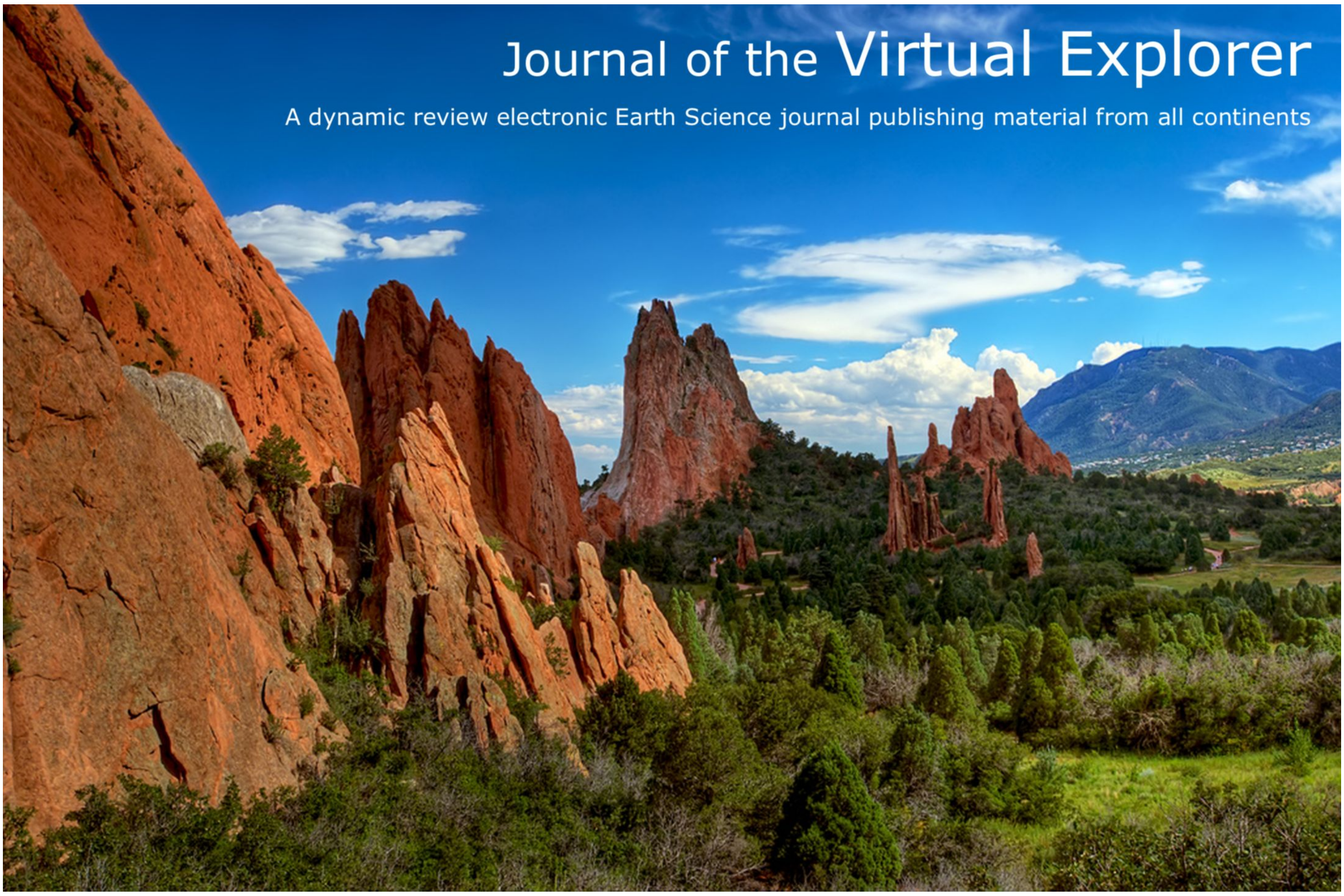

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The Neogene-Quaternary evolution of the Northern Apennines: crustal structure, style of deformation and seismicity.

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Abstract: Since Lower Miocene, the tectonic evolution of the Northern Apennines has been characterised by the contemporaneous activity and eastward migration of coupled compression (in the foreland) and extension (in the hinterland). Compression and extension are co-axial, i.e. the direction of maximum extension is nearly parallel to the maximum shortening induced by the previous compression.

As a result of this tectonic evolution, the Northern Apennines can be divided into two different crustal domains: a western Tyrrhenian domain, where extensional deformation destroyed the pre-existing compressional belt; and an eastern Adriatic domain where the compressional structures are still preserved.

The upper crust of the Tyrrhenian domain is thinned by a set of east-dipping low-angle normal faults, driving the onset and evolution of the syn-tectonic hinterland basins. The age of the syn-rift deposits testifies the regular eastward migration of the extensional deformation. The shallow structures of the Adriatic domain correspond to the arc-shaped Umbria-Marche fold and thrust belt where the timing of deformation is marked by the onset and evolution of syn-tectonic foreland basins.

The structural style of both compressional and extensional structures is strongly influenced by the mechanical anisotropy in the upper crust stratigraphy. Consequently, both thrusts and normal faults show marked staircase trajectories.

The present-day stress field (seismicity, boreholes break-out) as well as the on-going deformation (GPS data) of the region also reflect the contemporaneous activity of compression in the foreland and extension in the hinterland, supporting a uniformitarian view of the recent tectonic evolution of the Northern Apennines.

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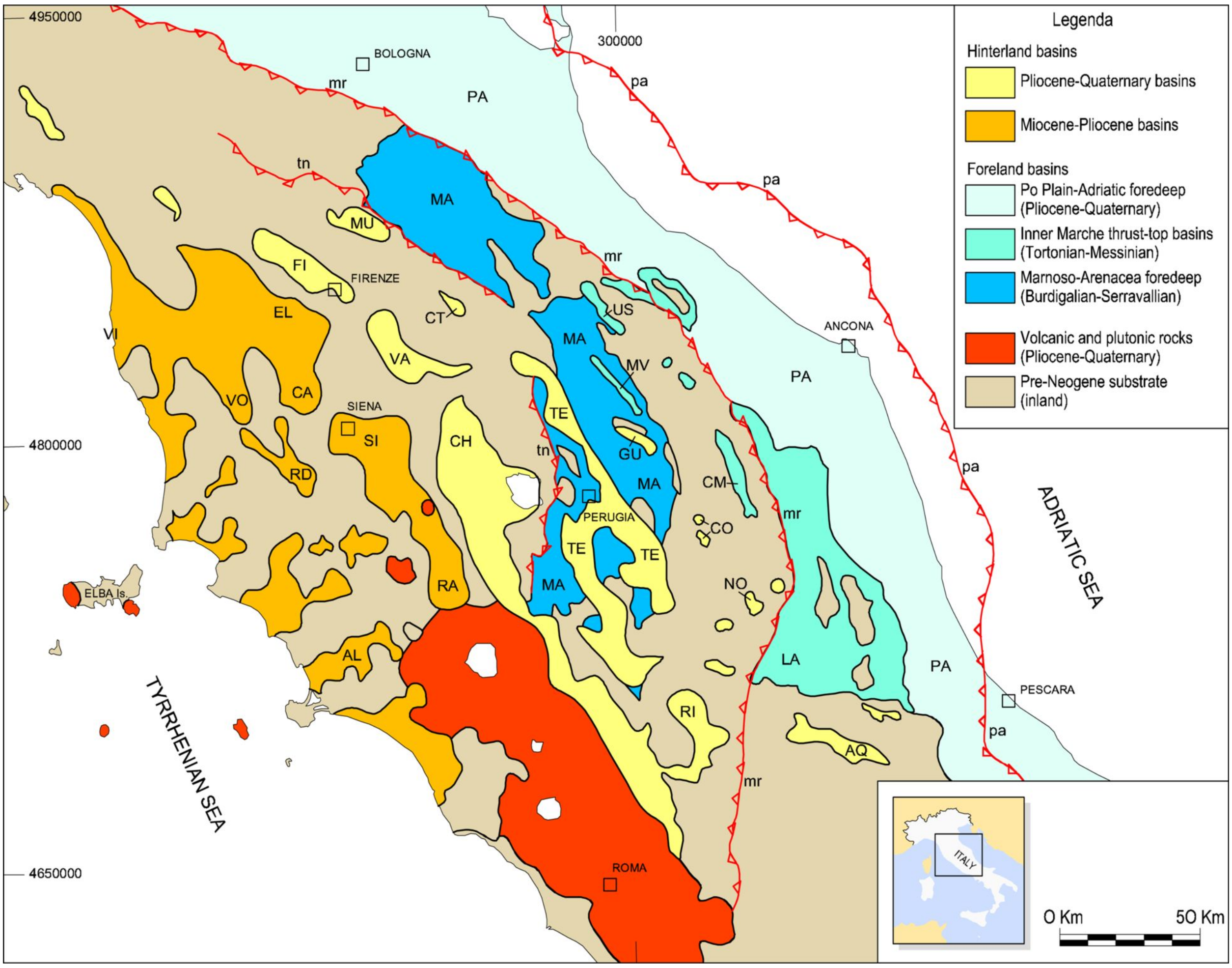
Chapter 1. Introduction

Location and limits of the study area

The Northern Apennines (NA) of Italy are an arcuate orogenic belt, formed within the convergent boundary between the European continental crust (the Corsica-Sardinia block) and Adria (considered either as a promontory of the African Plate or as an independent microplate). The NA are comprised between the Western Alps arc to the North and the Central-Southern Apennines arc to the South (fig. 1). From a geographical point of view, the on-land portion of the belt comprehends almost entirely the regions of Tuscany, Emilia-Romagna, Umbria and Marche, along with the NW portion of the Latium. The off-shore part comprehends the Northern Tyrrhenian Sea (from Corsica to the Tuscan coast), which is the hinterland of the NA (or of the Tyrrhenian-Apennines system); and the Northern Adriatic Sea, which is part of the present-day foredeep.

