Current knowledge on the crustal properties of Italy

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Abstract
The recent advances in experimental petrography together with the information derived from the super-deep drilling projects have provided additional constraints for the interpretation of refraction and reflection seismic data. These constraints can also be used in the interpretation of magnetic and gravity data to resolve non-uniqueness. In this study, we re-interpret the magnetic and gravity data of the Italian peninsula and neighbouring areas. In view of the constraints mentioned above, it is now possible to find an agreement between the seismic and gravity models of the Central Alps. By taking into account the overall crustal thickness, we have recognized the existence of three types of Moho: 1) European which extends to the north and west of the peninsula and in the Corsican-Sardinian block. Its margin was the foreland in the Alpine Orogeny and it was the ramp on which European and Adriatic mantle and crustal slices were overthrusted. This additional load caused bending and deepening and the Moho which now lies beneath the Adriatic plate reaching a maximum depth of approximately 75 km. 2) Adriatic (or African) which lies beneath the Po plain, the Apennines and the Adriatic Sea. The average depth of the Moho is about 30-35 km below the Po plain and the Adriatic Sea and it increases toward the Alps and the Tyrrhenian Sea (acting as foreland along this margin). The maximum depth (~ 50 km) is reached in Calabria. 3) Pery-Tyrrhenian. This is an oceanic or thinned continental crust type of Moho. It borders the oceanic Moho of the Tyrrhenian Sea and it acquires a transitional character in the Ligurian and Provençal basins (< 15 km thickness) while further thickening occurs toward the East where the Adriatic plate is overthrusted. In addition, the interpretation of the heat flow data appears to confirm the origin of this Moho and its geodynamic allocation.

Key words *deep crust properties – integrated interpretation*

1. Foreword

The term «crustal geophysics» is used in this paper to indicate depths greater than approximately 8-10 km, that is, the lower crust. The scientific progress achieved in recent years derives mainly from the results of 1) experimental petrography at high temperatures and pressures (fig. 1) and 2) the direct access to the rocks of the super-deep well of the Kola peninsula (fig. 2).

– Kern (1982) has summarized more than a decade of laboratory results on *P* and *S* wave velocities determined at temperature and pressure conditions of the lower crust and upper mantle. He showed that:

1) The propagation of elastic waves in igneous and metamorphic rocks depends heavily on 1) the existence and orientation of fracture planes and 2) the amount of pore-spaces. In general, strong external pressures will close the microfractures whereas an increase in temperature will increment the porosity owing to the different thermal expansion properties of the single mineral constituents. However, this thermal microfracturing will be progressively reduced as the lithostatic pressure increases. It has been found that a minimum of 1 Kbar confining pressure is needed to prevent the opening of fractures at a temperature-increase of 100 °C.

2) The dehydration reactions produce solid-fluid systems which lead to significant reductions in effective pressure (P_{eff}):

$$P_{\text{eff}} = P_{\text{lith}} - P_{\text{fl}} \tag{1.1}$$

The porous fluid pressure (P_{fl}) produced internally as a consequence of water release generates vesicles and spaces which remain open until the maximum experimental pressure (6 Kbar) is reached. This pore opening is responsible for the reduction in seismic velocity.

3) It appears that the velocity-density law should be considered quite approximate and it may display considerable inconsistencies.

4) The lack of a unique law that relates velocity and density derives from the different modes of variation of the above mentioned physical parameters (fig. 1).

– The results presented by Kern (1982) are corroborated by the observations made in the Kola well (fig. 2).

