Highlighting the crustal structure of southern Tuscany by the reprocessing of the CROP03 NVR profile

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ABSTRACT. The formation of the Tyrrhenian basin and the formation and evolution of the Apennines are believed to be the consequence of the settlement of two mantle domes, one in the southern Tyrrhenian Sea and one in the Tuscan-Latium area. These domes are considered the result of the interaction between the pre-existing lithospheric mantle (and crust) and a thermal input in presence of fluids derived from deep mantle sources. A consequence is the peculiar structure of the crust which interacts with the uppermost mantle, in a region where extensional regime, crustal thinning and magma intrusions have been identified utilizing geophysical data, and in particular the reprocessing of reflection seismic lines crossing an area where important geothermal fields have been exploited. The new models were derived from heat-flow and gravity anomalies, deep seismic soundings, S-waves velocity models constructed by shallow and deep tomographic inversion and subsequent modelling of flow and stress distribution in the lithosphere, deep reflection lines (CROP profiles). The reprocessed section of the deep seismic profile CROP-03, in the sector from the Tyrrhenian shore to the Siena graben, puts in evidence important features, that are compared directly to the land profiles CROP-18A and -18B and to LISA lines acquired offshore, in the Tuscan archipelago. The high resolution of reflection seismic images, and the AVO analysis, highlight the lithology and physical properties of the crustal rocks of southern Tuscany. The intrusion of mantle-derived magmas and lithology differentiation inside the lower crust, the vertical channels corresponding to the ascent paths of magma from crust-mantle transition and fluids dominating the upper crust of southern Tuscany represent the source and motor of the geothermal resources.

1. Introduction

Southern Tuscany is a key area for the exploitation of geothermal energy. The reservoirs were discovered and studied from near-surface evidence and the geological contest and, more recently, utilizing 2D and 3D reflection seismic profiling, research furthered by ENEL, the main company for electric power supply in Italy.

High amplitude reflecting seismic horizon, generally followed by a remarkable and thick high energy interval, was revealed within the metamorphic-crystalline units of the upper crust, at a depth of between 3 to 6 kilometres, when acquiring data over the main geothermal fields of Larderello and Mount Amiata. It was indicated as the K-horizon by Batini *et al.* (1978) and its unexpected reflectivity was mostly related to trapped fluids (occurrence of "bright spots"–like

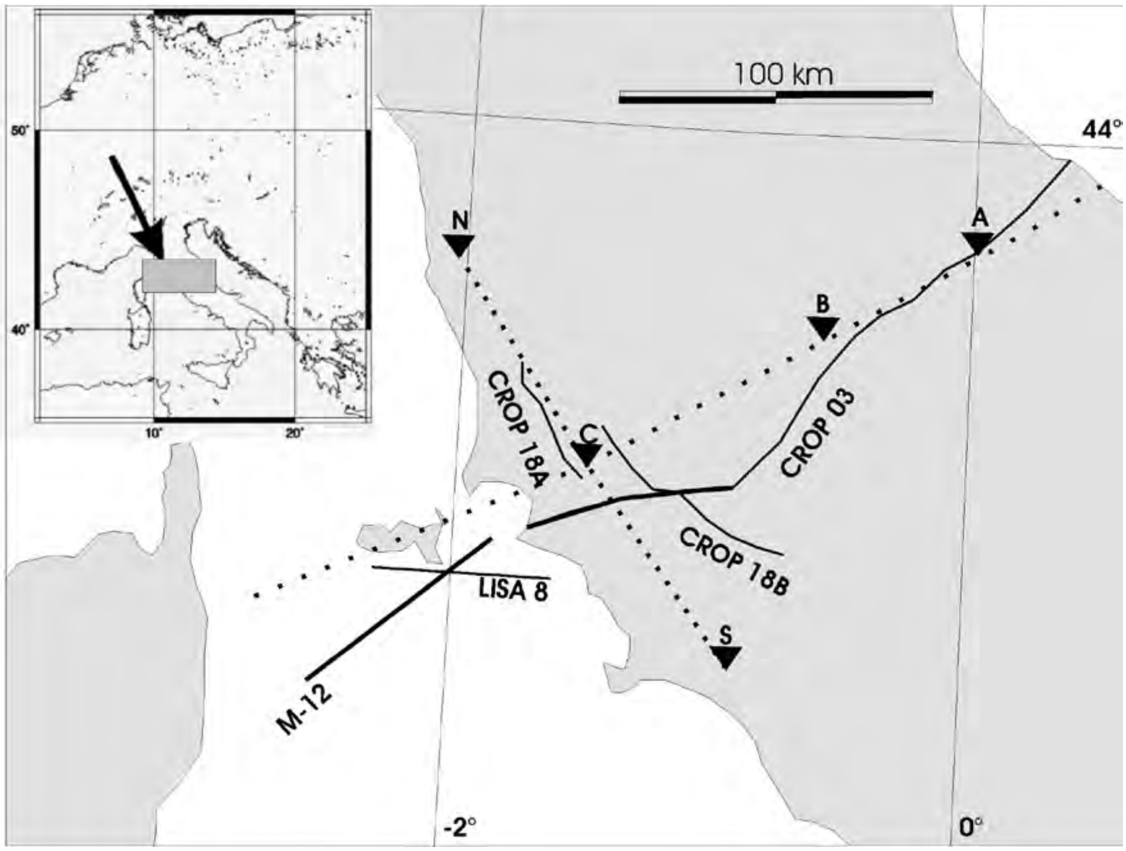

Fig. 1 - Location map of CROP seismic lines acquired in Tuscany, from Elba Island to the Northern Apennines. The analysed sector of the profile CROP-03 is indicated as thick line. Traces of profiles crossing the Tuscan archipelago are reported M-12, from Scrocca *et al.*, (2003); Lisa-08, from Mauffret *et al.*, 1999). The position of DSS profiles [N–S; C–B-A; from Giese *et al.*, (1981)] is presented with dotted lines. Dots indicate also the position of the sketch model presented in Fig. 3.

images). However, any deep well sampled that horizon, because of the presence of severe temperatures and fluid pressure conditions (Batini *et al.*, 1983). Therefore, the petrophysical and lithological properties of the K-horizon and the origin of high amplitude reflections within the metamorphic-crystalline basement cannot be proved directly.

To understand the crustal structures and their relationships to the geothermal fields three seismic lines (CROP-18/A, -18/B and CROP-03, Fig. 1) were acquired on land in the frame of the CROsta Profonda (CROP) project [Scrocca *et al.*, (2003) and references therein]. Other CROP lines have been presented by Finetti (2005). Together with wide-angle reflection and refraction deep seismic soundings (DSS) data (Giese *et al.*, 1981), the CROP seismic lines give information about the crustal structure of the geothermal province.

Several models have been recommended to describe the crustal process in the northern Apennines and southern Tuscany. For example, Doglioni *et al.* [(2005), and references therein] suggest a subduction model, which can be passive. Locardi and Nicolich (2005) put forward that after Middle Miocene the Tyrrhenian–Apennine region should be considered a closed kinematic system. Geodynamic forces were local and large lateral motions, induced by plate dynamics, were absent. They wrote that during the last 10-14 My these forces transformed a segment of the Alpine-Adriatic collision belt into a deep basin and into a new mountain chain. From the analysis of geophysical and geological data Locardi and Nicolich (2005) infer that mantle upwelling above a deep-seated thermal plume caused this tectonic "revolution". A support to this geological occurrence is given by the proposed presence of a 5 to 20km thick layer of dense magma (a soft mantle wedge, or mantle cushion) that underplates the western part of the Apennine range. This layer has contributed to the accretion of a new continental crust, an example of physical and chemical growing. The magmatic composition and the volcano-tectonic evolution indicate a rift tectonic environment and the transformation of the lithosphere mantle and crust by a thermal anomaly and by a fluid supply from deep mantle sources. A mantle plume model has been recently proposed also by Lavecchia *et al.* (2004) and Bell *et al.* (2004, 2005). Subduction (slab rollback or passive subduction) was widely discussed by Carminati *et al.* [(2005), with references therein].

