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# Three-dimensional numerical modeling of contemporary mantle flow and tectonic stress beneath the Central Mediterranean

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The structure, density and effective viscosity of the crust and uppermost mantle beneath the Central Mediterranean influence lithospheric deformation, mantle flow, and tectonic stress state. To estimate the contribution of buoyancy forces to regional dynamics, three-dimensional finite-element models are developed to determine contemporary uppermost mantle flow and tectonic stresses. We use density models for the crust and uppermost mantle derived from *S*-wave seismic velocities and constrained by gravity data. The viscosity model is constrained by the observed strain rate and regional heat flow data. The modeled movement of the uppermost crust is consistent with the northeast-oriented motion of the lithosphere and is in an agreement with the geodetic measurements. The modeled flow patterns of the lower crust and uppermost mantle are consistent with the regional observations. The models predict (i) northwest-oriented movements beneath the southeast part of the Adriatic Sea region, (ii) the northeastern subduction beneath the western part of the Adriatic Sea, (iii) the upwelling beneath the Tyrrhenian Sea and its eastern coast, (iv) the western movement of the Ionian Sea sub-plate, and (v) the subduction beneath the western Calabria region. Our models predict also a distinct compressional regime along the northeast part of the Italian peninsula and to the east of Sicily, and a tensional regime beneath the Tyrrhenian Sea, Umbria–Marche region, and Ionian Sea. The predicted tectonic stress regimes in the northern and central Apennines are in agreement with stress regimes derived from earthquake fault-plane solutions. Changes in the predicted crustal stress pattern and magnitude are likely to be caused by buoyancy-driven mantle circulation beneath the region rather than by gravitational potential energy differences in the crust itself. Based on the model results, we conclude that the buoyancy forces play an important role in the contemporary tectonics of the region.



## 1. Introduction

The Central Mediterranean geology has been mainly shaped by the interplay between the Eurasian and African plates. The extremely variable structure of the lithosphere–asthenosphere system in the region is the result of its complex geodynamic history. The Cenozoic to Quaternary regional evolution has been marked by the coexisting compression and tension developed between converging continental plates (e.g., Doglioni et al., 1999; Faccenna et al., 2004). However, the rate of convergence between the plates has been less significant compared to the east–west extension (e.g., Mantovani et al., 2002). The latter has been migrating from west to east and has been positioned behind a compression front migrating in the same direction. As a consequence, a number of extensional basins have formed behind the Apennines–Maghrebian compression front. Orogenic magmatism followed the eastward migrating extensional regime, becoming younger from Sardinia (Oligo-Miocene) to the Tyrrhenian Sea floor and to the Southern Tyrrhenian Sea (Peccerillo, 2003, 2005 and references therein).

The eastward migration of the Apennines compression front is accompanied by a fragmentation of the Apennines lithosphere, with progressive ending of the active subduction zone from the Northern Apennines to the south. The fragmentation of the Apennines lithosphere created sectors that had an independent evolution (Locardi, 1993; Sartori, 2003). This may explain the variable lithosphere–asthenosphere structure in the region.

Several authors suggested continuous west-dipping subduction of the Adriatic and Ionian plates beneath the southern margin of Europe (e.g., Carminati et al., 1998; Doglioni et al., 1999; Faccenna et al., 2004 and references therein). This follows an older subduction process, having an opposite dipping direction (e.g., Peccerillo and Martinotti,

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2006). The subduction front has been migrating more or less continuously eastward, up to its present position in the southern Tyrrhenian Sea (e.g., Cavazza and Wezel, 2003). The opening of the Tyrrhenian Sea basin is considered as related to backarc spreading behind the eastward migrating compression front (e.g., Sartori, 2003 and references therein). The integrated use of geophysical, petrological and geochemical data allowed to infer three possible types of processes that are likely governing the lithosphere dynamics in the Tyrrhenian and Italian Peninsula (Panza et al., 2007): delamination in the Northern and Central Apennines, slab detachment to continuous subduction in the Southern Apennines, slab roll-back and tearing with sideways asthenospheric flow through slab windows in the Calabrian Arc with a likely slab detachment.

The present-day configuration of the crust and upper mantle in the northern and central parts of Italy, that is the result of complex evolutionary stages of thickening and thinning of the lithosphere in interaction with the underlying asthenosphere, may contribute body forces to the regional geodynamics. As shown by Chimera et al. (2003), the imaged uppermost mantle supports the lithospheric delamination beneath the peninsula and provides a unifying background for petrological and geochemical studies of recent magmatism and volcanism in Tuscany. Furthermore, Ismail-Zadeh et al. (2004a) and Aoudia et al. (2007) have shown that buoyancy forces resulting from the density distribution govern the present dynamics of the lithosphere within North-Central Italy and can explain regional coexisting contraction and extension and the shallow depth and unusual distribution of intermediate depth earthquakes.

Internal buoyancy sources (due to spatial variations of the crustal and uppermost mantle density) influence a stress field by variations of gravitational potential energy, which in their turn produce tectonic (deviatoric) stresses. The principal purposes of this paper are to show (i) how the density and viscosity heterogeneities in the lithosphere–asthenosphere structure contribute to the regional dynamics and mantle flow beneath the central Mediterranean and (ii) how these heterogeneities govern the tectonic stress localizations. To estimate the contribution of the buoyancy sources to the modern tectonics of the region, we develop three-dimensional (3-D) finite-element models of the instantaneous crust–mantle dynamics and resulting tectonic stress, where the model input parameters (geometry, density, and viscosity) are based on the results of surface wave seismic tomography and the subsequent density and gravity modeling.