It is generally observed that seismic velocity increases with depth (*i.e.*, with lithostatic pressure, P_z) in a sedimentary sequence. A similar velocity increase occurs in crystalline rocks as the porosity decreases at the increase of the lithostatic pressure. However, as the rock dehydration process starts, the overall volume of the solid constituents decreases and microfracturing is favoured. In the Kola well, microfracturing alone was responsible for a three, four times porosity increase when compared to the unfractured neighbouring rocks. This increase

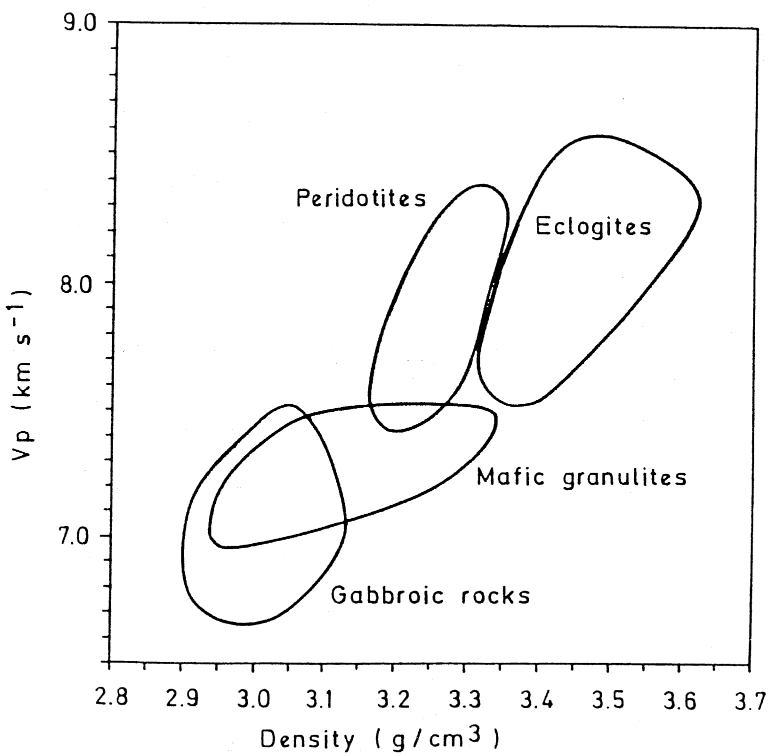

Fig. 1. Compressional wave velocity and density experimentally determined at 6 Kbar (Mengel and Kern, 1991).