Figure 1. Geological sketch of the Neogene-Quaternary basins of the Northern Apennines.



Hinterland Basins - AL: Albegna; AQ: Aquila; CA: Casino; CO: Colfiorito; CH: Valdichiana; CT: Casentino; EL: Valdelsa; FI: Firenze; GU: Gubbio; MU: Mugello; NO: Norcia; PA: Punta Ala; RA: Radicofani; RD: Radicondoli; RI: Rieti; SI: Siena; TE: Valtiberina; VA: Valdarno; VI: Viareggio; VO: Volterra. Foreland Basins - CM: Camerino-Matelica; LA: Laga; MA: Marnoso-Arenacea; MV: Monte Vicino; PA: Po Plain-Adriatic; US: Urbania-Serraspinosa. Major Neogene-Quaternay thrusts - tn: Tuscan Nappe thrust; mr: Main Ridge thrust; pa: Po Plain-Ariatic thrust.

The NA 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; Doglioni *et al.*, 1998). In this view, the Late Miocene-Quaternary tectonic evolution of the NA, on which this paper focuses, is superimposed on previous compressional events, related to the formation of the Alpine orogen

(Cretaceous-Eocene), and to the consumption of the oceanic lithosphere of the Western Thetys (Late Oligocene-Early Miocene), simultaneous to the rotation of the Corsica-Sardinia microcontinent (Alvarez, 1972; Carminati *et al.*, this volume; Molli *et al.*, this volume).

As the other arcs of the Apennines and of the Mediterranean region (e.g. Lister *et al.*, 1994; Jolivet *et al.*, 1998b), the NA incorporate tectonic units derived from the Mesozoic Thetys ocean and the adjacent continental passive margins. Top to bottom (i.e. hinterland to foreland), they are: the Liguride units (mainly oceanic), the Tuscan and the Umbria-Marche units (both deposited on the continental passive margin of Adria). The pre-orogenic successions are unconformably overlain by syn-compressional units, mainly consisting of turbidite sandstones, deposited in foreland basins (foredeep and/or thrust-top basins): the eastward younging age of these basins marks the progression of the compressional deformation from the hinterland to the foreland. Both pre-orogenic and syn-orogenic successions are involved in the compressional belt, which is successively uplifted and partially eroded. In the western part of the NA the compressional structures are disrupted by later extensional faults, bordering hinterland basins, where continental and/or shallow marine clastic successions are deposited: the age of these syn-extensional units also becomes younger from SW to NE. Neogene magmatic activity accompanies and post-dates normal faulting. Both compressional and extensional structures are segmented by transverse faults, which were active simultaneously to the main structures they are related to. The origin, role and relevance of these transversal structures is still debated (e.g. Pascucci *et al.*, 2007).

Aim and outline of the paper

This paper is not a summary of the very complex stratigraphy and tectonics of the NA region (for a recent review of these aspects see Carmignani *et al.*, 2001; 2004; Barchi *et al.*, 2001). Instead, it is focussed on the peculiar characters of the recent (Neogene to present) tectonic evolution of the region, and how this evolution is reflected in both shallow and deep structures (as depicted by geomorphological, geological and geophysical surveys) and eventually in the present-day stress and strain patterns (depicted by seismological and geodetic data, respectively). The aim is to show that at least the first-order features of the NA can be framed into a relatively simple scheme, beyond many local variations and peculiarities.

These concepts are not new and some of them are remarkably old, based on the collection of extensive datasets, mainly performed by field geologists of the last century: successively, modern geophysical surveys and up-to-date geodynamic models have offered new technical and theoretical support to these ideas.

After this introduction, Chapter 2 deals with the deep (crust and mantle) structure of the NA, as imaged by a multiplicity of geophysical data, revealing the existence of two well-distinguishable domains, namely the Tyrrhenian and the Adriatic domains.

This crustal structure is the result of a peculiar tectonic evolution, characterised by the contemporaneous activity and eastward migration of coupled compression (in the foreland) and extension (in the hinterland). Chapter 3 summarises the sedimentological, structural, morphological and magmatological evidences constraining the nature and timing of this long-lasting geological process.

Chapter 4 will go deeper into some details of the deformation style of the upper crust, illustrating how the mechanical stratigraphy of the upper crust influenced the geometry and kinematics of both compressional and extensional structures.

Chapter 5 will go back to the present-day setting, analysing the presently active stress and strain field, here regarded as a snapshot of the present-day stage of the NA evolution. This snapshot is captured through the seismicity (both historical and instrumental) and the geodetic (GPS) data.

The final chapter 6 compares the long-term tectonic evolution of the NA and their present-day setting, in order to propose a uniformitarian model of the tectonic process which built up the NA, explaining the observed space (horizontal and vertical) and time relationships between compressional and extensional deformation. Finally, alternative geodynamic scenarios are briefly illustrated, where the NA evolution can be framed.

Chapter 2. Crust and mantle structure: Tyrrhenian vs. Adriatic domain

The NA can be divided into two different crustal domains, whose distinction is based on their peculiar geophysical and geological features: a western Tyrrhenian Domain (TD), where extensional deformation destroyed the pre-existing compressional belt; and an eastern

Adriatic Domain (AD), where the compressional structures are more recent and still preserved. The transition between the two domains is pretty sharp and occurs along a relatively narrow, arcuate strip, which is here referred to as the “transition zone” (TZ).

The TD and the AD are characterised by contrasting gravity anomalies and heat flow values, and show relevant differences in the crustal thickness and internal structure, revealed by both active and passive seismic surveys.

Gravimetric and heat flow data

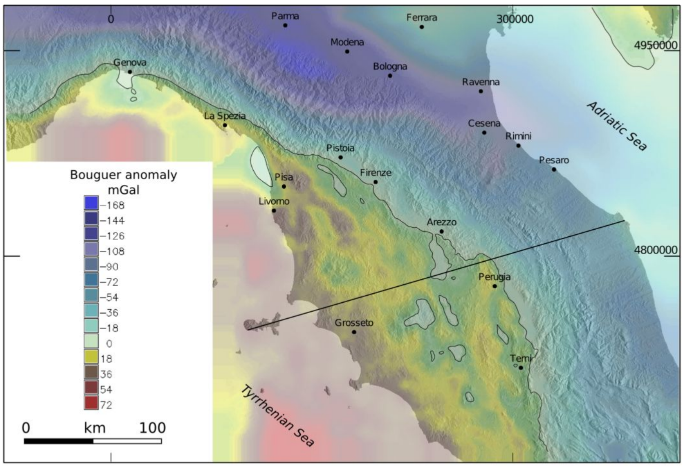
The gravity map (ISPRA, ENI, OGS, 2009) of the region (Bouguer anomalies, fig. 2) shows that the TD and the AD are characterised by positive and negative anomalies, respectively, and are separated by a sharp, arc-shaped step, corresponding to the TZ. On this map, the long-wavelength anomalies reflect the crustal structure whilst the short-wavelength anomalies reflect local and shallower features, mainly corresponding to recent syntectonic basins, filled with soft and light sediments.

The TD is characterized by long-wavelength positive anomaly (maximum +50 to +70 mGal), possibly imaging a mantle dome, centred beneath the Tyrrhenian coast. Scattered, short-wavelength anomalies correspond to the distribution of the recent (Late Miocene-Quaternary) hinterland basins of Tuscany, and of the intervening structural highs.

By contrast, the AD is characterized by long-wavelength negative anomaly (minimum < -150 mGal): the two gravimetric minima grossly correspond to the location of the depocenters of the Po Plain-Adriatic foredeep (beneath Reggio Emilia and Pescara respectively). If the negative anomaly is not due only to the light siliciclastic material of the foredeep (Royden and Karner, 1984), further contributes can derive from the stacking of slices of continental crust beneath the adjacent mountain ridge. In the AD the short-wavelength anomalies are less pronounced than in the TD and seem to be related to the thrust-top basins.

Heat flow data, collected in the NA (Della Vedova *et al.*, 1991; 2001; Mongelli and Zito, 1991) show that the TD is characterised by higher values of surface heat flow (usually > 150 mW m^{-2}) with respect to the AD (30-70 mW m^{-2}). From heat flow data Della Vedova *et al.* (2001) derived the temperature distribution with depth, highlighting the effect of shallow processes, active in the upper few km of the crust, such as fresh water circulation (within the carbonates of the Apennines ridge), sedimentation (in the foredeep areas), rapid erosion (mainly in the Alps), recent volcanic activity (in the geothermal areas of the TD). Correcting these disturbing effects (which still derive from the tectonic setting and evolution), Della Vedova *et al.* (2001) estimated the average values of the “undisturbed conductive heat flow” for the main tectonic provinces of Italy, rapidly decreasing moving from the TD (> 100 mW m^{-2}) to the Apennines Ridge (65-75 mW m^{-2}) till the Po Plain-Adriatic foredeep (40-50 mW m^{-2}).