Mapping Moho's isobaths from DSS data, two distinct mantle domes, in the Tuscan – Latium area and in the South Tyrrhenian Sea, were defined (Locardi and Nicolich, 1988). The domes evolved independently and had different rates and direction of migration. After a first eastward drift, the northern one migrated to the northeast gaining a rate of up to 2.3 cm/y, whereas the southern dome moved to the southeast at a much higher rate of 6-8 cm/y (Locardi 1986, 1988). The different migration rates and directions can account for the different tectonic evolution between the northern and southern Apennine range.

Fig. 2 shows a scheme for the mechanism of accumulation of fluids in a partly molten mass at the base of the crust, considered across the southern dome, about 100 km south with respect to our investigated area. The fluidized sector acts both as lubricant and as driving force for the tectonic movements. The model is consistent with the convective motions denoted by: 1) the opening of the Tyrrhenian basin; 2) the migration of the oceanized zone; 3) the volcanism and 4) the tectonic transport (Locardi and Nicolich, 1988, 2005). In Tuscany, the above mentioned lowest migration rates of the northern dome prevented large lateral displacements of the crust-mantle system, but the new geophysical data discussed in this paper can detail a model also for that region.

In Tuscany the volcanologic data show an eastward drift of the eruptive axes, since Tortonian, followed by their counterclockwise rotation, with the emplacement in its final position by Pleistocene. Peccerillo (2002, 2003) proposes an ages interval spanning from 12-14 My, in the western margin of the Tuscan Archipelago, to about 0.2 Ma to the east. The almost contemporaneous magmatic activity along single eruptive axes, hundreds of kilometres long, is another feature of the northern dome. In fact, in the same area (Southern Tuscany, Latium, Campania), the high potash "Mediterranean" volcanism erupted along an inland arc of about a 450 km length (Locardi, 1988) when the Apennine structure was already emplaced and its strong tectonic diversification along the range had already occurred. The eruption began almost contemporaneously 0.9–0.6 My ago and can be considered presently active and along the whole arc.

Peccerillo (2003) and Peccerillo and Lustrino (2005) suggest a completely different model to

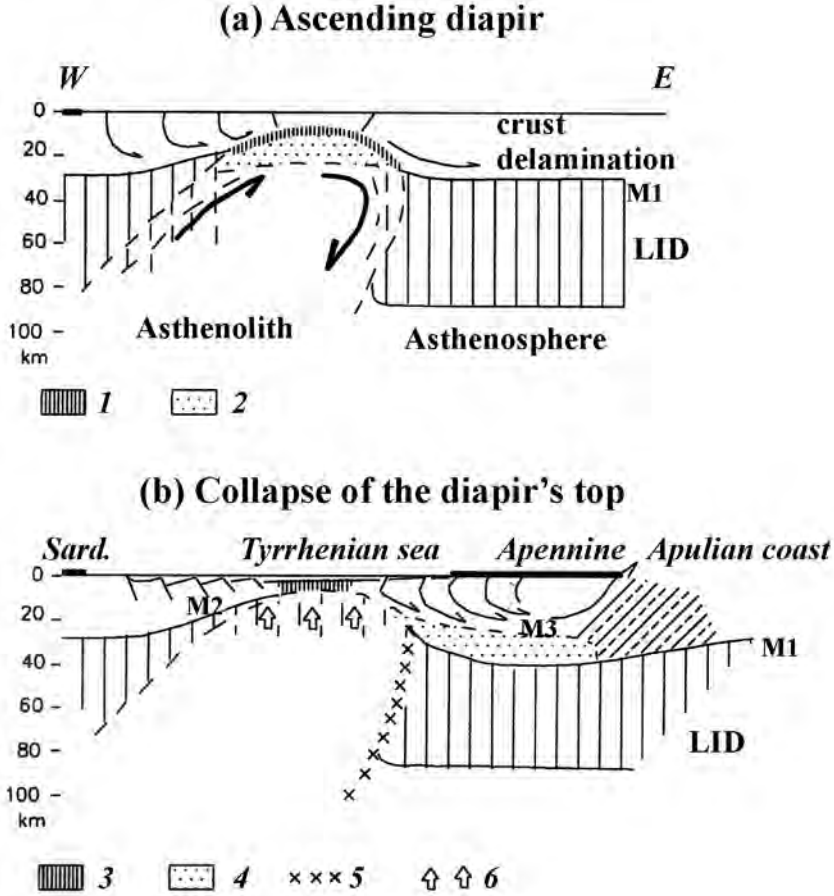

Fig. 2 - Evolutionary scheme proposed for the asthenosphere diapir in the southern Tyrrhenian [a sketch transect from Sardinia Island to the Adriatic coasts of Apulia, from Locardi and Nicolich, (2005)], about 100 km south with respect to our investigated area. M1= Adria Moho, M2=Tyrrhenian exposed mantle, M3= top of mantle derived underplated material. Thin arrows= crustal structure delamination mechanism. Thick arrows: convective movements inside the mantle. 1= crust reduced by delamination and successively melted; 2= zone of maximum fluids accumulation; 3= oceanization; 4= mantle-derived underplated material; 5= seismic arrays corresponding to Benioff's zone of the southern Tyrrhenian reaching more than one hundred km at depth ; 6= isostatic uprising.

explain the origin of Tuscan magmatism, also taking into account the interaction between crust and mantle and the consequent geodynamic contest.

Locardi and Nicolich (2005) infer a link between crustal anatectic and high potash mantle magmatism, which should correspond only to the area underplated by the heavy mantle material. They consider that the crust was molten above the migrating mantle dome where CO_2-dominated fluids exhaling from the mantle where it is accumulated.

Aoudia *et al.* (2004, 2005) inspected the crust and uppermost mantle evaluating the continental deformation and modelling the contemporary flow and stress distribution in the lithosphere. They investigated the contribution of internal buoyancy forces using the S-wave velocity model by Chimera *et al.* (2003), retrieved with the CROP-03 profile data and with a high

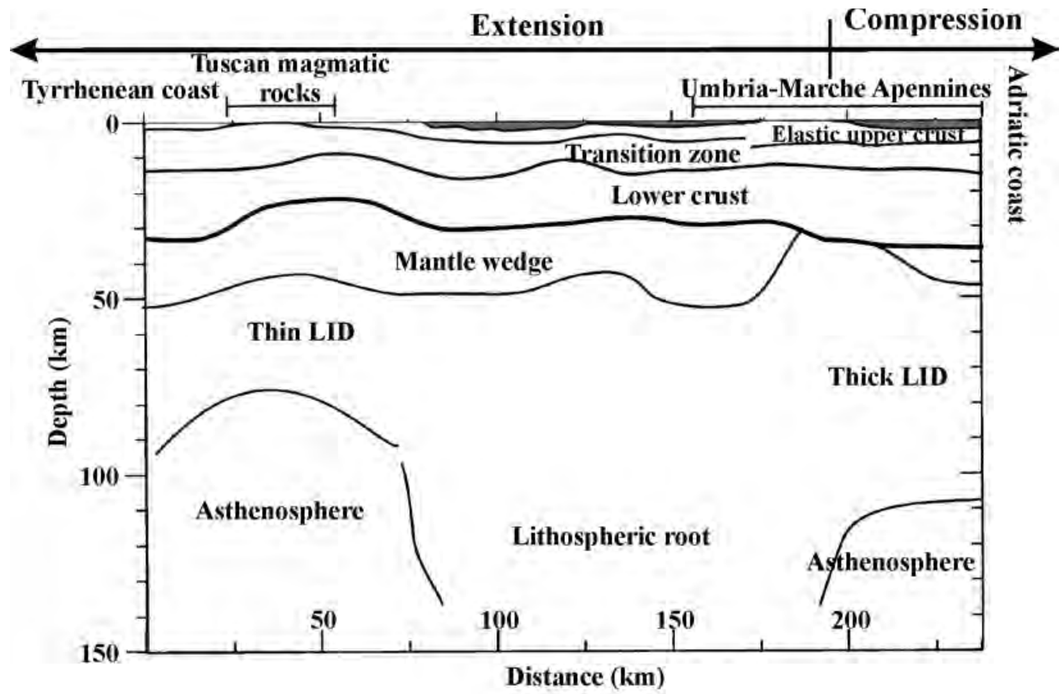

Fig. 3 - Sketch of the lithosphere-asthenosphere system, with a soft mantle wedge beneath Tuscany, approximately following the CROP-03 transect [after Aoudia *et al.*, (2004); position in Fig. 1].

resolution gravimetric survey utilized to estimate density values (Marson *et al.*, 1998) as a priory constraint of shallow and deep tomographic inversion of surface waves. A sharp well-developed low-velocity zone in the uppermost mantle (mantle wedge, cushion), between 30 and 50 km depth and more than 130 km wide, from the Tyrrhenian dying out beneath the Apennines, separates crust and lid (sketch model in Fig. 3).