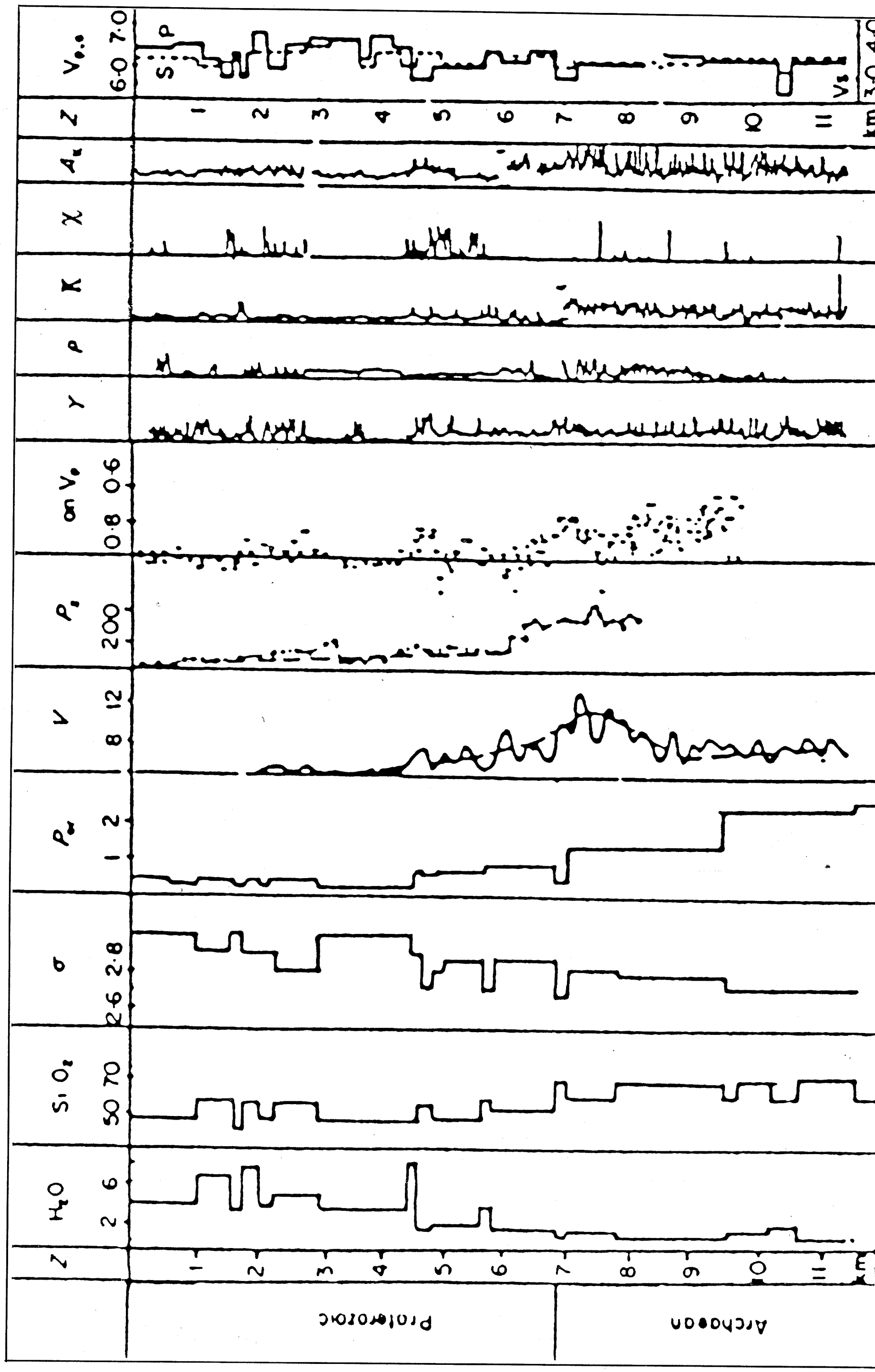

Fig. 2. Geological-geophysical data from the super-deep drilling of Kola (Pavlenkova, 1992). Z = depth (in km); H_2O, SiO_2 contents in %; V = drill-hole diameter; an = anysotropy. Coring results: γ = gamma; ρ = electrical resistivity; K = neutron; χ = magnetic susceptivity; A_K = acoustic velocity.

in porosity will also yield higher permeabilities. The freed H_20, trapped in the interstices of the fractured rocks, has increased the total volume by 1.7% and decreased the density from 2.9/3.1 to 2.8/2.9 g/cm^3.

In general, the released fluids will migrate or be trapped according to the permeability of the crystalline medium. Trapped fluids tend to increase the level of (micro-) fracturing and reduce the seismic velocity. In this regard, impermeable diabases were found between 2800 and 4600 m depth.

What has just been described is probably the single most important result that has emerged from the Kola project. It has demonstrated that seismic discontinuities do not necessarily reflect lithological variations but they may well be caused by variations in rock physical properties. In particular, the Kola well has shown that three mechanisms contribute to the formation of abrupt velocity changes, primary ingredients for seismic reflections. First, seismic velocity changes induced by different lithologies or petrologies. Examples include horizons of sedimentary or tufaceous layers. Secondly, extended microfracturing associated to tectonics can also lead to substantial velocity variations and originate seismic reflectors. It appears that both lithological and tectonically induced types of reflectors contribute to the current geologic structure of an ancient crust.

The third type of reflectors has been discovered in the metamorphic complex drilled by the Kola well. It was found that dehydration: 1) decreases the volume of the solid constituent; 2) increases the fluid pore pressure; 3) increases the overall initial volume; 4) reduces density; 5) favours microfracturing; 6) increases the overall porosity and permeability.

It appears that low velocity layers can result from this third mechanism that is, a change of the rock physical properties and not a compositional change.

The first two types of seismic discontinuities appear to occur in the upper crust and they are detected by means of near vertical reflection (NVR) surveys and, more sporadically, by wide angle (WA) data.The third kind of discontinuities are generally sub-horizontal and tend to occur at greater depths. They are generally imaged by both NVR and WA.