Figure 2. Gravity map (Bouguer anomalies) of the Northern Apennines.



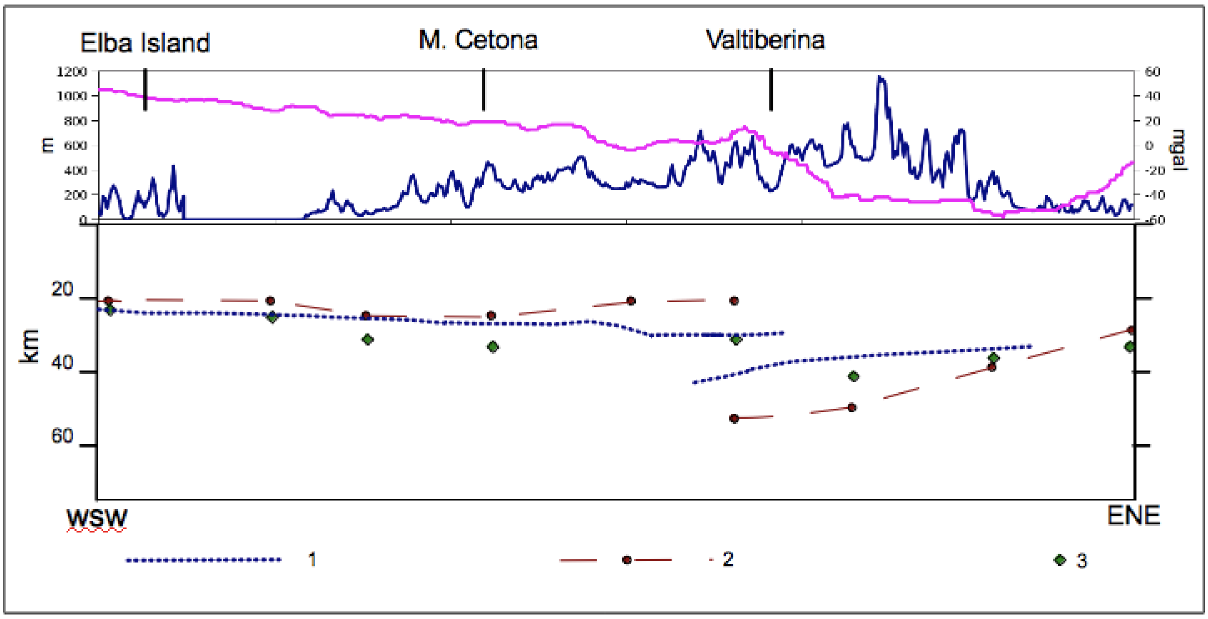
The map is obtained from rasterization of gravity anomaly contours from ISPRA, ENI, OGS (2009). The solid black contour line is the zero value. The underlying shaded relief is derived from the SRTM (Shuttle Radar Topography Mission) 90 m resolution DEM available at http://srtm.csi.cgiar.org/index.asp (Jarvis *et al.*, 2008). The straight solid line is the trace of the section of fig. 3.

Active and passive seismic data

Since the 70's of the last century seismic refraction experiments investigated the crustal structure of the NA and of the adjacent regions. The results were synthesised on Moho isobath maps (e.g. Scarascia *et al.*, 1994; Scrocca *et al.*, 2003), highlighting the presence of a relatively thin Tyrrhenian crust (20-25 km thick), contrasting a thicker Adriatic crust (30-35 km), separated by a sharp Moho step beneath the TZ (the same area where the larger variation of the gravimetric anomalies is observed, see fig. 2). The re-elaboration of these data and the acquisition of new seismic refraction profiles showed a doubling of the crust beneath the TZ (Val di Chiana-Val Tiberina region), laterally extending for about 30 km (Ponziani *et al.*, 1995; De Franco *et al.*, 1998). Seismic refraction data also highlighted a relatively slow Moho (7.7 km/s) in the TD and a faster Moho (about 8.0 km/s) in the AD; gravity modelling confirms the presence of a softer and warmer upper mantle beneath the Tuscany and a denser and older one as proper of the Adria lithosphere (Marson *et al.*, 1998).

The structure of the deep crust and of the upper mantle has been also imaged analyzing the teleseismic waveforms recorded by a regional array crossing the northern Italian peninsula (Piana Agostinetti *et al.*, 2002; Mele and Sandvol, 2003). The crust-mantle boundary revealed by the receiver function analyses is generally consistent with the results of the seismic refraction experiments (Fig. 3), showing a shallow Thyrrenian Moho, slightly deepening towards the east from 22 to 25 km, and a deeper west-dipping Adriatic Moho. However, significantly different estimates were proposed for the depth of the Adriatic Moho below the main ridge of the Apennines, e.g. > 50 km in Mele and Sandvol (2003) and about 35 km in Piana Agostinetti *et al.* (2002).

Figure 3. Moho depth profiles across the Northern Apennines.



Moho depth after: 1- Ponziani *et al.*, 1995; 2- Mele and Sandvol, 2003; 3- Piana Agostinetti *et al.*, 2002. In the upper diagram: the blue line is topography; the red line is a gravimetric profile (data after ISPRA, ENI, OGS, 2009). The trace of the section is on fig. 2.

The map of the lithosphere thickness, derived by passive seismology, shows that the TD is characterized by a strongly reduced lithosphere (about 30 km), whilst the AD lithosphere reaches a thickness of 70 to 90 km (Calcagnile and Panza, 1981; Suhadolc and Panza, 1989). In the TD a thin lithosphere is also supported by a study of thermal data that 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 is reflected in the high surface heat flow and in the positive Bouguer anomalies found in this domain (Marson *et al.*, 1998) . The reduced thickness of the lithospheric mantle (less than 15 km thick) indicates a considerable amount of partial melting (Peccerillo and Panza, 1999). More controversial is the lithosphere thickness in the AD: beneath the Apennine chain, a thickness of about 60 km is derived from the dispersion of the surface waves (Suhadolc and Panza, 1989), whereas a larger thickness (100-150 km) is derived from the P-Wave analysis (Babuska and Plomerova, 1995).

During the 90's, a further contribution to the knowledge of the crustal structure of the NA was offered by the Italian deep crust seismic project (CROP), which acquired crustal scale seismic reflection data across the NA, both onland and offshore. The profile CROP03 (Pialli *et al.*, 1998) extends across the whole central Italy, from

Punta Ala (Tyrrhenian coast) to Gabicce (Adriatic coast), intersecting all the main tectonic units of the NA. CROP profiles M12A and M16 represent the offshore continuation of the CROP03 profile across the Tyrrhenian and the Adriatic Sea, respectively (Finetti *et al.*, 2001; Scrocca *et al.*, 2003). A further profile (CROP18) was acquired onland across the geothermal areas of southern Tuscany (Brogi *et al.*, 2005).

The CROP 03 profile has shown that the TD and the AD exhibit a distinct reflectivity pattern at any crustal level. In the upper crust of Tuscany (TD), extensional tectonics, active since the late Miocene, has largely obliterated the previous compressional structures. In contrast, compressional structures are still clearly preserved in the Umbria-Marche Apennines (AD) where extension has been active for a much shorter time period, about 3Ma, and has only affected the westernmost part of the region. Local earthquake tomography confirms that the compressional structures are still the most visible features of the upper crust, as they have not been obliterated by the active extension, yet (e.g. Chiarabba and Amato, 2003; Di Stefano *et al.*, 2009).

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

Beyond these valuable researches and results, the deep crustal structure and the Moho geometry of the NA is still partially unresolved, particularly in the zone where the transition between the TD and the AD occurs. The Moho doubling is poorly imaged by the CROP03 profile, where the TZ is characterised by 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. These problems are reflected by the contrasting interpretations, which have been proposed for the CROP03 profile by different Authors (Barchi *et al.*, 1998a; Decandia *et al.*, 1998; Finetti *et al.*, 2001; Lavecchia *et al.*, 2003; Pauselli *et al.*, 2006). It is also worth to note that significant differences affect the velocity models of the NA crust, derived by active seismic (e.g. De Franco *et al.*, 1998) and passive seismic (e.g. Di Stefano *et al.*, 2009) experiments.

Approximately beneath the TZ (where the doubling of the crust was detected by the refraction data) tomographic images obtained by different Authors with different methods (e.g. Amato *et al.*, 1993; Spakman *et al.*, 1993; Piromallo and Morelli, 2003) show a pronounced high-velocity anomaly, recognised till the depth (670 km) of the mantle transition zone. This anomaly is commonly interpreted as a submerged slab possibly linked to the crust of the AD. The shape and position of the velocity anomaly are slightly different in the different studies. The most relevant differences pertain to the length of the slab and its lateral and along-dip continuity. At least the shallower part of the slab (down to 100-200 km of depth) should consist of thinned continental crust (i.e. the passive margin of Adria), which was the original substratum of the Apennines units, now stacked in the NA fold-and-thrust belt.