The puzzling hypocenter location of the sub-crustal seismicity, which does not define a Wadati-Benioff seismic zone (Selvaggi and Amato, 1992; Barchi *et al.*, 2006), is explained by Aoudia *et al.*, (2005), by the particular geometry and buoyancy of the delaminating lithosphere. In the crust, their model predicts compression regime at the eastern part of the profile and tension in Tuscany over the mantle wedge with severe earthquakes in the region where the lithosphere is highly deformed.

Geophysical data acquired across the Tuscan geothermal province support this model. Important information is represented by refraction and wide-angle reflection deep seismic soundings [DSS; Giese *et al.*, (1981)], which defined the top of the lower crust at a 12 to 16 km depth, marked by an interface with a refraction velocity of 6.8 – 7.3 km/s. The Moho discontinuity is at 22-25 km depth (Nicolich, 1989; Locardi and Nicolich, 2005), characterized by an anomalous low-interval velocity of 7.5-7.8 km/s for the culmination of the above mantle wedge beneath the geothermal province. Simplified crustal models along two DSS profiles (N-S directed NW-SE and C-B-A, WSW-ENE) are reported in Fig. 4. The lower part of the crust is

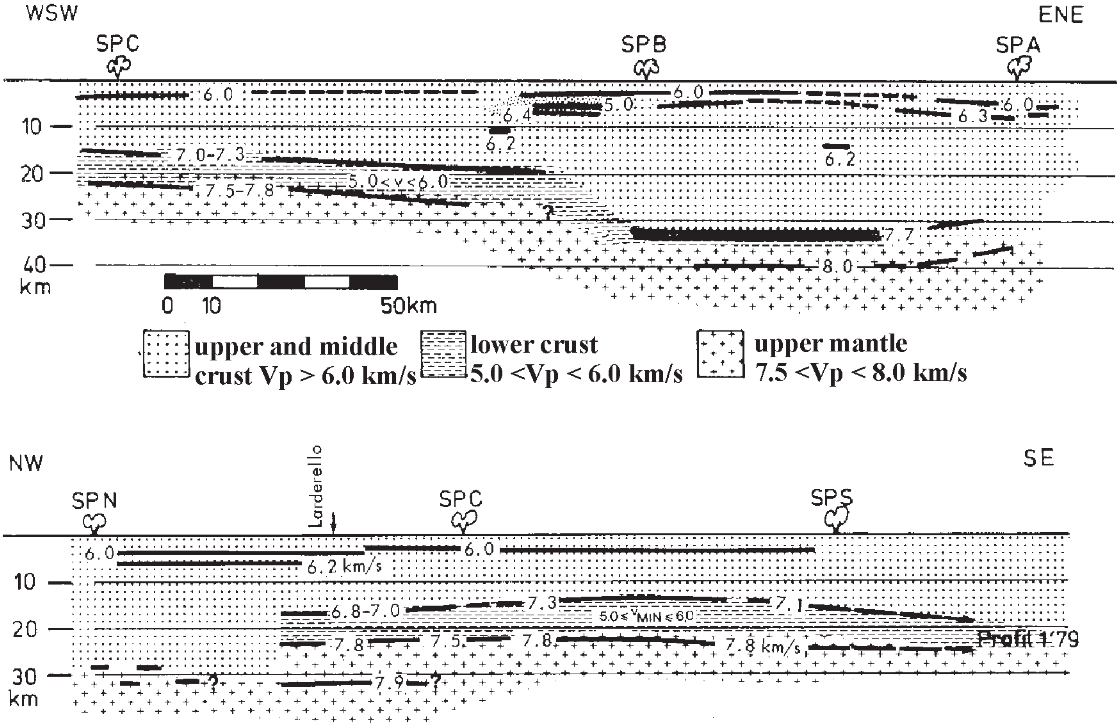

Fig. 4 - Interpretation scheme of DSS data (from Giese *et al.*, 1981) along the C-B-A and N-S profile (position in Fig. 1).

marked and has been defined as anomalous with the presence of high and low velocity layers. The rapid velocity changes can favor the observed reflectivity at DSS's wavelengths. The velocity model was refined by Ponziani *et al.* (1995) and by De Franco *et al.* (1998) mantaining the differentiation of the crustal layers down to the Moho.

Seismic velocities, that are intermediate between the velocities of the mantle and those of the lower crust (7.5-7.8 km/s) are generally interpreted as underplating of heavy mafic melts extracted by a hot mantle plume (Bonatti and Seyer, 1987). The area where those values have been found is delimited by the highest heat-flow values (Della Vedova *et al.*, 1991, 2001), that correspond to the extension of the mantle wedge and are in agreement with an upward flow field, suggesting that the mantle wedge is partially molten.

Gravity data, when the Bouguer anomalies are computed with a reduction density of 2.670 kg/m^3, are evidence of an excess of mass in the lower crust of the geothermal province, which is isostatically undercompensated (Velicogna *et al.*, 1996; Marson *et al.*, 1998). In addition, the geoid height shows a steep gradient moving from western Tuscany towards the Umbria-Marche regions (Barzaghi *et al.*, 2002). On the contrary, local scale negative anomalies, when utilizing a density of 2.400 kg/m^3, indicate a deficit of mass in the upper crust (Nicolich and Marson, 1994; Tinivella *et al.*, 2005).

Pluton emplacements, studied by Acocella and Rossetti (2002), require roof uplift to provide

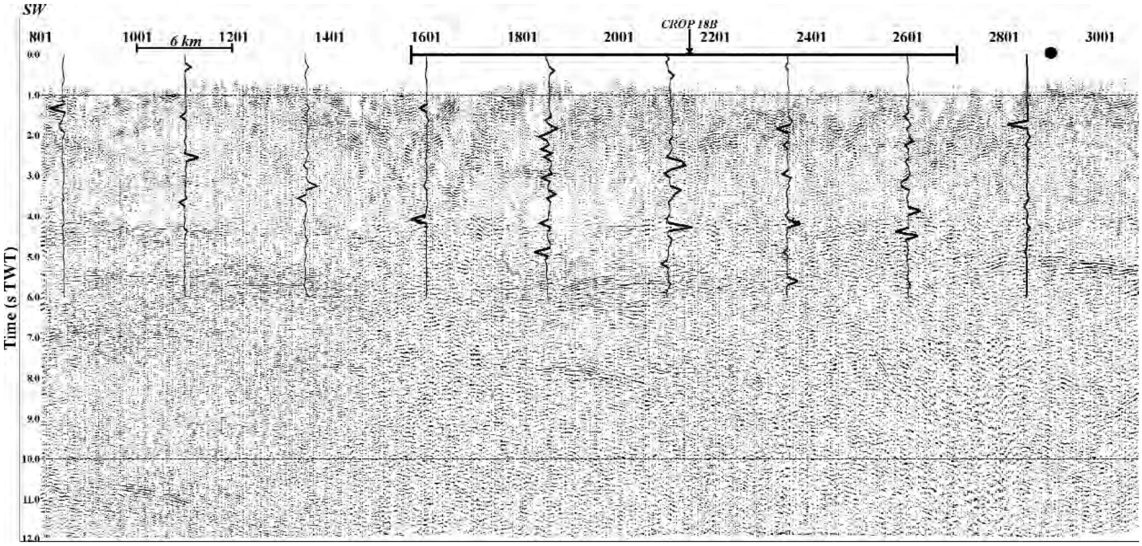

Fig. 5 - Stack section after true-amplitude processing of the western portion of the CROP-03 profile (see Fig. 1). Datum plane: + 200 m above sea level. The ratio between P- and S-waves reflectivity is over-imposed to the section at selected CDP (see the text for details). An arrow indicates the crossing point of the line CROP-18B. The horizontal solid line shows the overlapping with the receivers of the wide-angle acquisition with shots fired in correspondence of the black circle.

the pathway to the ascent of granite magmas in the brittle-extending crust to give the necessary space to the ascent and emplacement of the volcanic staff. A broad uplift of more than 600 m for Pliocene deposits characterize the geothermal fields. A geological and tectonic description of the southern Tuscany can be found in Decandia *et al.* (1988), Liotta *et al.* (1998), Brogi *et al.* (2003, 2005), with references therein.