2. Crustal geophysics

2.1. *Potential methods (gravity and magnetics)*

Because the data acquisition and interpretation for both gravity and magnetics is similar, we will overview them together in this study.

We will also restrict our analysis to the long wavelength features which are indicators of the deep structure. We also remark that the interpretation of the potential data is highly nonunique and, whenever possible, we will avail of the information provided independently by other geophysical data sets or by the surface geology.

Bouguer gravity anomalies – The first map was compiled by the Italian Geodetic Commission in 1963 (1:1 000 000, 3000 stations). Chronologically, this map was followed by the one prepared within the CNR-Progetto geodinamica, Modello Strutturale d'Italia (1:1 000 000, OGS, 1973). Currently, a 1:500 000 map is available which made use of 300 000 station measurements and was prepared by both CNR and AGIP.

The main features of the regional gravity field can be summarized as follows (fig. 3):

a) Positive Bouguer anomalies, indicators of thinned crust (~ 20 km), have been mapped along the Ligurian and Tyrrhenian sides of the peninsula. In some places the Moho appears to be very shallow (*e.g.*, in the Tyrrhenian and Ligurian Seas, the Sicily Strait and the Ionian Sea; see fig. 12). Positive Bouguer anomalies have also been found in the Ivrea zone and of the Euganei-Lessini-Berici hills.

b) The following negative Bouguer anomalies have been mapped: 1) the very long wavelength anomalies that occur in the Alps. This feature indicates the existence of well defined roots and of LVZ; 2) some intermediate wavelength anomalies which extend from Piedmont to Sicily and indicate the presence of sedimentary basin (hydrocarbons; see fig. 4).

c) The strong gradients of the gravity field indicate that the Moho has been severely faulted with up to 10 km offsets. This occurs along the entire Apennine chain at the transition between the Adriatic and the Tyrrhenian domains. In fact, the load of the Tyrrhenian crust has downfaulted the thinned Adriatic crust.

d) The experimental results obtained by Mengel and Kern (1991, fig. 1) have also led to an explanation for the Milan positive anomaly of the Alps-Apennines on the EGT geotraverse. According to the first hypothesis (Werner and Kissling, 1985; Schwendener and Müller, 1990), a dense body is located at depths of 80-100 km whereas, in the second one (Cassinis *et al.*, 1990), the anomaly can be modeled by a density distribution in agreement with the DSS model of Buness and Giese (1990, fig. 5). The gravimetric model of Buness (1992) has shown that by taking into account the experimental data of Mengel and Kern (1991) a good fit to the observed anomalies can be obtained (fig. 6).

Magnetic anomalies – A total of 260 000 km of aeromagnetic survey was completed by AGIP between 1970 and 1980. In the subsequent years, some smaller parts at the northern border which had remained uncovered by the aeromagnetic profiles were surveyed by the CNR. These data have allowed the magnetic bodies within the sedimentary rocks to be mapped to produce a detailed map of the magnetic basement (fig. 7). Overall, this basement displays many similarities with that imaged by seismic data although consistent differences occur in parts of the peninsula characterized by un-susceptible metamorphic formations.

2.2. *Deep seismics*

Apart from the two profiles ECORS-CROP in the Western Alps (1986-1987) and NFP.20-CROP (1988-1989) in the Central Alps (results published by Roure *et al.*, 1990 and Blundell *et al.*, 1992), NVR data (fig. 8) of the Italian peninsula are not available yet. However, the first profile carried out on the peninsula (CROP-04, 1989-1991) is still in the seismic processing stage while the second one (CROP-03) has just been acquired. Similarly, the marine seismic data are also in the processing stage. At this point, it appears that our knowledge of the crust in Italy has to be based on the extensive DSS data acquisition program that was carried out and completed by the Italian Explosion Seismology Group (CNR) in the years 1956-1981 which involved data for a total of 20 135 km (fig. 9). This program was followed by the DSS profiles of the European Geotraverse-Southern segment (EGT-S: Alps-Tunisia) from 1982 to 1985 (4685 km; fig. 10). The analogic instruments of the '60s have been replaced by digital ones and most of the analogic data have been digitized, the acquisition and processing technologies have been improved, so that it is possible to thrust to the crustal geophysical results integrated by DSS constraints we are now presenting.