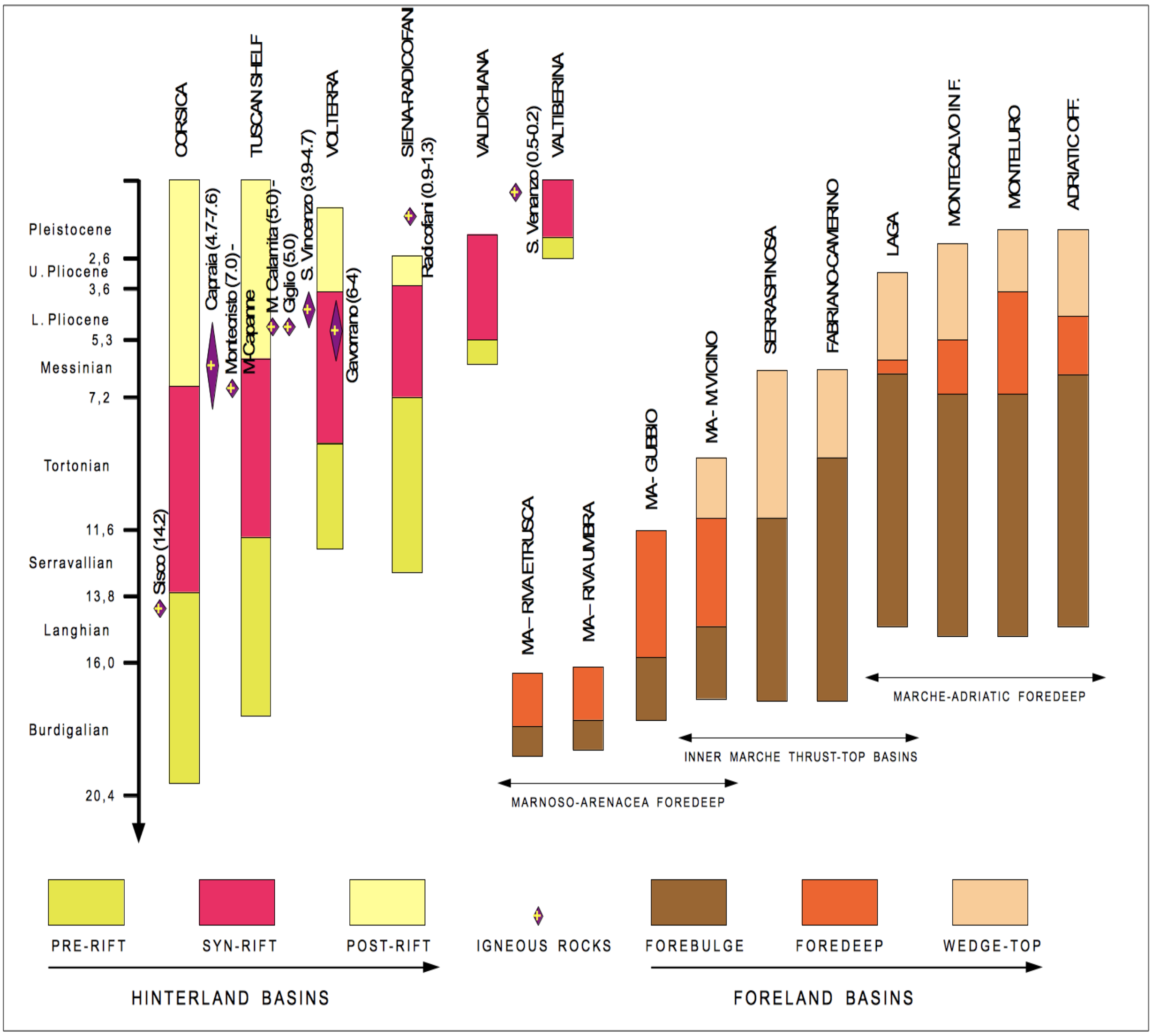
The anomalous high S-wave velocity at depths exceeding 50 km confirms the presence of lithospheric roots below the TZ (Calcagnile and Panza, 1981; Du *et al.*, 1998; Pontevivo and Panza, 2002). These roots are also marked by rare, but regularly recorded, seismic activity (Amato *et al.*, 1997) that reaches a maximum depth of 90 km below the Val Tiberina Basin (see chapter 5).

Chapter 3. Tectonic evolution

The Neogene-Quaternary tectonic evolution of the NA is characterized by the contemporaneous activity and eastward migration of coupled compression (in the foreland) and extension (in the hinterland): this is a key-feature for understanding the crustal structure (chapter 2), as well as the distribution and kinematics of seismicity (chapter 5) of this region. Compression and extension are co-axial, i.e. the direction of maximum extension (i.e. WSW-ENE) is nearly parallel to the maximum shortening induced by the previous compression. Extensional tectonics is also marked by later magmatic activity, also migrating eastward.

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 (fig. 4).

Figure 4. Ages of the hinterland and foreland syntectonic basins of the Northern Apennines.



The basins are considered along a WSW-ENE transect, grossly corresponding to the CROP03 profile. The horizontal distance is not in scale. The ages of the basins are derived from the literature, particularly from previous compilations by Pascucci *et al.*, 1999; Barchi *et al.*, 2006. The ages of the igneous rocks are after Poli 2004, ages after compilation of Serri *et al.*, 2001.

The distinction between the extensional (hinterland) basins and the compressional (foreland) basins is also supported by the analysis of the magnetic fabric (AMS pattern, Mattei *et al.*, 1997). In the literature the compressional nature of the foreland basins is widely recognized, whilst the extensional character of the hinterland basins is still controversial (e.g. Bonini and Sani, 2002 vs. Pascucci *et al.*, 2006, showing contrasting interpretation of the Radicofani basin).

The compressional wave

As Merla (1951) early recognized, the eastward moving NA compressional belt has generated progressively younger foreland basins (foredeep and piggy back basins, Ricci Lucchi, 1986; Ori *et al.* 1986) that were successively incorporated in the fold and thrust belt: the Adriatic Sea represents the younger and easternmost foredeep, active until at least the early Pleistocene. A regional review of the flexural foreland basins of the Apennines has been recently offered by Casero (2004).

The areal distribution of the Miocene-Quaternary foreland basins of the NA is shown in Fig. 1: however, it is necessary to keep in mind that the present-day distribution of these deposits, which presently crop out mostly in the syncline areas, does not necessarily reflect the original shape and size of the original basins, since the width and reciprocal distance between the basins has been substantially reduced by later orogenic contraction.

The diagram in Fig. 4 summarizes the ages of the successions 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 syn-compressional basins typically evolve through three subsequent stages (fig. 4) reflecting the depositional zones of a foreland basin system (e.g. Decelles and Giles, 1996). The resultant sedimentary succession consists of, bottom to top: i) forebulge/backbulge deposits, typically consisting of hemipelagic marls (e.g. Schlier Fm.); ii) foredeep deposits, mainly turbidite sandstones, with variable amounts of intercalated marls (e.g. Marnoso-Arenacea Fm.); iii) wedge top deposits, showing a larger variability of sedimentary environments and facies. During their evolution these basins were affected by frequent variations in the direction of the paleocurrents (e.g. longitudinal from NW vs. transversal from SW), as well as in the provenance of the sediment supply (e.g. Alps vs. Inner Apennines). Synsedimentary tectonics is reflected by the recurrent occurrence of large-scale mass wasting complexes, emplaced in the inner margin of the Oligocene-Miocene foredeep basins of the NA (Lucente and Pini, 2008): these complexes, including both extrabasinal (i.e. olistostrome) and intrabasinal displaced sediments, also show progressively younger ages moving from west to east. The Pliocene-Quaternary denudation complexes, recognised by Ori *et al.* (1986) in the seismic profiles through the Adriatic foredeep may be modern analogues of the Miocene mass wasting complexes.

Along the CROP03 transect, the Miocene-Quaternary foreland basins of the NA have been divided into three major groups, corresponding to, from West to East: i) the Marnoso-Arenacea foredeep; ii) the Inner Marche thrust-top basins; iii) the Marche-Adriatic foredeep.

1. The Marnoso-Arenacea Fm. (Late Burdigalian – Late Serravallian) was deposited in a 'complex foredeep' (sensu Ori *et al.*, 1986), presently cropping out in the region comprised between the eastern front of the Tuscan Units and the main ridge of the Apennines (Fig. 1). The onset of the foredeep was driven by the internal embrication of the innermost Tuscan units, that later overthrust the inner part of the Marnoso-Arenacea basin (Barchi *et al.*, 1998b). Synsedimentary tectonics during the Late Serravallian is suggested by the presence, within the Marnoso-Arenacea succession, of olistostromes and proximal fan deposits, as well as of coeval, slumping phenomena (Ricci Lucchi and Pialli, 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 (Menichetti and Pialli, 1986). The sedimentation is closed by the Lower-Middle Tortonian M. Vicino sandstones, a thrust-top basin, developed over the easternmost part of the foredeep, immediately to the west of the main Apennines Ridge (fig. 1).