The results of the recent surface wave seismic tomography studies conducted in the Central Mediterranean are summarized in Section 2, and the density and viscosity models are discussed in Sections 3 and 4, respectively. The numerical model of the regional mantle dynamics and relevant computational approach are described in Section 5. The results of the numerical modeling are presented in Section 6 and discussed in Section 7. We discuss uncertainties and limitations of the study in Section 8 and present concluding remarks in Section 9.

## 2. Regional surface wave tomography

Understanding the lithosphere–asthenosphere structure beneath the central Mediterranean is fundamental for gaining a better insight into geophysical processes, including tectonic stress generation and seismicity, but also for unraveling the past geodynamic history of the region. Information on the properties of the upper mantle and lower crust can be gained by geophysical investigations, namely from seismic tomography and gravimetric studies. The Italian peninsula and surrounding regions were intensively studied by means of seismic surface wave tomography (e.g., Panza et al., 2007 and references therein).

The surface wave dispersion curves (Pontevivo and Panza, 2002; Pontevivo, 2003; Pontevivo and Panza, 2006) determined by either the two-stations method (Panza, 1976) or frequency-time analysis to extract the fundamental mode of Rayleigh waves (Levshin et al., 1972, 1992) were used to produce tomography maps employing the 2-D tomography algorithm by Yanovskaya and Ditmar (1990). To develop models of *S*-wave velocity (vs. depth), a grid (of the size $1° \times 1°$), comparable with the resolving power of the dispersion data, was imposed on the tomography maps. For each cell the tomographic values on the four knots were averaged to define its local (cellular) dispersion curve.

The analyzed cellular dispersion curves range from 7 s to 150 s, are appropriate to explore the *S*-wave velocity structures down to about 350 km (Panza et al., 2007). The lateral resolving power of the dispersion data is of about 200 km, however, the availability of a priori independent geological and geophysical information on the uppermost part of the crust improves the lateral resolving power (Pontevivo and Panza, 2002; Panza et al., 2004). Therefore, for each cell ($1° \times 1°$) the uppermost (6–10 km) crustal structure was fixed according to independent data (Panza et al., 2003). The structure below the inverted layers is the same for all cells and has been fixed according to published data (Du et al., 1998). Using the non-linear inversion method (Valyus et al., 1969; Knopoff, 1972; Panza, 1981), *S*-wave velocity vs. depth was determined for the cells covering the Italian peninsula and surrounding areas (Panza et al., 2007). The full database of the Earth velocity models (up to 350 km depth), used in this study and covering the geographic area delimited in Fig. 1, are reported in Panza et al. (2003), Farina (2006), Panza et al. (2007), and Panza and Raykova (2008). Examples of seismic velocity models sampling the first 100 km and crossing the northern, central, and southern parts of the region are shown in Fig. 2. These sections highlight a complex interaction between the crust and the uppermost mantle.

The southern Alpine region (Fig. 2, cells e0–e1) is marked by a thick crust with the Moho depth ranging between 35 km to 45 km; this is consistent with independent estimates of the crustal thickness in the region (e.g., Nicolich and Dal Piaz, 1990; Marone et al., 2003). In the Po Basin (cells e2–e3) a high-velocity lower crust seems to be absent, and a layer with *S*-wave velocity $V_S$~3.40–3.45 km $s^{-1}$ is detected in the depth-range from 7–8 km to 28–30 km. The uppermost mantle beneath the Po Basin is composed by relatively low-velocity mantle material (below the Moho and to the depth about 60 km) underlain by a high-velocity lithospheric lid. Along the Adria Dinarides front (cells e4–e5) a high-velocity lower crust extends to a depth of about 35–40 km. A high-velocity lid is detected from about 60 to 85 km in cell e4 and below 60 km in cell e5.

Along the Tyrrhenian coastline (cells b0–b1) the crustal thickness varies from about 19 km (cell b0) to 22 km (cells b1). The crustal thickness (cells b2–b3) of the North and Central Apennines is estimated to be 35 to 40 km. A mantle wedge (8 km thick with $V_S$~3.95 km $s^{-1}$) is determined (cell b0) below the Moho overlying the upper mantle. An uppermost mantle layer (cell b1) starts at a depth of 23 km. The relatively low-velocity uppermost mantle layer seems to be absent beneath cell b2. Along the Adriatic coastline (cells b4–b5) the crust is about 35 km thick, and it is characterized by an about 15–20 km thick lower crust. In the Central Apennines and near the coasts of Abruzzo (cells b3–b4) a relatively high-velocity lid is determined below the Moho, and it reaches a depth of about 100 km. The mantle lithosphere (cell b5) reaches a depth of about 165 km. In the central Adriatic Sea (cells b6–b8) the crust is about 25 km thick. A low-velocity layer below the Moho (cells b6–b8) is overlying the mantle lithosphere. This layer (cell b8) is resting on the top of a high-velocity lid.

The crustal thickness varies from ~20 km in the Southern Tyrrhenian Basin (cells C1–C4) to 44 km in the Messina Strait (in cell C5–C6). A low-velocity layer (about 10 km thick) is detected in cells C3 and C4 and gets thicker (30 km) in cells C1 and C2. The asthenosphere (cells C1–C4) is likely to be present at a depth of below 70 km. In general, the modeled velocity of the upper mantle (cells C5 and C6) is high in both, the lithosphere and the asthenosphere. The modeled crustal structure shows a thin crust (cells C7–C9) beneath