3. Crustal features

3.1. *Basement*

The following seismic features characterize the crystalline (metamorphic) basement. It is transparent to the transmission of seismic waves, rigid and fractured so that there are few, scanty and dispersed reflectors. As it has been proved by the Kola and KTB experiments, the imaged horizontal reflectors cannot belong only to those of the first or second type discussed above but they must be of the third type (*i.e.*, modification of the mechanical properties or of the physical state).

It is generally assumed that the top of the crystalline basement corresponds to the depth having 6.1 km/s of longitudinal wave velocity, v_p. However, this limit is rather uncertain in that the sedimentary rocks may also reach similar v_p velocities and, therefore, the transition between the sedimentary cover and the crystalline basement is often ill-defined. In a similar manner, the basement defined by the mag-

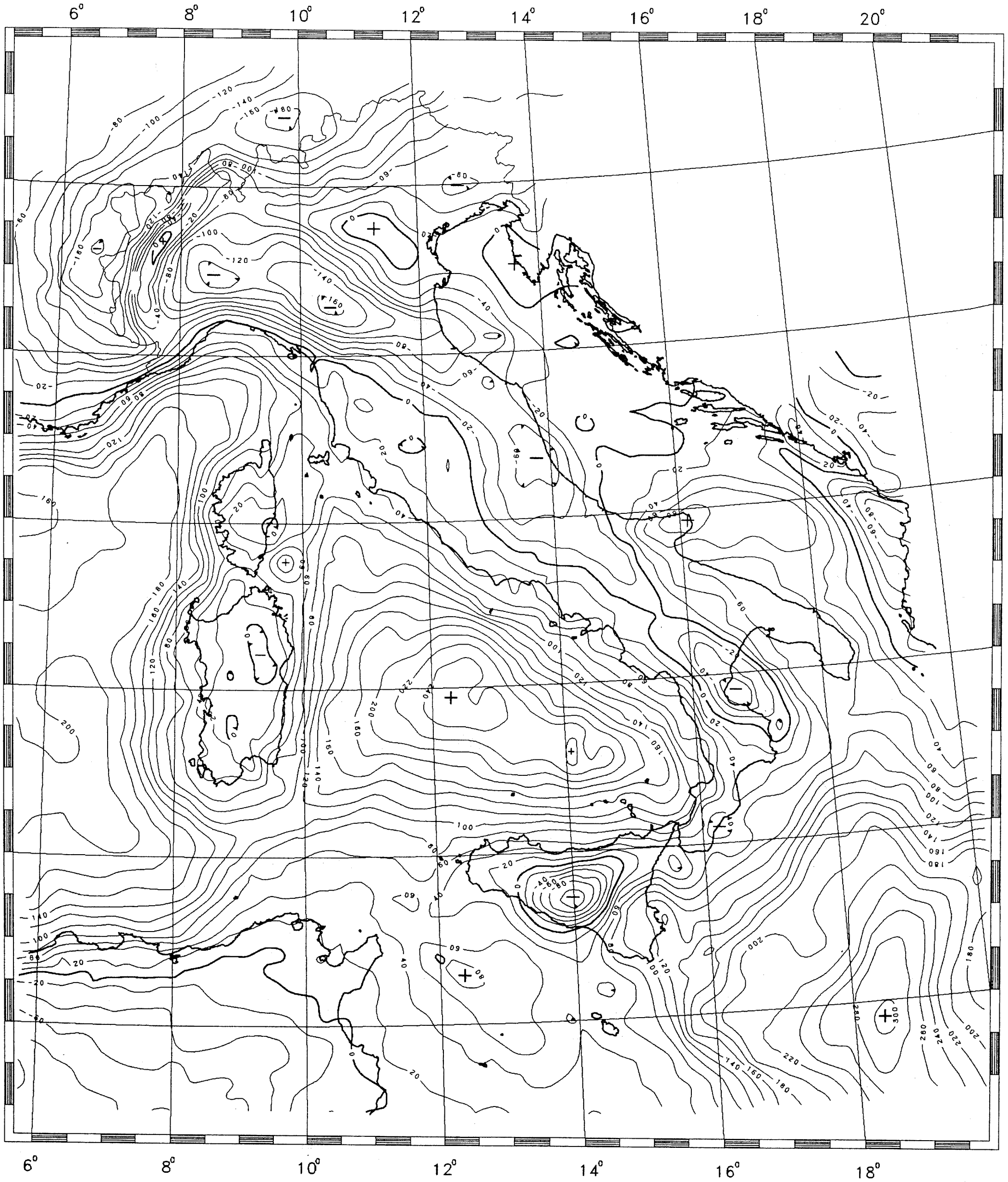

Fig. 3. Bouguer gravity anomalies (equidistance 20 mGals).