In this paper, we present the seismic images of the deep crustal structures corresponding to the geothermal fields of Tuscany and the results of non-conventional analyses (such as, true amplitide processing, pre-stack migration and amplitude versus offset). We stress that the resolution and the penetration of the crustal data cannot provide a detailed description of the very complex near-surface geology and of the mantle structures.

2. CROP-03 reprocessed seismic data

The recent deep seismic reflection data, recorded in the studied area [CROP-03 and CROP-18A, B profiles: Pialli *et al.*, (1998); Accaino *et al.*, (2005b)], are of great interest. The results of reprocessing and analysis of the available data of the western sector of the seismic line CROP-03 (indicated with a thick line in the map of Fig. 1) are presented through a multidisciplinary approach here. The main objective of the study was determined by the need to obtain the structure of the crust completed by information on the composition and fluids content. Improving the signal/noise ratio and preserving the reflection amplitudes, the distinct seismic units are differentiated offering a direct image of the variable lithological and petrophysical properties

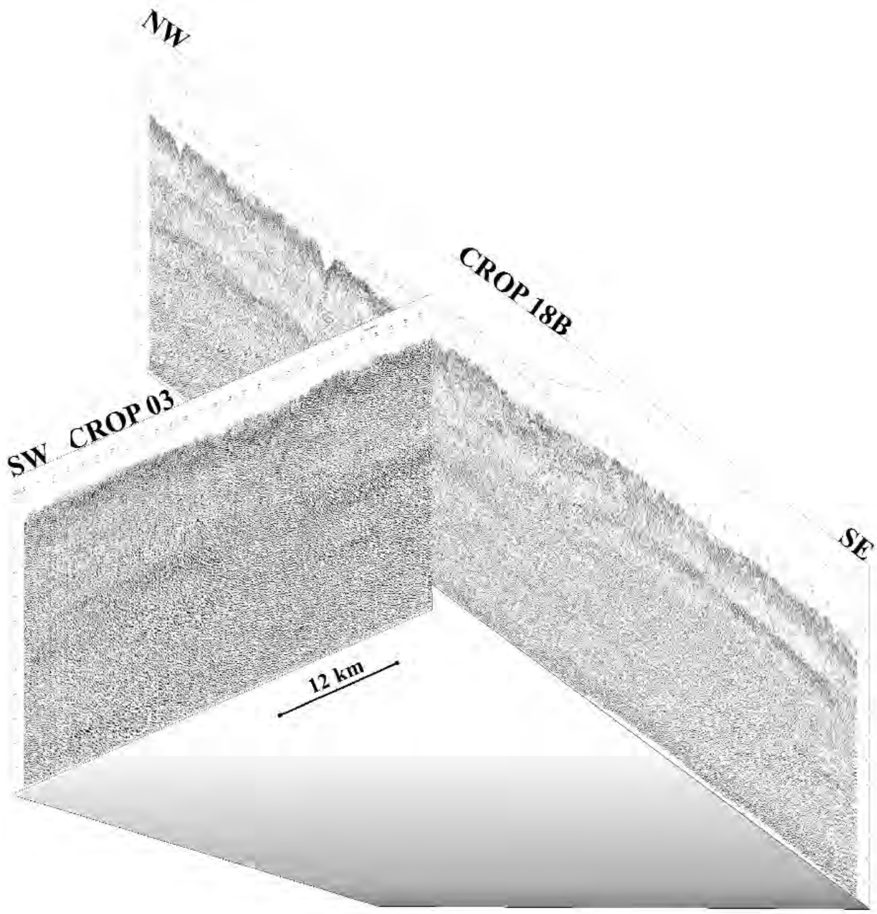

Fig. 6 - Stacked sections of the lines CROP-18B (Accaino *et al.*, 2005b) and CROP-03 plotted in perspective. The continuity between the two sections in correspondence with the main reflection zone starting at about 4 s TWT is evident. The K-horizon reflecting interval is imaged only beneath the Mount Amiata geothermal field.

characterizing the crust of the geothermal province.

The applied processing flow chart is consistent with the elaboration performed on the other CROP lines in the area [CROP-18A and -18B; details in Accaino *et al.*, (2005b)] allowing the comparison of different seismic sections. Shot/receiver geometry applied to the raw data defined the common mid points (CMPs) producing correctly stacked signals. Surface-consistent amplitude balancing completed the recovery of the amplitude losses (spherical divergence and absorption phenomena or variable ground coupling of the geophones and diverse energy of each shot). Source and receiver static corrections were evaluated, after picking the first arrivals, by means of the commercial software (GEOTOMO©) that utilises tomographic approaches. Surface consistent deconvolution has been applied selecting a prediction lag of 24 ms (that corresponds to the second zero crossing) and an operator length of 140 ms. A preliminary velocity analysis, after a trimmed dip filter (Holcombe and Wojslaw, 1992), with the data corrected to a floating datum, has been evaluated. With this procedure we have removed the ground roll with a

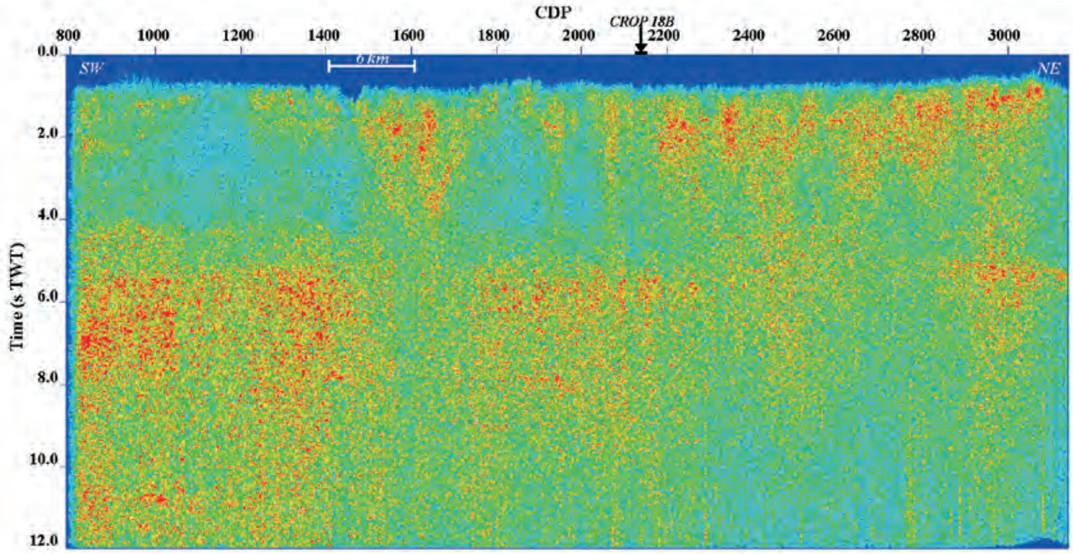

Fig. 7 - Instantaneous amplitude obtained from stacked section with the highlighted image of the lower crust and with blue transparent zones related to magma intrusions through the lower crust and within the upper crustal layers.

satisfactory result. The sequence preserved the amplitude spectrum significantly. To improve the signal-noise ratio, F-X deconvolution was applied in the shot domain after normal moveout (NMO) and field static corrections. Removing NMO and statics, a new stack velocity analysis was carried out. A refined NMO correction and residual statics, using a surface consistent procedure, were then computed and applied. Finally, a time variant filter and AGC, with a time window of 3 s, completed the stack section shown in Fig. 5.

