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Structure and rheology of lithosphere in Italy and surrounding

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\textbf{ABSTRACT}

We define the structure and rheology of the lithosphere in Italy and surrounding, combining the cellular velocity models derived from nonlinear tomographic inversion with the distribution vs. depth of hypocentres to assess the brittle properties of the Earth’s crust. We average, over cells sized 1° × 1°, the mechanical properties of the uppermost 60 km of the Earth, along with seismicity, grouping hypocentral depths in 4-km intervals. For most of the cells, the earthquake energy is concentrated in the upper crust (4–12 km). For some regions, where orogenic processes occur, the release of earthquake energy is shallower and limited to the uppermost 10 km of the crust. Ambiguities in the structural models are minimized considering the hypocentral distribution, mainly to define the location of the Moho boundary, when its identification, based on shear-wave velocities, is not straightforward.

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\textbf{Introduction}

Anderson (2007) gives a good and long-overdue treatment of how the Earth’s crust, along with seismicity, grouping hypocenters to assess the brittle properties of the crust. We average, over cells sized 1° × 1°, the mechanical properties of the uppermost 60 km of the Earth, along with seismicity, grouping hypocentral depths in 4-km intervals. For most of the cells, the earthquake energy is concentrated in the upper crust (4–12 km). For some regions, where orogenic processes occur, the release of earthquake energy is shallower and limited to the uppermost 10 km of the crust. Ambiguities in the structural models are minimized considering the hypocentral distribution, mainly to define the location of the Moho boundary, when its identification, based on shear-wave velocities, is not straightforward.

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earthquakes with unspecified kind of magnitude are plotted as purple circles.

We used the revised ISC (2007) catalogue for the period 1904–June 2005 and, using the relation of Richter (1958) log$_E$ = 1.5$M_s$ + 11.4, for each cell we computed the total energy released by the earthquakes, grouping hypocentres in 4-km intervals. The value of $M_s$ is either directly taken from the ISC catalogue or computed from currently available relationships between $M_s$ and other magnitudes (Peishan and Haitong, 1989). As the hypocentral depths did not exceed 60 km there was no need to correct $M_s$ for the focal depth (Herak et al., 2001). The energy of the earthquakes, for which only $M_d$ is known, is calculated as follows. The regression relation constructed for the study region from the earthquakes for which $M_d$ and $M_l$ are known, is used to determine $M_l$ which is then converted to $M_s$. The relationship obtained $M_s = 1.12M_d - 0.76$ is in agreement with similar relationship computed by Gasperini (2002) for the Italian region. The earthquake
energy distribution vs. depth (log \( E - h \)) is shown in Figs 1–3 in the right hand graph for each cell. The logarithm of the energy for the grouped hypocentres was normalized to the maximal value \( \log E_{\text{max}} \) for each cell (written on the horizontal energy axis). In each cell we considered all earthquakes listed in the ISC catalogue with depth and magnitude specified (red bar histogram) and only the earthquakes for which the depth had not been fixed a priori (black line histogram).

**Discussion**

The \( \log E - h \) distribution for most of the cells is concentrated at depths of 4–12 km (Figs 1–3), similar to the distribution of earthquake frequency reported by Meletti and Valensise (2004) and with the maximum slightly shallower than in the distribution of hypocentres in the continental crust shown by Ponomarev (2007). The energy histograms of all events (red bars) and for the events for which the depth has not been fixed a priori (black line bars) have a similar behaviour (with some obvious exceptions for the cells with weak seismicity). Consistent with a classic Coulomb/Byrlee (brittle/ductile) transition, where the rheology and mechanical properties of rocks follow the Sibson’s law for the upper crust, and a power law creep in the lower crust, the earthquake energy is concentrated in the upper crust (4–12 km); only in a few cells in the transition of the central-western Alps (f-3, f-2, f-1), the eastern Alps-Dinarides (e4, e5, d5), and the Dinarides-Hellenides (a9, A9) the earthquake energy is concentrated in the uppermost 10 km of the crust, generally with a maximum in the surface layers. These are thrust zones where orogenic processes are in progress, with predominant compression (Scandone and Stucchi, 2000), as it is evident from the focal mechanisms (Meletti and Valensise, 2004; Guidarelli and Panza, 2006; Pondrelli et al., 2006). The energy release in the crust is dominant for the uppermost 60 km even in the zones with intermediate and deep seismicity (c2, b8, a9, A5, A9, C9, D3, D4, D5).

