moreover, the shallow compressional events, characterising the external area, could be geometrically connected, through a west-dipping major thrust, to the deep earthquakes, recorded in the lower crust-upper mantle, below the mountain range, as also suggested by Lavecchia *et al.* (2003b, 2004).

7. Conclusions

The marked geological and geophysical differences between the Tyrrhenian and Adriatic domains, the onset and evolution through time of the hinterland and foreland basins, the orientation of the present-day stress and strain fields, all underline the peculiar tectonic setting of the Northern Apennines, supporting the presence of contemporaneous, co-axial, eastward-migrating compressional and extensional coupled adjacent zones. Since 17 Ma, the previously shortened Tyrrhenian domain has been affected by an extensional regime, whilst the Adriatic domain has been subjected to contraction. In a same structural region, extension and compression are also contemporaneously active within the upper crust and at deeper crustal/subcrustal levels, respectively.

This process is effectively imaged by the distribution and characteristics of instrumental seismicity, representing the present-day, instantaneous image of the tectonic evolution of the Northern Apennines. The major cluster of upper crustal seismicity marks the border between the extended (i.e. Tyrrhenian) and the not-yet-extended (i.e. Adriatic) domains: more precisely, it is located at the western border of the Adriatic domain. This seismicity, including the moderate ($5 < M < 7$) seismicity of the axial zone, represents the “noise” produced by the eastward migrating extension, propagating towards the foreland.

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