The section is also plotted in perspective with the seismic line CROP-18B [Fig. 6; Accaino *et al.*, (2005b)] to put in evidence the correspondence of the main reflecting intervals recognisable on both lines. The high reflectivity, below 5 s TWT, is present in both sections, while the K-horizon, characterizing the neighbouring Mount Amiata geothermal field cannot be followed on the line CROP-03. The re-processing confirms a quite flat reflector around 4 s, which can be correlated with the regional interface already considered on CROP-18A and B, below the K-horizon, and associated to a lithologic change (Tinivella *et al.*, 2005).

The main reflectivity along the CROP-03 section can be identified directly in the instantaneous amplitude section (Fig. 7), presented after the seismic attributes analysis. As distinctly evidenced, the deeper part (from 5 to 8 s TWT) provides reflections with much higher energy compared with the shallow part (above 4 s). The Moho is marked by the termination of the high-amplitude continuous, fairly thick reflecting intervals at 7.5-8 s TWT. Somewhere a lack of reflected energy occurs, like the transparent vertical corridors crossing the lower section. The shallow non-reflecting zones can be explained by the presence of energy scattering by multiple magma intrusions.

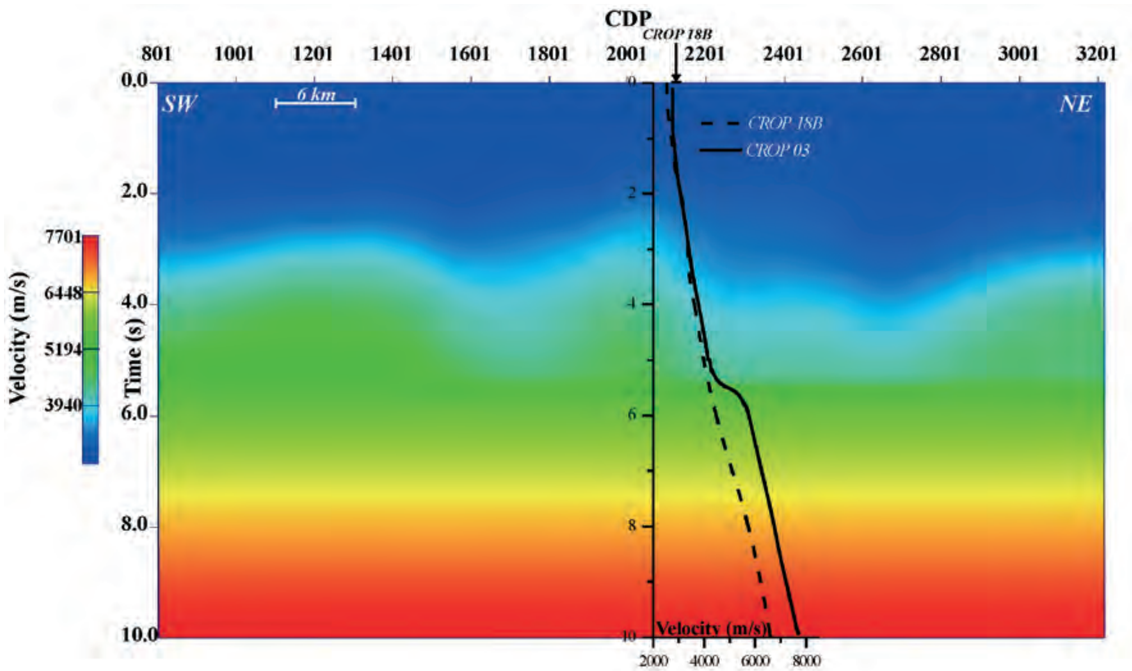

Fig. 8 - Root-mean-square velocity distribution used for the pre-stack time-migration. The arrow indicates the position of the line CROP-18B. A comparison between the RMS-velocity from the profile CROP-18B (dotted line) and CROP-03 (solid line) is presented in correspondence to their crossing.

3. Evaluation of the seismic velocities

The velocity analysis was performed by using iteratively the pre-stack migration, employing the commercial software GeoDepth©. At the first inversion step, we considered a velocity model with a constant velocity, equal to 2500 m/s. Thus the pre-stack time-migration is computed and then the Common Image Gathers (CIGs) were analysed, determining the residual move-out corrections from the semblance analysis. The model is updated by using the following rule: the flatness of an event on CIG yields a qualitative estimation of the error in the determination of the velocity model (Yilmaz, 2001). The residual moveout analysis of the CIGs and the update of the velocity model were iteratively repeated until the residual moveout was near zero. We used a layer-stripping approach, updating the velocity following the time sequence. In this way, velocity versus time is better constrained. The main advantage of using this iterative procedure is that the picking in the pre-stack domain is not necessary, particularly difficult in the case of poor land-recorded data.

The deeper velocity information (below 5 s TWT) cannot be obtained, because the maximum available offsets do not allow a detailed move-out analysis. So, we introduced a velocity gradient determined (i) from the velocities identified by refraction data (Giese *et al.*, 1981), and (ii) upgrading the seismic imaging of the pre-stack time-migration. The criterion followed was to obtain the same depth for the base of the crust in the reflection section image as defined by DSS

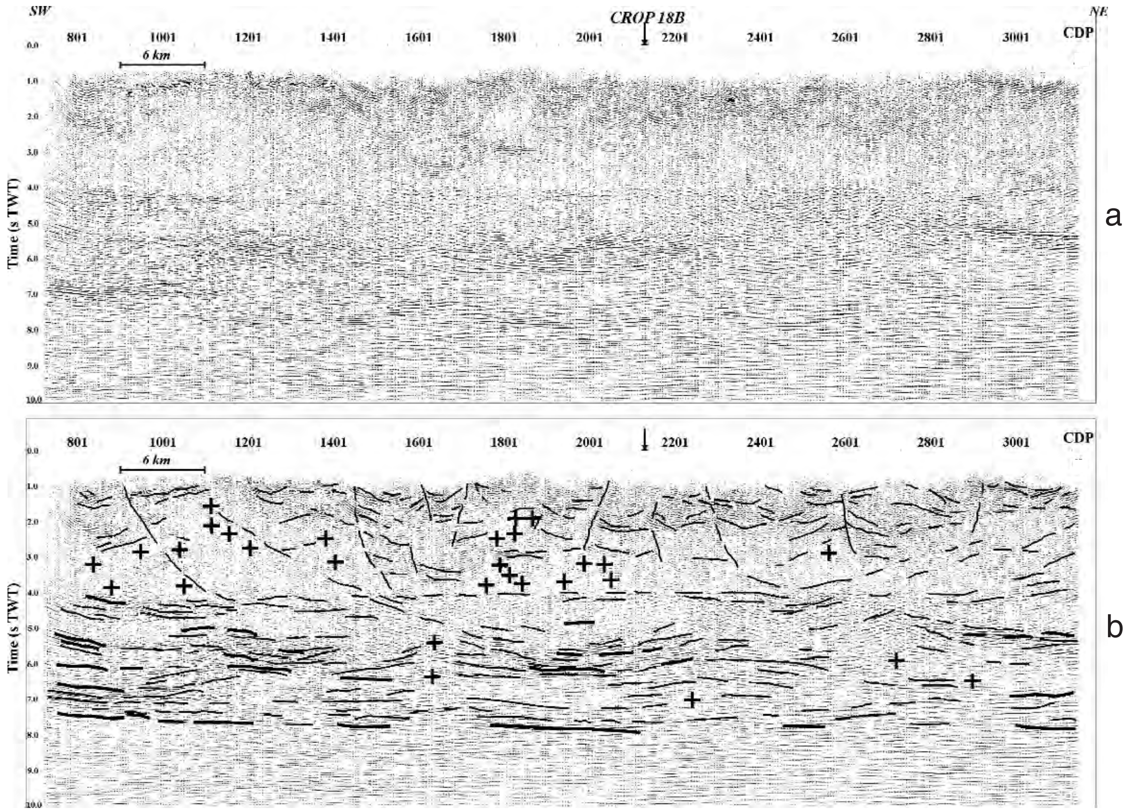

Fig. 9 - a) pre-stack time-migration; b) line-drawing. The crosses indicate the possible magmatic intrusions associated with low reflecting energy. The line-drawing of the upper crust was not controlled with the outcroppings and is only indicative of the brittle nature of that part. The main reflectors in the lower crust are indicated with thick segments. The Moho boundary corresponds to the lower termination of the lower crust lamellae. The arrow indicates the location of the line CROP-18B.