The ambiguities in our structural model are reduced with the addition of the hypocentre information, mainly in defining the location of the Moho (Panza et al., 2007a). The cell a2, under the Albani hills and the Tyrrhenian offshore, is a clear example, where the definition of Moho is difficult using \( V_s \) alone (Fig. 2). In this cell and in most of the southern Tyrrhenian Sea \( V_s \) values usually observed in the crust (\( \leq 4.0 \) km s\(^{-1}\)) are interpreted as mantle material (Panza et al., 2003,
2007a,b), with a relatively high percentage of partial melting (Bottinga and Steinmetz, 1979; Green and Falloon, 2005). The upper crust in a2 reaches a depth of about 7 km with $V_s$ about 3.3 km $s^{-1}$, and it overlies a 30-km-thick layer with $V_s$ about 3.9 km $s^{-1}$. This layer is on top of a 20-km-thick high-velocity lid. According to the log $E-h$ distribution and other data compilations (e.g. Nicolich and Dal Piaz, 1990) the 30-km-thick layer, with $V_s$ about 3.9 km $s^{-1}$, can be reasonably assigned to the crust only up to a depth of about 25 km, where the recorded seismicity stops. The shut-down of the distribution of seismicity at a depth of about 25 km (Fig. 2) leaves room for an aseismic, 12-km-thick, mantle layer. Therefore, the 3.9 km $s^{-1}$ layer can be interpreted as high-velocity lower crust down to a depth of about 25 km, above a very low-velocity, soft mantle. The neighbouring cell a3 (Fig. 2) exhibits a negative velocity gradient within its 20-km-thick crust that overlies a low-velocity uppermost mantle layer with a $V_s$ of about 4.2 km $s^{-1}$ and a thickness of about 30 km. Below this layer, $V_s$ increases with depth to an average of 4.35 km $s^{-1}$.

In e4 (Fig. 1) below about 23 km of depth, the 30-km-thick layer with $V_s$ $\sim$4.0 km $s^{-1}$ can be lower crust (Moho boundary at $\sim$53 km depth) or soft mantle (Moho boundary at $\sim$23 km depth). The Moho depth defined by other studies varies between 35 and 45 km (Nicolich and Dal Piaz, 1990; Marone et al., 2004; Dezes and Ziegler, 2005). The question can be settled considering the seismic energy distribution (Fig. 1) that is concentrated in the uppermost 30 km with some weak events below this depth: it is reasonable to assign the first 7 km of the 30-km-thick layer to the brittle continental crust, placing the Moho depth at about 30 km, and to interpret the remaining 23 km as very soft mantle.