2. The Inner-Marche thrust-top basins (Cantalamessa *et al.*, 1986) developed in the Tortonian-Messinian time interval, when no foredeep basin of regional extent existed, but there were only minor basins, restricted between the growing Umbria-Marche folds. Most of these basins (Urbania, Serraspinosa, M. Turrino, Fabriano, Camerino) are presently located above the main ridge of the Umbria-Marche Apennines, or immediately to the west (M. Vicino) and to the east (Laga) of the mountain belt (fig. 1).

3. The Marche-Adriatic syn-orogenic succession was deposited in a foredeep in the Late Messinian-Quaternary time interval. The onset of this basin was related to the major compressional phase of the Umbria-Marche Apennines (Late Messinian-Early Pliocene). Soon after its onset, the forebulge of the Adriatic foreland reached the outermost Dinarides and stopped migrating, so that the topography of the Marche-Adriatic foredeep was affected by a set of syn-sedimentary folds and thrusts (complex foredeep, sensu Ori *et al.*, 1986). The structural setting and the timing of deformation of the Marche-Adriatic foredeep, as well as that of its northern equivalent, the Po Plain, have been depicted in detail in many papers, combining surface geology, biostratigraphy and good quality seismic reflection profiles, calibrated by many boreholes (e.g. Pieri and Groppi, 1980; Castellarin *et al.*, 1985; Argnani *et al.*, 1991; Coward *et al.*, 1999; Scarselli *et al.*, 2006).

Summarising, the age of the foreland basins (fig. 4) effectively illustrates the general eastward migration of the contractional deformation from Late Burdigalian in the Tiber Valley zone to Adriatic coast and a regular superposition of depositional environments, where small mobile basins restricted between the growing ridges are superimposed on regionally extended foredeep basins (Ori *et al.*, 1986). However, the migration is not completely regular: the Marnoso-Arenacea and the Marche-Adriatic basins are regionally extended foredeep, related to the main embrication events of the Tuscan Units (Late Burdigalian-Serravallian) and of the Umbria-Marche Units (Late Messinian-Pliocene) respectively. In the intervening time period (Late Tortonian-Early Messinian) no proper foredeep was present, but only minor basins, restricted between the growing Umbria-Marche folds.

The rate of migration of the compressional wave can be estimated by progressively restoring balanced cross-section through the Umbria-Marche fold and thrust belt. A study made by Basili and Barba (2007) along the CROP03 profile concluded that deformation advanced steadily towards NE, at a time-average rate of about 6 mm/yr. More detailed studies, performed where the timing of deformation is well constrained by syntectonic deposits, suggest a not entirely steady migration of deformation, where a major factor of perturbation is the occurrence of major fall or rise of the base level (e.g. Scarselli *et al.*, 2007).

It is well known that surface transport processes can influence the evolution of a thrust belt, modifying its regular migration from the hinterland to the foreland (e.g. Simpson, 2006). For example, during the Messinian salinity crisis (about 5.6 Ma) the Mediterranean sea level experienced a drop of about 1500 m, followed by a rapid early Pliocene marine ingression. These perturbations affected the rate of migration of the compressional front (and the magnitude of shortening) in the Marche-Adriatic region (Scarselli *et al.*, 2007) and in the Po Plain area as well (Castellarin *et al.*, 1985). Quaternary glaciations may also have had significant effects on the late evolution of the thrust belt and of the related basins.

The extensional wave

Simultaneously with the development of the eastward moving NA compressional belt, at its back the TD 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, described by Elter *et al.* (1975) earlier, was successively confirmed by other research (e.g. Lavecchia *et al.*, 1984; 1987; Carmignani *et al.*, 1994 among many others).

The diagram in Fig. 4 summarizes the ages of the successions infilling the extensional hinterland basins along the CROP03 transect and its offshore continuation across the Northern Tyrrhenian Sea. In strict analogy with the compressional history, the age of the successions systematically decreases from the Corsica basin through the Northern Tyrrhenian Sea and Tuscan Mainland, to reach the Umbria-Marche Apennines ridge in the Early Pleistocene, where extension is presently active (Bartole *et al.*, 1991; Jolivet *et al.*, 1998b; Pascucci *et al.*, 1999, 2007; Collettini *et al.*, 2006).

The diagram of fig. 4 also shows that magmatic activity also migrates eastward, following the formation of the extensional basin. 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 has already occurred. See Peccerillo (2005) for a comprehensive review of the Pliocene-Quaternary volcanic processes of the Apennines.

Pascucci *et al.* (1999) describe the evolution of a typical extensional hinterland basin, consisting of three major subsequent stages: the first pre-rift stage is characterised by narrow bowl-shaped basins or flat-like deposits, whose location is possibly affected by the pre-existing topography and tectonics; during the second syn-rift stage the activity of the major extensional faults promotes the subsidence of triangular shaped, asymmetrical half-grabens; finally, during the third post-rift stage wide bowl-shaped or blanket-type deposits are draped above the previously formed depressions. The three subsequent stages have followed each other through time and space, moving from the westernmost offshore (i.e. Tyrrhenian) areas towards easternmost inshore (fig. 4).

The hinterland basins and the extensional faults driving their evolution have been effectively imaged 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; 2007). 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 M. Amiata (Calamai *et al.*, 1970; Brogi, 2004).

These studies produced significant advances in the knowledge of the extensional process.

The extensional deformation is strongly asymmetric and dominated by a set of east-dipping, low-angle normal faults (Barchi *et al.*, 1998a; Decandia *et al.*, 1998), whose location is strictly connected with the position of the shallow marine and/or continental syn-tectonic basins.

The low-angle normal faults are detached within the upper crust. Beneath Tuscany, the basal detachment corresponds to a prominent reflector (k-horizon, Cameli *et al.*, 1993), located at an average depth of 10 km, and it slightly deepens towards the east, reaching a maximum depth of about 5s (up to 14 km). This is the same depth of both the Brittle/Ductile transition inferred from the thermal field (Pauselli and Federico, 2002) and the cut-off of seismicity (Chiarabba *et al.*, 2005; De Luca *et al.*, 2009). The underlying lower crust shows an anomalously reduced thickness (about 9 km), possibly produced by ductile flow. In this view, the Tyrrhenian Domain has been extended by brittle/semi-brittle fault zones in the upper crust, coupled with prevailing ductile flow in the lower crust.

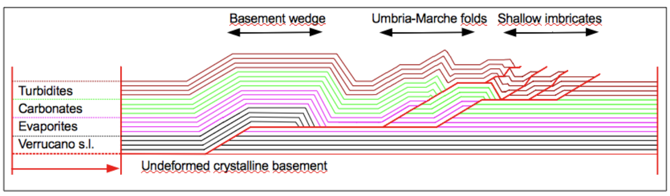
The easternmost fault of this extensional system is an active, east-dipping low-angle normal fault (Alto-Tiberina Fault, ATF), exposed in the Umbria region, at the western border of the AD (Boncio *et al.*, 1998; 2000; Collettini and Barchi, 2002). The ATF cuts the thickened crust down to a depth of about 14 km, greater than that reached by the older, no longer active faults of the TD, reflecting incipient extension. Strong evidence for the present-day activity of the ATF is furnished by micro-seismicity surveys, where most of the recorded seismicity fits the trace of the ATF (Boncio *et al.*, 1998; Chiaraluce *et al.*, 2007). The focal mechanisms show extensional kinematics with NNW-SSE trending planes, parallel to the ATF fault, 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. The present day activity of ATF (and of some of its high-angle splays) is also suggested by geomorphological evidence (Cattuto *et al.*, 1995), and confirmed by GPS data, showing about 2.5 mm/yr of NE extension across the High Tiber valley (D'Agostino *et al.*, 2009; Hreinsdóttir and Bennett, 2009), as well as by high resolution seismic reflection survey across the SW margin of the Sansepolcro basin (Delle Donne *et al.*, 2007). ATF is suitable to represent the presently active expression of the extensional wave of the NA.

Chapter 4. Mechanical stratigraphy and style of deformation

The structural style of both compressional and extensional structures of the NA is strongly influenced by the mechanical anisotropy within the upper crust. In fact the sedimentary cover is composed by alternating competent and not-competent units, whilst the shallower part of the basement often correspond to a low-Vp horizon, interpreted as a mechanically weak layer (e.g. Brogi and Liotta, 2008; Mirabella *et al.*, 2008).

In the AD, the Miocene-Quaternary compressional structures affect the Umbria-Marche Mesozoic-Tertiary succession, widely described in the literature (e.g., Centamore *et al.*, 1986; Cresta *et al.*, 1989, and references therein). From a mechanical point of view, this succession can be divided into four main litho-structural units. top to bottom: turbidites, carbonates, evaporites and basement s.l. (fig. 5). The upper part of the "basement" includes Permian-Middle Triassic clastic rocks (e.g. Verrucano Formation) and/or Late Palaeozoic meta-sedimentary rocks. The mechanical boundaries among these units correspond to the main *décollement*s of the compressional structures, as well as to the best seismic markers, driving the interpretation of the seismic profiles.