(Fig. 3), also after tests performed on an acceptable velocity-dependent migration of the data. Note that Fig. 3 reports interval velocities in each layer, whereas Fig. 8 reports root mean square (RMS) velocity functions . The two velocities are related by Dix's equation:

$$V_{\mathrm{int}}^2\left(t_n - t_{n-1}\right) = \left(V_n^2 t_n - V_{n-1}^2 t_{n-1}\right),$$

where V_{int} is the interval velocity within the layer boundaries n and $n\text{-}1$; V_n and V_{n-1} are the RMS velocities down to the layer boundaries n and $n\text{-}1$, respectively, and t_n and t_{n-1} are the two-ways times to these boundaries.

The RMS velocity model (Fig. 8), obtained by the above iterative procedure, was smoothed and a pre-stack time-migration computed (Fig. 9a), improving the correct positioning of the seismic horizons and of dipping anomalous events.

In the lower window (Fig. 9b) the line-drawing is shown. The details of the shallow part (about the first 3 km depth) can be obtained by using wells and available geological information, which

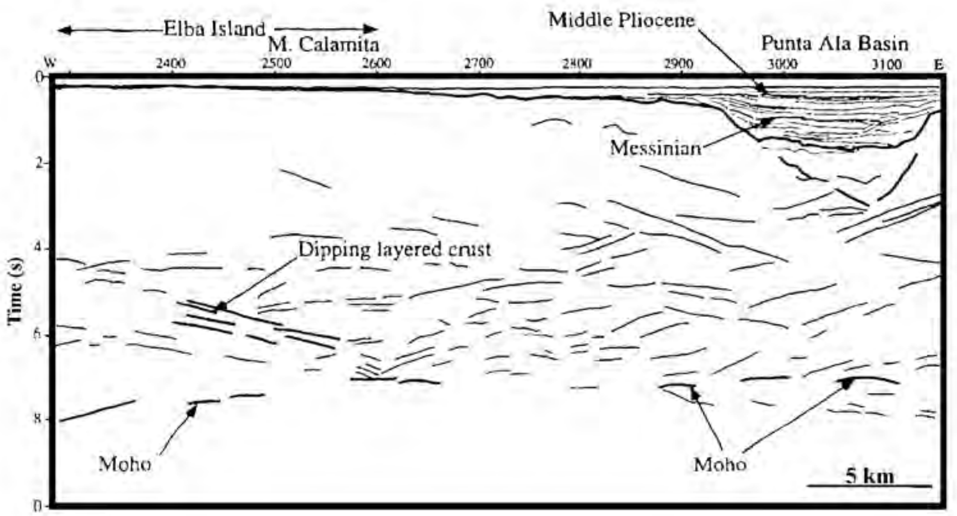

Fig. 10 - Line-drawing of the profile LISA-08 [from Mauffret *et al.*, (1999); position in Fig. 1]. Note the dipping reflection in the eastern part of the section. The lower crust interval matches the onshore data of CROP-03. The Moho is located at about 7.5-8 s TWT.

are exhaustively described in Decandia *et al.* (1998).

The termination of layered reflectivities within the lower crust is considered to coincide with the Moho discontinuity at about a 23 km depth.

The reprocessed data of the CROP-03 section do not offer the interpreter any seismic image that can allow the definition of the very intriguing features presented in the paper of Finetti *et al.* (2001). On the contrary, our data match those recorded in the Tuscan Archipelago by Mauffret *et al.*, (1999) very well (LISA project). In Fig. 10, the line drawing of the LISA-08 line, which continues inside the Tuscan Archipelago domain the lower crust structures made known on land.

In Fig. 10, the line drawing of the LISA-08 line, continue inside the Tuscan Archipelago domain the lower crust structures pointed out on land.

We already mentioned that the seismic section reveals the presence of sub-vertical conduits or fractures that can be related to the ascent of plutons from the lower to the upper crust, identified also on CROP-18B and -18A (Accaino *et al.*, 2005b; Tinivella *et al.*, 2005). To better evidence this aspect, we present a pre-stack depth-migrated section by using the final velocity model in the sector between the CDPs 1100 and 2100. Fig. 11a indicates the section where the opposite dip of the reflectors with respect to a vertical channel is marked Fig. 11b shows the similar feature, as revealed on the CROP-18B (from Tinivella *et al.*, 2005).

4. AVO analysis

Both the stacked section and the pre-stack time-migration sections reveal the presence of a quite flat reflector at a depth (~ 9 km – 4 s TWT) already recognised in the other seismic lines acquired in Southern Tuscany (Accaino *et al.*, 2005a, 2005b; Tinivella *et al.*, 2005). To identify if

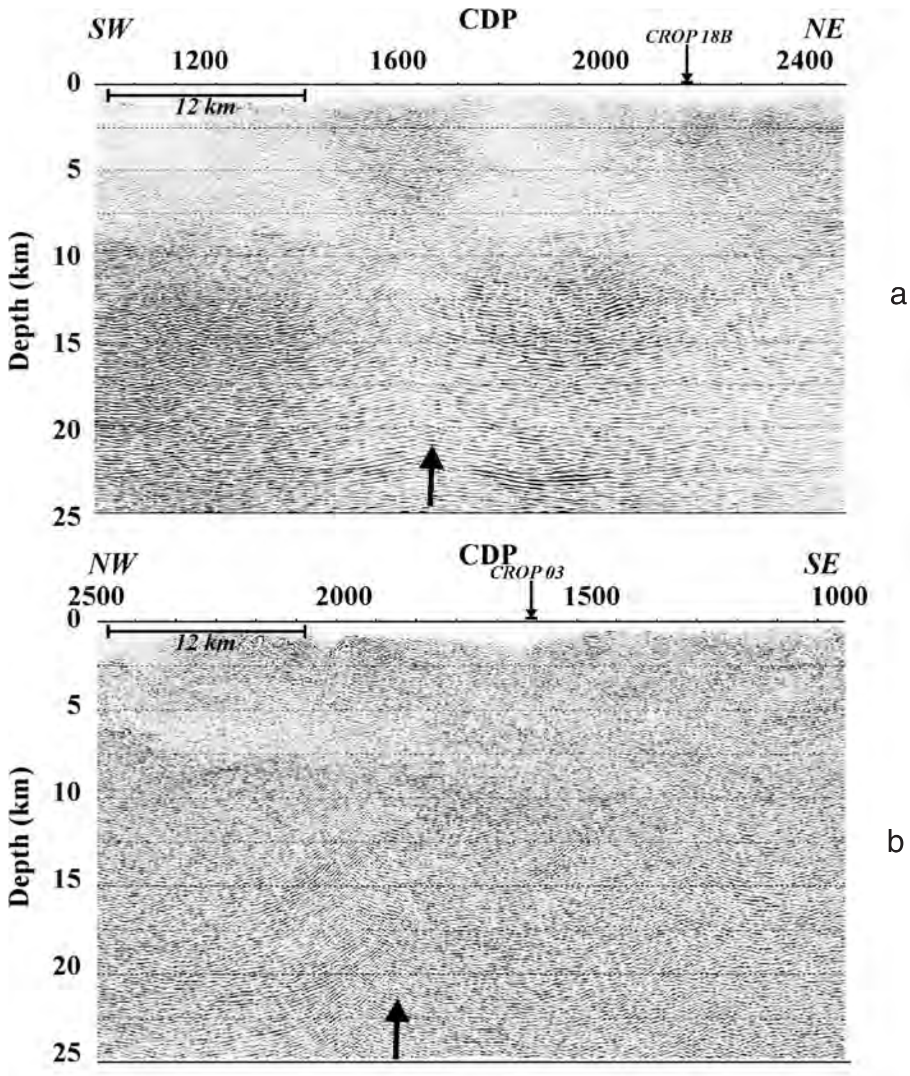

Fig. 11 - a) Detail of the pre-stack depth-migration in the central part of the line CROP-03 as recognizable in this figure and in Figs. 7, 9 or 12 looking at CDP numbers. b) image of a vertical channel obtained on the line CROP-18B (Accaino *et al.*, 2005a).

the reflections are related to the lithology or to pore fluids changes, we completed an amplitude versus offset (AVO) analysis by using the GeoDepth© (Paradigm Geophysical) commercial software. We followed the Aki and Richards's (1980) approach for the linearization of the Zoeppritz equations. The main inputs are the pre-stack true amplitude data, the velocity model, and the acquisition geometry, while the outputs are two sections that reported the P- and S-wave reflectivities, linked to the main P- and S-wave velocity contrasts. For more details about the procedure see Tinivella *et al.*, (2005).