The interpretation of $V_s$ models in the cells with recent volcanism (b1, b2, a2, a3, A-2, A-1, A3, A4, B-2, B-1, B1, B2, B4, C3, C4, C5, D2, D5) is given by Panza et al. (2007a). The seismicity, even if weak, in B2, B3, and B4 (Fig. 3) plays a key role in the definition of the Moho that can be placed just below the deepest event in each cell, namely, at depths of 11 km (B2), 13 km (B3), and 12 km (B4). In B2, the remaining 5 km, with $V_s$ $\sim$4.05 km $s^{-1}$, represent mantle material, cooler than the underlying very hot 8-km-thick mantle layer with average velocity 3.15 km $s^{-1}$ and density 3.1 g cm$^{-3}$, followed by a layer with velocity 4.2 km $s^{-1}$. Similarly, in B3 the remaining 5 km of the layer...
with material 3.6–4.0 km s$^{-1}$ is mantle material cooler than the underlying, very hot, 8-km-thick mantle with average velocity 3.1 km s$^{-1}$ and density of 3.1 g cm$^{-3}$, followed by a layer with velocity 4.2 km s$^{-1}$. In B4, the hypocentre’s distribution and gravity modelling lead to two distinct intermingling 11.5-km-thick layer with velocity 3.3 km s$^{-1}$ and density 3.15 g cm$^{-3}$ as mantle, followed by a layer with average velocity 4.2 km s$^{-1}$. There is no seismicity in B1 but, by analogy with the neighbouring cell B2 and with the results of gravity modelling, the layer with average velocity 3.6 km s$^{-1}$ is defined as mantle with density 3.1 g m$^{-3}$. It overlies a layer with velocity 4.2 km s$^{-1}$.

Another interesting case is observed in the area of Stromboli and Messina strait to Southern Calabria, in C5 (Fig. 3). The nature of the crust is difficult to define as a layer with $V_s$ in the range 3.8–4.0 km s$^{-1}$ reaching a depth of about 44 km overlies high-velocity mantle material. The analysis of the seismic energy distribution (details given in Panza et al., 2007a) leads to define this feature as a two-faced (Janus) crust/mantle layer. If we consider only the hypocentres at sea, this layer is totally aseismic and therefore it can be reasonably assigned to the mantle, consequently the crust has an average thickness of ~17 km. On the other hand, if all events in C5 are considered, the ‘Janus’ layer is occupied only by hypocentres located either in Sicily and Calabria (continental area) or close to their shoreline (Panza et al., 2007a). In such a case the layer can be reasonably assigned to the brittle continental crust which turns out to be ~40 km thick.

Finally, in d0, c1 (Fig. 2), D1, D2 and D6 (Fig. 3), seismic activity is recorded also in the uppermost mantle layer, where $V_s$ is relatively low, but where the average heat flow does not exceed 65 mW m$^{-2}$ (95 mW m$^{-2}$ for c1) according to Hurting et al. (1991). Gravity modelling in this region confirms that these layers belong to the mantle. The brittle behaviour of this uppermost mantle material, with average velocities 4.0–4.3 km s$^{-1}$, can be an example of the magma-assisted rifting model discussed by Buck (2004) and the eclogite mantle ‘engine’ by Anderson (2006).

From the integrated use of geophysical, petrological and geochemical data it is possible to confirm the presence of three processes that probably govern the present lithosphere dynamic in the Italian Peninsula and surroundings: delamination in North and Central Apennines (e.g. d0, d1, c1, b0, b1, a3) with mantle wedges that fill in the space left by the removed lithosphere along the W-directed subduction zones (Doglioni, 1991; Chimera et al., 2003; Panza et al., 2007b); slab detachments (e.g. A5, A6, B7), with low crust delamination and formation of orthopyroxene-rich uppermost mantle layer with strong crustal signatures (Lustrino, 2005), to continuous subduction in the Southern Apennines; slab roll-back and tearing with sideways asthenospheric flow through slab-windows in the Calabrian Arc (Panza et al., 2007a) with likely slab-detachment (e.g. C4, C5, D5, D6).

Conclusions
The synoptic representation of $V_s$ models of the uppermost 60 km of the Earth and of the log$E$–$h$ distribution is used to define the Moho when its identification based on $V_s$ alone is ambiguous. For most of the cells, the earthquake energy released is maximum in the depth range of 4–12 km, i.e. mainly in the upper crust. For some regions where orogenic processes are in progress, the release of seismic energy is concentrated in the uppermost 10 km of the crust. The brittle behaviour of the uppermost mantle, with relatively low average velocities accompanied by relatively low heat flow, is well consistent with the magma-assisted rifting model and the eclogite mantle ‘engine’.

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