Figure 5. Thrust trajectory and mechanical stratigraphy of the sedimentary cover in the Adriatic Domain.



Schematic view of a thrust fault propagating through the Umbria-Marche succession, generating a set of progressively shallower structures, bottom to top: basement wedges (*décolled* within the upper part of the basement s.l.), Umbria Marche folds (*décolled* within the Late Triassic evaporites) and shallow imbricates (*décolled* at the top of the carbonates).

The stratigraphic and structural setting of the TD upper crust is more complex, consisting of a thick stack of embricated Liguride and Tuscan Units, successively dismembered by the Miocene-Quaternary extensional structures (e.g. Carmignani *et al.*, 1994 and this volume; Molli *et al.*, this volume). Beyond the structural complexity, this also results into a mechanical alternation of competent (e.g. Macigno, Tuscan carbonates and dolomites) and less competent (e.g. Ligurids, evaporites, phyllites) units.

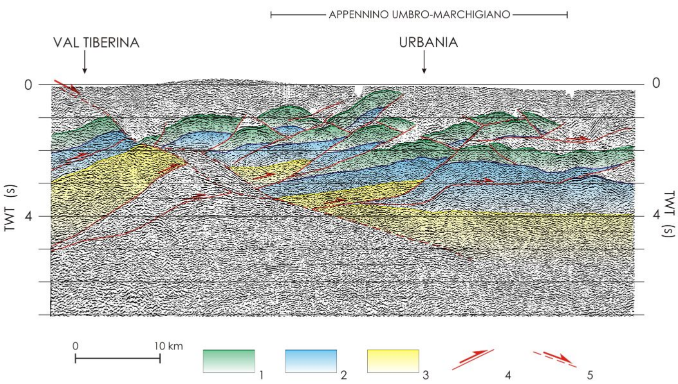
Structural style of the compressional belt

Contrasting interpretations of the structural style of the Umbria-Marche fold and thrust belt have been proposed by different Authors, mainly by integrating surface geology data with seismic reflection profiles (e.g. Ghisetti *et al.*, 1993 for a discussion). The proposed interpretations can be divided into two main groups. A first group of authors interpreted Umbria-Marche Apennines as a typical thin-skinned thrust belt, where thrusts developed in a regular, piggy-back sequence, over a main basal *décollement*, corresponding to the Triassic evaporites (Bally *et al.*, 1986; Calamita and Deiana, 1986; Hill and Hayward, 1988). A second group recognized the involvement of the basement in the major thrust sheets (Lavecchia *et al.*, 1987; Sage *et al.*, 1991; Barchi, 1991; Barchi *et al.*, 1998b), some of these invoking the reactivation of previous normal faults (tectonic inversion, e.g. Coward *et al.*, 1999; Butler *et al.*, 2006). Contrasting interpretations of the basement geometry and depth, and of its involvement in the major thrust sheets, have also been inferred from aeromagnetic data (Arisi Rota and Fichera, 1985; Cassano *et al.*, 1998; 2001; Speranza and Chiappini, 2002).

The style of deformation beneath the main mountain ridge of the AD is effectively imaged by a portion of the CROP03 profile, where two prominent reflections can be recognised, corresponding to the Marne a Fucoidi Formation (close to the top of the carbonates) and to the top of the basement s.l. (fig. 6). Other reflectors (e.g. top evaporites, top carbonates multilayer, Messinian evaporites) can also be traced locally.

At depth, at least the shallower part of the basement is involved in the major thrust sheets, progressively deepening towards the east. Above these structures, the Marne a Fucoidi reflections describe a set of wide anticlines (involving the carbonates and the evaporites) whose geometry and wavelength resemble the typical Umbria-Marche folds, cropping out in the main mountain ridge. The shallower part of the section, above the Marne a Fucoidi seismic marker, is characterized by numerous reflections, imaging a complex pattern of short wavelength structures disharmonically overlying the carbonates.

Figure 6. Interpretation of a portion of the CROP03 profile across the main mountain ridge of the Apennines (modified after Barchi e al., 1998).



The seismic profile illustrates the style of deformation of the subsurface structures of the Umbria-Marche Apennines, in particular the relationships between the basement wedges and the Umbria-Marche folds.

This structural setting supports a style of deformation, characterized by a system of multiple *décollement*s, where different sets of structures are generated at different structural levels, linked to each other and developed in a hierarchical mode. The size (i.e. wavelength and amplitude) of the structures depends on the depth of the *décollement*, which they sole to, so that the deeper the *décollement*, the larger the structures (fig. 5).

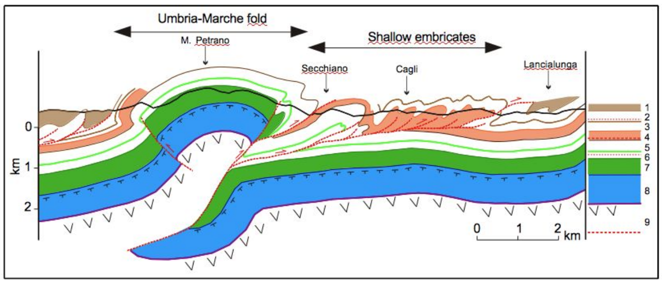
The most important classes of structures seismically and geologically detectable are:

1. shallow imbricates, detached on the top of the carbonates and involving the turbidites; along the section of figs. 6 and 7, these structures show a wavelength of few hundreds meters;
2. Umbria-Marche folds, detached at an intra-evaporites level and involving the carbonates, with structures that have a 5 to 10 km wavelength;
3. basement thrusts, corresponding to the major structural units of the region (having a 25–35 km wavelength) and that transfer the shortening from the upper part of the basement up to the sedimentary cover.

The structures generated at different structural levels may be connected along asymmetric major thrust systems, propagating across the whole upper crust, from lower towards upper crustal levels and branching into a progressively larger number of splays (fig. 5), so that shallower structures are more numerous and possess a shorter wavelength than the corresponding deeper structures.

The relationships between the Umbria-Marche folds and the shallow embricates are also illustrated in the geological section of fig. 7, crossing the culmination of a NE verging anticline (M. Petrano anticline) in the inner ridge of the Umbria-Marche Apennines. The M. Petrano anticline involves the entire carbonates succession and is décolled in the underlying evaporites. The major thrust propagates upward into a flat trajectory in the marly horizons at the top of the carbonates, splaying out into a set of small, narrow, shallow seated embricates.

Figure 7. Geological section throuh the inner ridge of the Umbria-Marche Apennines (modified after Massoli *et al.*, 2006).



The geological section illustrates the style of deformation of the shallow structures of the Umbria-Marche Apennines, in particular the relationships between the Umbria-Marche folds and the shallow imbricates.

1- Marnoso-Arenacea; 2- Schlier; 3- Bisciaro; 4- Scaglia Cinerea and ScagliaVariegata; 5- Scaglia Rossa and Scaglia Bianca; 6- Marne a Fucoidi; 7- Jurassic multilayer; 8- Calcare Massiccio; 9- Anidriti di Burano. The red dotted lines mark the main *décollement*s.

This style of deformation is also supported by many other studies, mostly published in the last 20 years and based on shallower commercial profiles, crossing both the mountain ridge and the adjacent deformed foredeep, from the Po Plain to the Marche-Adriatic region (e.g. De Donatis *et al.*, 1998; Turrini *et al.*, 2001; Massoli *et al.*, 2006).

It is important to note that the wavelength of the embricated structures, detached above the Mesozoic-Paleogene carbonates, depends on the thickness of the syn-orogenic clastics. Where the Tertiary clastic succession is very thick (e.g. Po Plain, Pescara basin), the “shallow” structures show a much longer wavelength (4.5 to 8.2 km, Massoli *et al.*, 2006) than in the inner part of the Umbria-Marche fold and thrust belt (e.g. fig. 7).

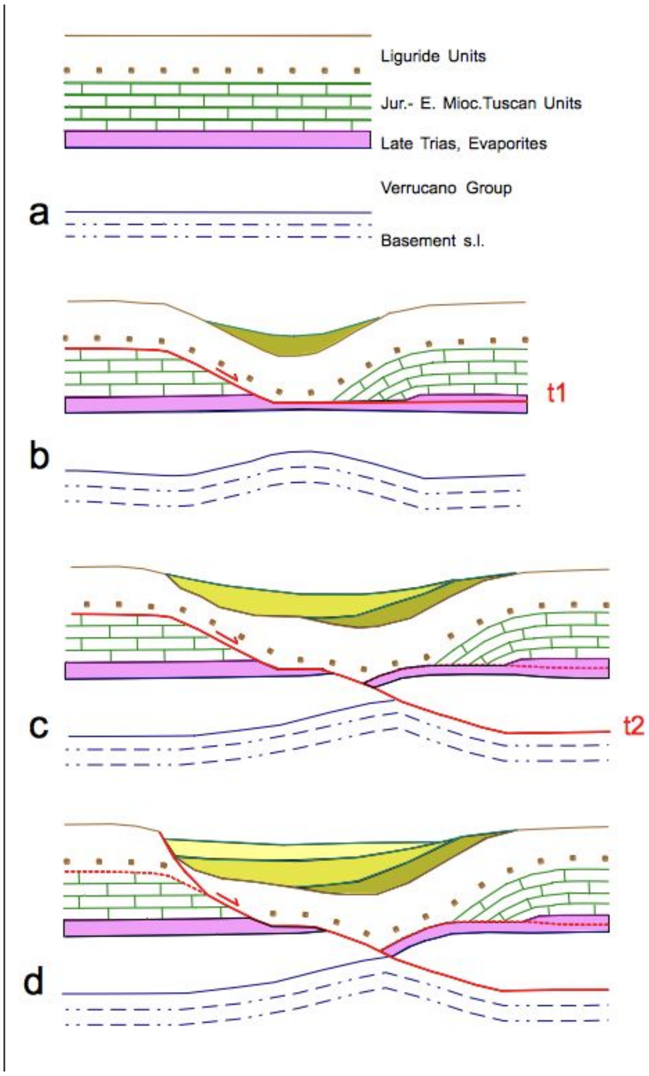
Structural style of the extensional belt

As the mechanical stratigraphy of the Umbria-Marche succession affects the flat-ramp-flat geometry of the compressional structures of the AD, the complex stratigraphy of the Tuscan upper crust also affects the geometry of the major east-dipping normal faults of the TD, which are characterised by a marked staircase trajectory, influencing the deformation of the hangingwall blocks and the evolution of the hinterland basins as well (fig. 8).