The P- and S-wave reflectivity sections are shown in Fig. 12 and reveal (i) the presence of a

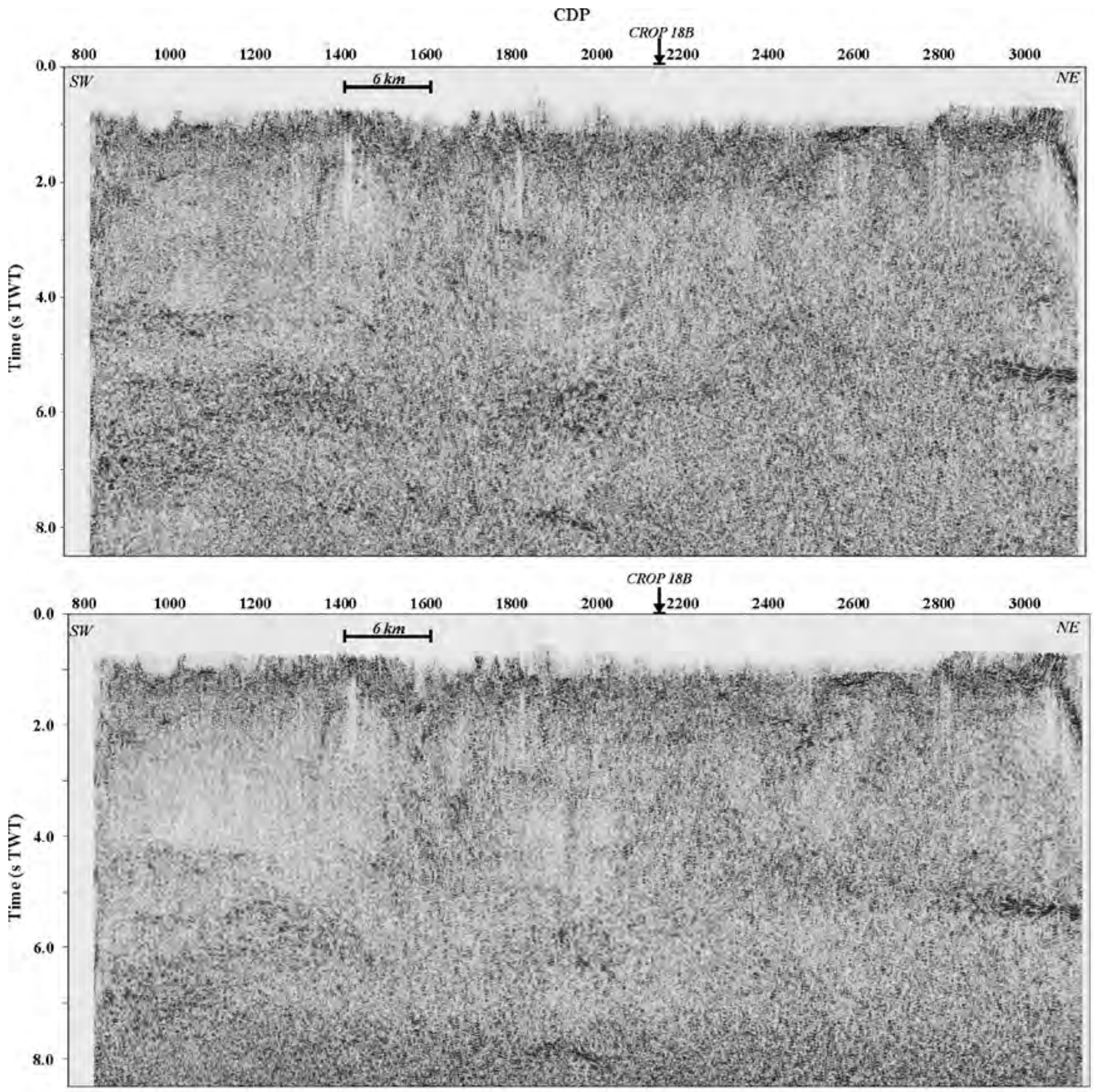

Fig. 12 - P-waves reflectivity (upper section) and S-waves reflectivity (lower section) after AVO inversion. The arrow indicates the intersection with the line CROP-18B. Note the low reflection zone in the upper part of the two sections and the energy present in both sections at about 4 s TWT.

low S-energy above 4 s TWT [probably related to fluids in the pore space in agreement with Tinivella *et al.*, (2005)], and (ii) the relatively high-energy reflectors below 5 s TWT in both sections. Thus, the AVO results confirm the absence of important lithologic changes and support the presence of fluids above 4 s TWT.

To help the interpretation of the AVO results, we evaluated the ratio between the P-wave and S-wave velocity reflectivity sections, providing information about the Poisson ratio contrast. The results are reported at selected CDPs in the stacked section (see Fig. 5) and indicate the presence of the reflector at about 4 s TWT with high P/S reflectivity ratio most likely caused by a lithologic change.

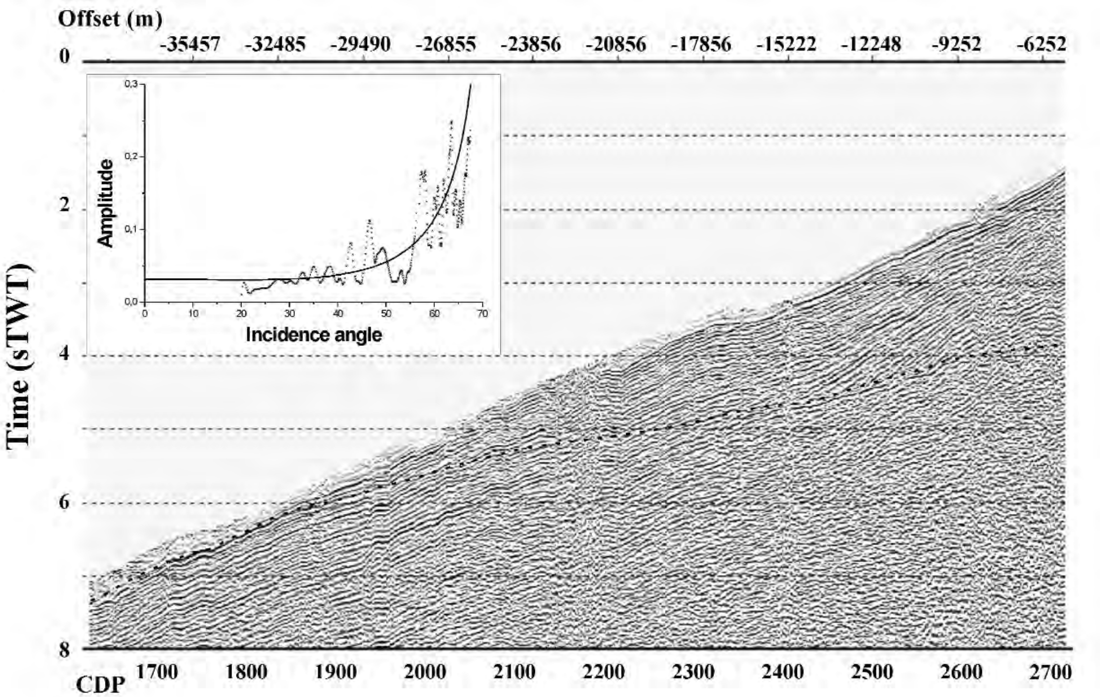

Fig. 13 - Wide-angle records after amplitude compensation (see text for details). The position of shot and receivers are indicated in Fig. 5. The dotted line points to the picked reflector. The insert shows the energy versus incidence angle extracted from the data (stars) and the best theoretical fit (continuous line).