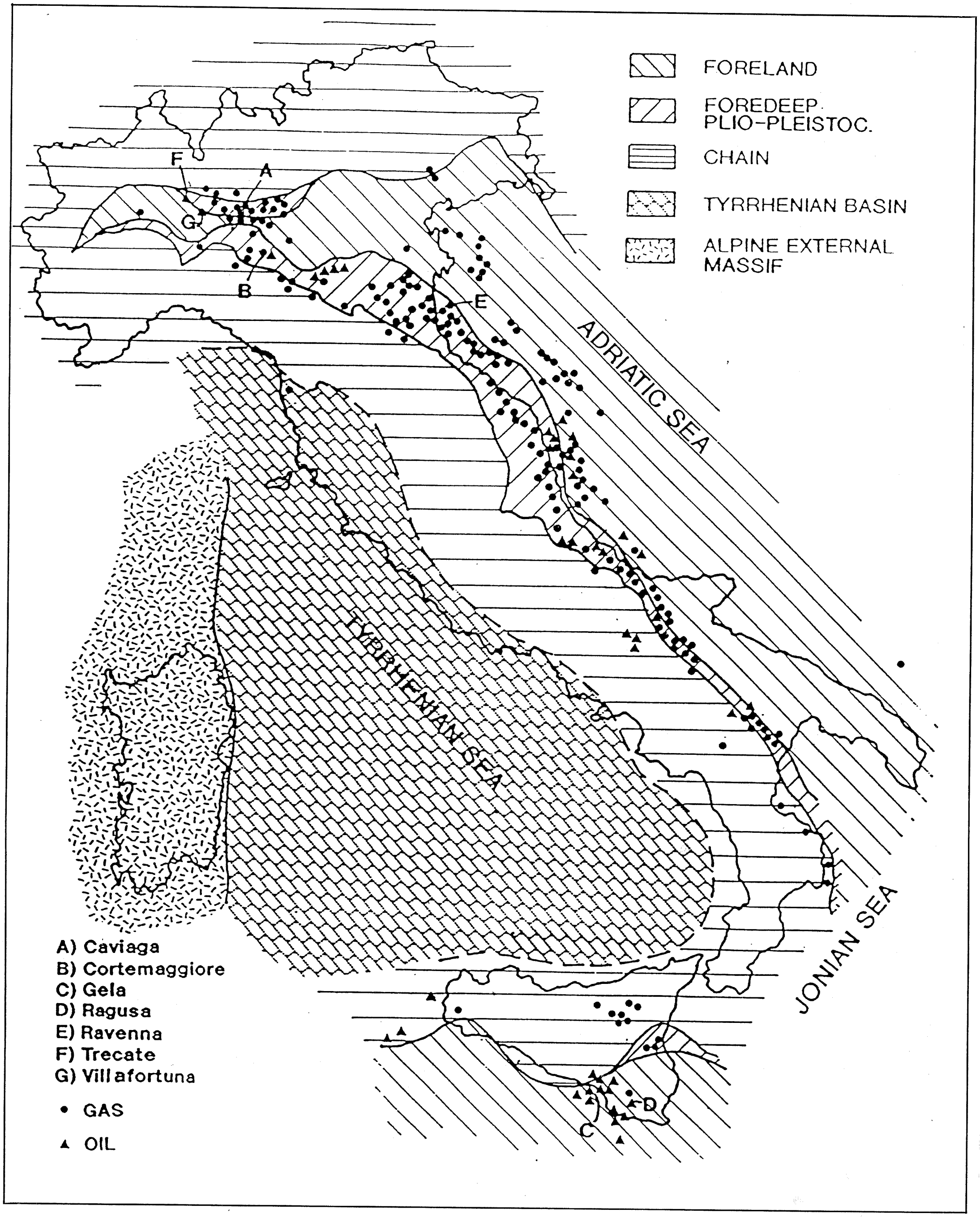

Fig. 4. Hydrocarbon findings in Italy: in the great majority in the peri-apenninic external fore-deep sedimentary basins (Bilgeri, 1991).

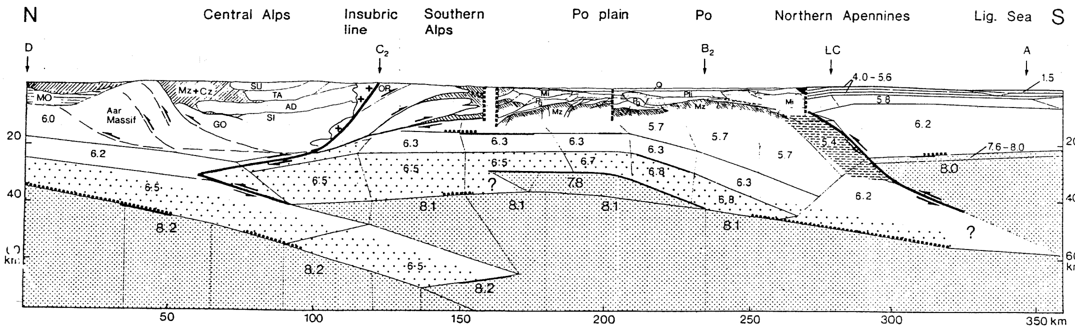

Fig. 5. EGT-S DSS section through the Central Alps and the Northern Apennines (Buness and Giese, 1990).