The presence of low-angle normal faults in the Tuscan mainland was initially hypothesised to explain anomalous stratigraphic relationships in the Mesozoic successions (i.e. “serie ridotta”; e.g. Lavecchia *et al.*, 1984). Successively, in the geothermal areas of Southern Tuscany the availability of large data-sets of surface and subsurface data (seismic reflection profiles, calibrated by a dense network of boreholes) allowed many Authors to reconstruct in detail the down-dip geometry of the major extensional detachments, producing asymmetrical boudinage of the involved crustal volumes and the “lateral segmentation” of the competent units (e.g. Brogi and Liotta, 2008), with main flats within (or at the base of) the Liguride units and the Triassic evaporites (Calcare Cavernoso), and the ramps cutting trough the carbonates of the Tuscan Units. At a larger scale, low-angle detachment with staircase trajectory produce the boudinage of the entire upper crust (Brogi, 2004; Brogi *et al.*, 2005).

A pronounced staircase trajectory also characterises the still active ATF, whose geometry has been reconstructed over an area of about 1800 km^2 using a network of seismic reflection profiles (Barchi *et al.*, 1999; Chiaraluce *et al.*, 2007). In this case, the main *décollement*s driving the curvi-planar geometry are located at the top of the carbonates and within the phyllitic basement. Since the ATF may be considered a seismogenic fault, its staircase geometry potentially produces down-dip segmentation of the fault, thus affecting the dimensions of the active fault segments and the maximum expected magnitude, to be considered for seismic hazard evaluation.

Figure 8. Normal faults with staircase trajectory and related hinterland basin in the Tyrrhenian Domain (modified after Brogi and Liotta, 2008).



a- schematic stratigraphy of the TD; b- the movement along the t1 detachment produces lateral segmentation of the Tuscan carbonates and the onset of a bowl-shaped basin within a syncline ramp; c- the movement along the t2 detachment produces further extension and the enlargement of the basin; d- high-angle faults, splaying out from the main detachment propagate up to the surface: the basin evolved into a semi-graben, driven by a major east-dipping detachment.

The staircase trajectory of the major detachments can have significantly affected the onset and evolution of the hinterland basins, generated at their hangingwall (fig. 8). The initial bowl-shaped geometry of the hinterland basins has been used to support the hypothesis of a syn-compressional development of the hinterland basins (e.g. Bonini and Sani, 2002). However, this geometry can be also explained in an extensional context: in fact, bowl-shaped basins may be generated within ramp synclines at the hangingwall of curvi-planar normal faults, as hypotheised by Brogi and Liotta (2008) for the Radicondoli basin; alternatively, the upward propagation of normal faults can produce the dragging of the syntectonic strata along the basin flank (extensional fault propagation folds, e.g. Withjack *et al.*, 1990), as illustrated by Sachpazi *et al.* (2003) for the faults bordering the Corinth Gulf basin.

Chapter 5. Active tectonics and seismotectonics

The Northern Apennines are a tectonically and seismically active region, releasing almost continuous micro-seismicity and relatively frequent moderate (5<M<7) earthquakes, mostly concentrated along the mountain ridge (Chiarabba *et al.*, 2005; Gruppo di lavoro CPTI, 2004). This is a relevant social problem, since this is a densely populated region, rich of cultural heritage. From a scientific point of view, seismological data provide a reliable and detailed image of the present-day stress field of the region and, along with geodetic data, depict the contemporary (short term) kinematics of the region, to be compared with the geological (long-term) tectono-sedimentary evolution, described in chapter 3.

The distribution of the seismicity is also related to the lithosphere structure and rheology, depending on both composition and local geotherm (i.e. T-depth profiles).

Rheological profiles of the lithosphere and seismicity distribution

Pauselli and Federico (2002) produced a regional scale study of the rheology of the NA crust, based on the study of the thermal field along the CROP03 transect, aimed to locate the brittle/ductile transition (as defined by Brace and Kohlstedt, 1980). According to this study, in the westernmost part of the region (TD) the brittle/ductile transition occurs at a depth of 10 to 12 km; moving farther east, it deepens from 12 km (beneath the TZ) to about 25 km (in the AD). In the easternmost part of the profile the rheological behaviour of the lithosphere is more complex, with rocks being brittle at a depth less than 25 km but also at depth greater than 34 km. A more complete study, performed by Pauselli *et al.* (2010), in which a high-pressure brittle fracture mechanisms (Zang *et al.*, 2007) is included in the thermo-rheological model across the NA, shows important lateral variations of both total lithosphere strength and of the strength distribution with depth. The TD is characterised by a “crème brulée” structure, where only one load-bearing layer (the top half of the upper crust, 6 to 8 km thick) contributes to most of the lithosphere strength. In the AD the lithosphere is much stronger and a “jelly sandwich” structure is present, with both lower crust and upper mantle contributing to the lithosphere strength.

The different rheological profiles characterising the TD and the AD are reflected by the different distribution of the earthquake’s hypocenters (Chiarabba *et al.*, 2005; De Luca *et al.*, 2009).

The TD is characterised by scarce, shallow-seated (mostly < 10 km deep), low-magnitude seismicity, concentrated in the shallower part of the upper crust: the reduced thickness of the seismogenic layer can be easily referred to the weak and thinned crust of the Tuscan region and to the geothermal and post-magmatic processes active in the area (e.g. Liotta and Ranalli, 1999). In the AD the distribution of the seismicity is more complex. Both instrumental and historical seismicity (Gruppo di lavoro CPTI, 2004) including the larger (5 < M < 7) extensional earthquakes of the region, are concentrated along a NW-SE trending belt, narrower in the northern part (about 20 km wide), wider in the southern part. Across this belt, grossly corresponding to the TZ and to the westernmost part of the AD, crustal seismicity deepens from west to east, down to a depth of 25 km, locally highlighting two separate clusters, located in the upper and lower crust respectively. The eastern part of the AD is characterised by a diffuse microseismicity, mostly located in the shallower part of the crust (< 15 km deep) and by few moderate events (M < 5.5). The AD is also characterised by less frequent, but regularly recorded intermediate seismicity, occurring along a west-dipping structure down to about 70 km. This sub-crustal seismicity is generally interpreted as connected to the descending Adriatic continental slab.

Summarising, the available data highlight a good correspondence between the distribution of the shallow seismicity and the rheology of the upper crust, which is thicker and stronger in the AD with respect to the TD. For a full comprehension of these relationships, better earthquake locations are needed, along with more detailed rheological profiles, keeping in count e.g. the mechanical variations within the upper crust, where different rock-types (soft sediments, sedimentary rocks, crystalline rocks) are superposed. For example, below the axial ridge of the Apennines the distribution of the seismicity suggests that the brittle crust is partitioned into three different layers (Mirabella *et al.*, 2008): a low-velocity horizon, corresponding to the shallower part of the basement, decouples the overlying sedimentary cover (where moderate seismicity is produced) from the underlying crystalline basement (where only microseismicity occurs).

High-pressured fluids can also play an important role in earthquake triggering (Miller *et al.*, 2004), in a region where a huge CO_2 flux is observed (Chiodini *et al.*, 2004).

Finally, we have to consider that the seismicity is not only a function of the rheological properties of the lithosphere, but also reflects the active tectonic structures and processes.

Present-day stress and strain field

The focal mechanisms of the earthquakes reflect the contemporaneous activity of extension in the hinterland and compression in the foreland area. This pattern was early described since the 80's (e.g. Lavecchia *et al.*, 1984; 1994; Frepoli and Amato, 1997) and is now supported by a large collection of focal mechanisms, whose quality and quantity have continuously increased with time (e.g. Pondrelli *et al.*, 2006, De Luca *et al.*, 2009).