5. Wide-angle shots

During the acquisition of the seismic line CROP-03 several high energy shots were used to record expanding spread data (Gualtieri *et al.*, 1998). In this paper we consider three shots located at CDP 2885 position (Fig. 5) and recorded until 40 km of distance. The details of the acquisition geometry are reported in Table 1. Because different source energy was used during the three acquisitions, we considered the overlapping traces to rescale and equalize the amplitudes (Fig. 13). The reflector at about 4 s TWT is recognizable at near offsets.

Picking this reflector we extracted the amplitudes, averaging signal within a window wide 0.05 s. The amplitudes were rescaled applying a geometrical spreading correction considering the

Table 1 - Acquisition geometry of the three wide-angle shots.

Dynamite (kg)	First offset (m)	CDP at first offset	Last offset (m)	CDP at last offset
80	5412	2706	16,837	2326
150	16,537	2335	27,994	1954
200	27,694	1963	39,051	1586

effects of the offsets (Ursin, 1990) and utilizing the velocity model obtained for the line CROP-03. The offset was converted into incidence angles by using the velocity information. Finally, the amplitudes versus incidence angles were rescaled with respect to the maximum amplitude and filtered. Note that the extracted amplitudes are not precisely related to the reflection coefficients at the selected reflector, because the reflection points at different offsets are broadly distributed along the reflecting interface. Observing the geometry (quite flat) and the reflectivity of the interface (quite constant strength amplitude, as evidenced in Fig. 7) along the seismic line, we can suppose that the assumption of the coincidence uncertainty of the reflection point at all offsets has the same order of magnitude of other errors included in the analysis, such as the velocity model. These amplitudes are related to the contrasts of the physical parameters that caused the reflection, i.e. the P- and S-waves velocities and density, by using a theoretical model for viscoelastic isotropic media [single-phase model: Carcione, (1997)]. The algorithm for determining these parameters is the least-square method; we estimated the physical parameter contrasts minimising the function that represents the square difference between the amplitudes obtained from seismic data and those obtained from the theory. The results (see the insert in Fig. 13) indicate that P-wave reflectivity is controlled by density variation, that is by the presence of changes of lithology.

6. Discussion

The comparison of the velocity functions of CROP-03 and CROP-18B (Fig. 8) in proximity of their crossing point indicates the excellent match between the two profiles above 5 s TWT, confirming the goodness of the two approaches (tomographic inversion, used in CROP-18, and CIG analysis). On the contrary, below 5 s the two velocity profiles present strong differences. Because the two seismic lines are dip (CROP 03) and strike (CROP 18B) lines with respect to the main regional structures (Apenninic and anti-Apenninic), the velocity difference can be related to anisotropy [see also Margheriti *et al.*, (1996)]. The presence of anisotropy in the lower crust has been evidenced by many authors [see for example the recent papers of Godfrey *et al.*, (2002), Shapiro *et al.*, (2004), Wilson *et al.*, (2004), Lueschen, (2005)]. Weiss *et al.* (1999) point out that the seismic anisotropy is particularly strong in layered structures, such as the seismic lamellae that are typical of the lower crust. They reached the conclusion that the deformation that occurs by slip on discrete faults in the upper 15 km of the crust is distributed as strain over a progressively broader region below, a typical situation for Tuscany. Millahn *et al.* (2005) have found in the Tauern Window (Transalp profile crossing the Eastern Alps) a 10% velocity anisotropy attributed to E-W elongation of the rock texture caused by paleo-stresses, that is N-S compression and E-W extrusion and stretching.

In our case, the percentage anisotropy (A) is between 10 and 20%, calculated by using the following relationship:

$$A = (V_{03} - V_{18B})/V_{03}$$

where V_{03} is the maximum velocity in the WSW-ENE direction and V_{18B} (NW-SE direction) is the minimum velocity. Seismological studies already revealed the presence of anisotropy in southern Tuscany (Margheriti *et al.*, 1996) with an approximately E-W orientation of the high-velocity direction attributed to the stretching of the upper mantle below the Tyrrehenian Sea.

The seismic sections and the AVO analyses on both wide-angle and near-vertical reflection seismic confirm the presence of a regional reflector at about 4 s (i.e., ~ 9 km depth) that can be positioned at the base of a more transparent interval. The AVO inversion confirmed also the occurrence of non-reflecting sectors that can be justified accepting the occurrence of magma ascent along vertical conduits within the lower crust, and of energy scattering by intrusions in the upper crust. The presence of overpressured trapped fluids can explain the reflectivity in the upper crust (K-horizon) revealed on the neighbouring geothermal fields (Tinivella *et al.*, 2005).

Dini *et al.* (2005) presented the assumption that crustal sources for magmas (granites) are located at 14-23 km depth (5 to 8 s TWT). They estimated an efficient, fast magma extraction from that source, while mantle derived magmas are supposed to be present at the base and in the lower crust (again the interval between 14 and 23 km).

Heat flow sources are confined to at depths of 3 km or less, contributing to the very high values in correspondence of the geothermal fields and at short wavelengths: local advection of hot fluids. To take account of regional conductive heat transfer with larger wavelengths, the contribution of deep sources (at 8 km or larger) is obligatory (Bellani and Della Vedova, 2003).

The intrusion of mantle-derived magmas is commonly referred to as magmatic underplating and has been studied by Sinigoi *et al.* (2003, and references therein) along the considerable outcroppings of lower crust bodies in the Ivrea-Verbano zone (Southern Alps in NW Italy). The heat released at the base of the crust induced anatexis in the overlying crustal rocks with production of granitoid melts, quickly migrating towards higher crustal levels, leaving behind progressively depleted restites, the residual product of the fusion processes, representing high-grade metamorphic and refractory rocks. A sharp distinction between denser materials and less dense anatectic melts and metamorphosed volcano-sedimentary rocks is the end result.

Seismic images can support the model suggested by Serri *et al.* (1993), i.e., the partial melting of the uppermost mantle and the delamination processes, even if they relate it to the subduction processes of the Adriatic continental lithosphere (Doglioni *et al.*, 1999, 2005). Our analysis of seismic data does not support the existence of active subduction zones activated by external convergence forces.

Upper mantle mafic intrusion in the lower crust layers is consistent with volcanological observations as well as with reflection seismic images and with a layered lower crust marked by velocity (and density) alternations in the DSS profiles. In the upper crust volcanic products, fluids and overpressures are accepted by our analysis. From the theoretical modelling, the porosity in the overpressured zones turn out to have a value near 5.5% (Tinivella *et al.*, 2005), with a consequent decrease of the density that can partly justify the negative gravity Bouguer anomalies observed in correspondence of the main geothermal fields.

7. Conclusions

The formation of the Tyrrhenian basin and the formation and evolution of the Apennines are the effect of two asthenospheric mantle domes. These domes should be considered, to a large extent, as a physical and chemical transformation of the pre-existing lithospheric mantle. The magma formation was induced by heat and mantle metasomatic fluid supply from deep sources; it is produced at the top of the ascending mantle plume and is accreted at the base of the crust-underplating or continental crust accretion in a rift tectonic environment.

In the southern Tuscan delamination of the uppermost lithosphere beneath the base of the crust is observed by tomographic inversion of surface S-waves from seismological data, defining the presence of a mantle wedge. Modelling shows that lithospheric deformations are driven by buoyancy forces interrelated with this mantle wedge.

In conclusion, our integrated analyses of geophysical data are in agreement with the above described hypothesis and, in particular, we can recognise in the seismic images presented here the rising of hot melts from below, with unstable equilibrium among rock physical properties, temperatures, fluid pressures and confining pressures at different levels of the crust, source and motor of the geothermal energy resources.

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