The axial ridge of the Apennine belt is characterized by moderate to large normal-fault earthquakes (5<M<7), mostly occurring on NW-trending adjacent segments (Amato *et al.*, 1998, Chiaraluce *et al.*, 2003). The contiguity of these segments delineates an elongated extensional belt that obliquely crosses the arc-shaped, pre-existing compressional structures and is mostly confined in the upper 6-8 km of depth. This extensional belt longitudinally crosses the whole NA and it is clearly marked by the alignment of several intra-mountain basins, which, from NW to SE, are labeled the Lunigiana-Garfagnana, Mugello, Casentino, Sansepolcro, Gubbio, Colfiorito and Norcia basins (fig. 1). This is part of a much longer alignment, affecting the Central (L’Aquila, Fucino) and the Southern (Irpinia, Val D’Agri) Apennines as well.

In Northern Umbria, the seismogenic normal faults splay out from a major low-angle NE-dipping detachment (Alto Tiberina Fault, see also chapter 3), whose presence and activity is supported by both active and passive seismic data (Chiaraluce *et al.*, 2007 and references therein). This fault represents the most recent, still active extensional master fault, driving the extension of the NA hinterland.

A belt of shallow (D < 15 km), contractional or traspressional earthquakes occurs in the eastern part of the AD, in the Po Plain-Adriatic foreland, from the southern edge of the Po Plain (e.g. Reggio Emilia, Forlì), down to the Adriatic coast, from Pesaro to Ancona to Porto S. Giorgio. This seismicity could mark the position of the currently active compressional front of the NA, even if the activity of the compressional front in the AD is presently debated (e.g. Di Bucci and Mazzoli, 2002; Scrocca, 2006).

In an intermediate position, between the presently active extensional and compressional belt, some earthquakes are registered at a depth of 15-25 km, showing transpressional or compressional kinematics. In this region, also some significant historical earthquakes occurred (e.g. Cagli, 1791). These events could represent the expression of the deeper part of the Adriatic crust, even if the oblique kinematics also supports the hypothesis that these events are nucleated along transfer faults, segmenting the major thrusts.

The Italy stress map (Montone *et al.*, 2004), merging the seismological data with other geological information (i.e. borehole breakout data and structural data of active faults), confirms and reinforces this setting, showing two coupled, NW-SE trending, nearly parallel belts of extension and compression.

In the last few years GPS data provided new effective constraints to the kinematic framework of the region (D’Agostino *et al.*, 2009; Hreinsdóttir and Bennett, 2009). At present these data indicate active extension across the main ridge of the Apennines, with an average rate of about 2.5-3 mm/yr, in good agreement with the seismicity data and the other stress filed indicators. The kinematics of the compressional belt are still undefined, and different hypothesis have been made about the present-day tectonic setting of the Adriatic region (e.g. D'Agostino *et al.*, 2008) and about the activity of the Po Plain-Adriatic thrust (Scrocca, 2006).

It is certain that geodetic data will improve continuously in the next years, giving a further contribute to the definition of the present-day tectonic setting.

Chapter 6. Final remarks

The present and the past

As we have seen in the previous chapter, the present-day stress field of the NA is characterised by compression in the foreland and extension in the hinterland (Montone *et al.*, 2004). In particular, at shallow crustal levels we have compressional earthquakes in the Po Plain-Adriatic region and extensional earthquakes along the axial ridge of the Apennines. The presently active compressional and extensional belts are clearly superposed to the alignment of the compressional and extensional basins that were active in the early Pleistocene (e.g. Lavecchia *et al.*, 1994; Scrocca *et al.*, 2007).

This observation supports a uniformitarian view, where the long-lived tectonic process that governed the evolution of the NA is still active and is now expressed by the present-day stress field and by the on-going deformation as well. If the present is a key for the past, then also the past is a key for the present, and we can use long-term geological observation to better understand the on-going tectonic processes. In this view, the seismicity depicts a snapshot of the present-day tectonic setting, the last stage of a much longer tectonic history, active since about 17 Ma, whose main feature is the progressive eastward migration of a coupled pair of compression and extension, from the Tyrrhenian to the Adriatic region.

The outstanding similarity of the general framework, supporting the uniformitarian concept, overwhelms the local differences and irregularities in the rate of migration of the compressional and extensional waves (see chapter 3). These irregularities could be partly related to the rise and fall of the base level, induced by climatic or eustatic changes, occurred during the long-lived tectonic process, deeply affecting the onset and evolution of both hinterland and foreland basins, modifying the pattern of surface transport processes (i.e. erosion and sedimentation, see chapter 3).

Barchi *et al.* (2006) described the tight space and time relationships between compression and extension, highlighted by the geological data about the age of the syntectonic basins (long-term image) as well as by the instrumental seismicity pattern (short-term image). These relationships are expressed through three simple “rules”, controlling the switch from the compressional to the extensional regime.

The first rule describes how the tectonic regime varies in the horizontal space, and says that at any time compression and extension are contemporaneously active, in the foreland and in the hinterland, respectively.

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

The third rule describes how the tectonic regime varies with depth, and says that extension can occur at shallow levels (in the uppermost crust) at the same time and position of compression at greater depth (lower crust and upper mantle).

These rules are strictly geometrical and they are valid independently from the adopted geodynamic model, where the space and time relationships between compression and extension are framed.

Geodynamic setting

The NA are characterised by the coexistence and interaction of two different tectonic environments, which need to be understood and framed into an appropriate geodynamic setting. The AD is a typical thrust and fold belt, where a sedimentary succession, originally deposited on a continental passive margin, is involved at the periphery of a collisional orogen. The TD, with a thinned crust and lithosphere, displays all the typical geological and geophysical features of an extensional belt (Lister and Davis, 1989), such as positive Bouguer anomalies, high heat flow and strong magmatic activity.

Many different hypotheses have been proposed in the last 40 years, that can be referred to three main groups of models, i.e.: extension-dominated models; compression dominated models; and complex models, where extension and compression derive from a unique process and are of the same order of magnitude.

The first group of models assigns the main role in the dynamics of the Apennines to extension. Extension is driven by the eastward shifting of an asthenospheric rising plume, and produces active rifting in the Tyrrhenian Sea (Lavecchia, 1988; Decandia *et al.* 1998; Lavecchia *et al.*, 2003). In this view, compression in the foreland would be a secondary effect, induced at the outer border of the extending region by horizontal forces (rift push), generated by the strong difference in the lithospheric thickness between the thinned TD and the unthinned AD (Lavecchia *et al.* 2003).

A second group of models calls for a quasi-continuous, eastward thrusting of the Apenninic crust and tends to minimize the role of extension (Boccaletti *et al.*, 1997; 1999). Following this hypothesis, Finetti *et al.* (2001) proposed a crustal section across the entire NA system (from eastern Corsica to the Adriatic offshore) based on the interpretation of the deep crustal profiles M12A, CROP03 and M16. In this view, the hinterland basins were formed as compressional (thrust-top) basins, driven by out-of sequence thrusts, and the extension is a recent process, superposed on a previous compressional history (e.g. Boccaletti *et al.*, 1997; Bonini and Sani, 2002).

The third group of models invokes a combination of extensional and compressional mechanism and suggests that the eastward shift of the Northern Apennines orogenic system was driven by the roll back of a subducting slab (Scandone, 1980; Royden *et al.*, 1987; Doglioni 1991; Cavinato and DeCelles, 1999) and/or by delamination (Channel, 1986; Channel and Mareschal, 1989) of a crustal slice dipping westward into the upper mantle.

This hypothesis, based on the roll-back of the subducting Adriatic slab, can explain better most of the geological and geophysical features of the NA. In fact, the Apennines wedge is formed over a low-strength *décollement*, possesses a low topographic expression and is characterised by extension at the rear of the wedge; al these features are typical of orogens associated with the roll-back of a subduction zone. A model of west-dipping subduction and retreat (roll-back) of the Adriatic slab also explains the presence of the cold body imaged by mantle tomography and the presence of the subcrustal seismicity.

However, many uncertainties still remain: the deep seismicity is scarce and its location with respect to the supposed Adriatic slab needs to be improved (e.g. Lavecchia *et al.*, 2003); moreover, the depth of the earthquake foci does not exceed 90 km. The location and geometry of the Adriatic slab is also matter of debate: the lateral and vertical continuity of the slab imaged by tomography is complex and controversial, possibly due to the complexity and segmentation (Royden *et al.*, 1987; Scrocca, 2006) of the Apennines subduction. In particular it is not clear why the cold body is not present below the adjacent central Apennines, where a similar tectonic evolution is observed, including the migration of both compressional

(Cipollari and Cosentino, 1997) and extensional basins (Cavinato and De Celles, 1999), and a similarity exists in both the crustal structure and the distribution of the seismicity (Ghisetti and Vezzani, 2002).

In conclusion, a satisfactory geophysical and geological model of the (Northern) Apennines is still to be produced. We need further reliable information about the geometry of the transition zone (see e.g. fig. 3). We also need to reconsider the geological constraints, in order to quantify and compare the rates of deformation related to both the extensional and compressional wave, and their variations through time.

Finally, it is important to consider that the problem of the geodynamic setting involves a lithospheric scale and cannot be addressed considering only the NA region, but needs to consider the entire Apennines belt and possibly the entire Mediterranean area.

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