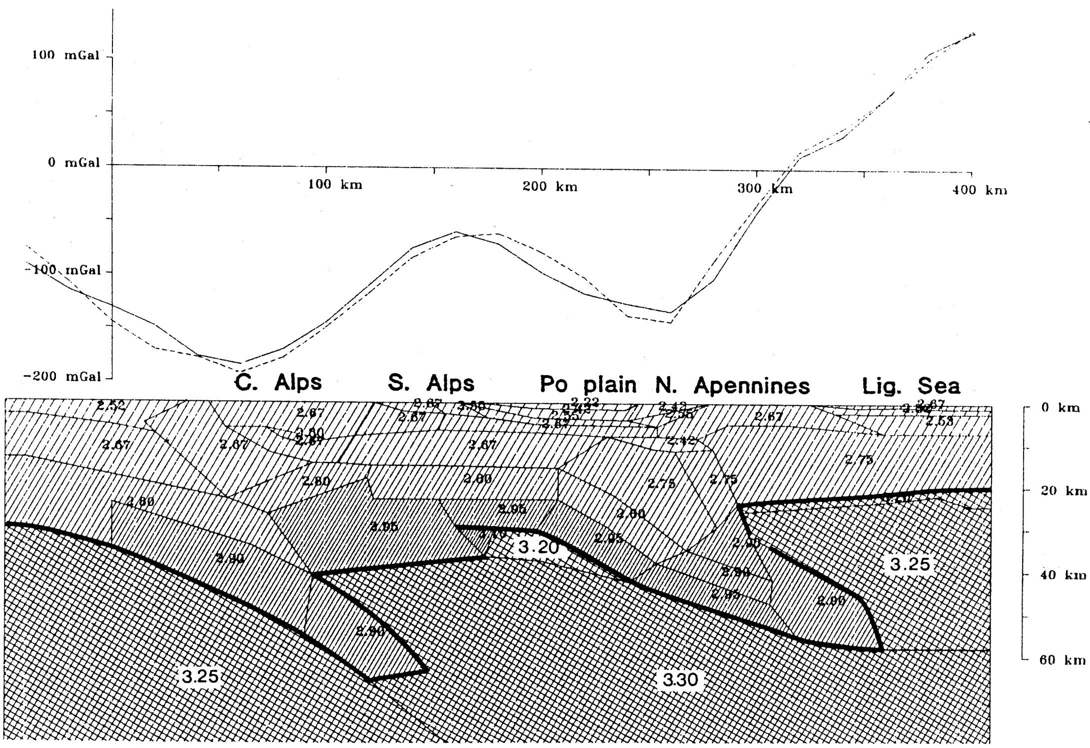

Fig. 6. Gravity model deduced from the seismic model of fig. 5 (Buness, 1992).

netic data is also badly defined when unsusceptive metamorphic formations are in place.

Within these limitations, the general features of the magnetic basement from the aereomagnetic surveys (fig. 7) indicate intense rifting beneath the Adriatic plate, elongated NW-SE along the peninsula, W-E in the Po plain. On the peninsula, the depth of the magnetic basement is mainly greater than 10 km and deeper than 14 km in the Pontremoli area with a subsequent rapid rise (by faults ?) SW-wards.

The Low Velocity Zones (LVZ) generally found in the upper and intermediate crust of the Alps, Northern Apennines and Po plain (fig. 11) can now be attributed to changes in the physical conditions, particularly microfracturation, porosity and permeability as explained in section 1.

3.2. *Lower crust*

It is well known that the main features of the lower crust are: ductility, flaking and silica depletion with respect to the upper crust. The increase in ductility is probably due to the temperature increase associated to the geothermal gradient, flaking is caused by the presence of magmatic materials whereas the increase in basic constituents is presumably derived from the vertical movements of basaltic bodies.

In addition, at the bottom of the crust the presence of trapped fluids at very high pressures can favour the formation of magmas, the transformation of the layers in contact with the latter and the formation of astenoliths. Another support for this mechanism of hydrothermal accretion of the lower crust (*i.e.*, undercrusting) comes from the horizons of serpentinite that have been found in the oceanic rifts.

In fact, in the most conspicuous mantle outcrops such as those that can be observed in some very eroded mountain ranges, the predominant rock type is peridotite which is made up of olivine in a particularly altered form called serpentinite. The olivine and in particular the serpentinite displays densities lower than those of the common mantle rocks and therefore it floats on the top of the mantle itself. The melting point of the olivine is rather high and it lags behind when the hot basaltic magma, owing to decompression, also decreases its density; therefore, it can migrate toward shallow levels and by so doing, separates from the upper mantle to originate volcanoes and produces massive fluid circulation by convection.

We can also see in the Earth a natural mechanism for the formation of two layers. The crust which is predominantly basaltic and which took billions of years to be formed, is the lighter component of the system and it forms the shallowmost layer. Because of its intermediate density, the olivine remains trapped between the crust and the mantle. The deep magmas pass through the olivine rich layer in their ascending movement.

The advancement in the understanding of the formation, presence and behaviour of fluids at depth has contributed new elements for the evaluation of the constituents of the tectonosphere and has also shed new light on the interaction between crust and mantle and the dynamic behaviour of the crust. In particular, we now have additional information on the formation of faults, the role of fluids in the lubrication and the displacement of large masses, especially in the orogenetic and seismically active areas.

The presence of fluids in the lower crust is the key factor to understanding the internal mutations and, especially, to explaining the reduced coupling (*i.e.*, detachment) between upper and lower crust at the origin of the ongoing large overthrusting and gravitational sliding phenomena.

3.3. *Moho discontinuity*

The Moho discontinuity is now viewed more as a transition zone between the crust and the upper mantle than as a sharp contact between two distinct lithologies (fig. 11). In fact, it has been found that the Moho is often represented by some smooth change in seismic velocity. In this latter case, the Moho depth is assumed to correspond to that of the largest velocity gradient.

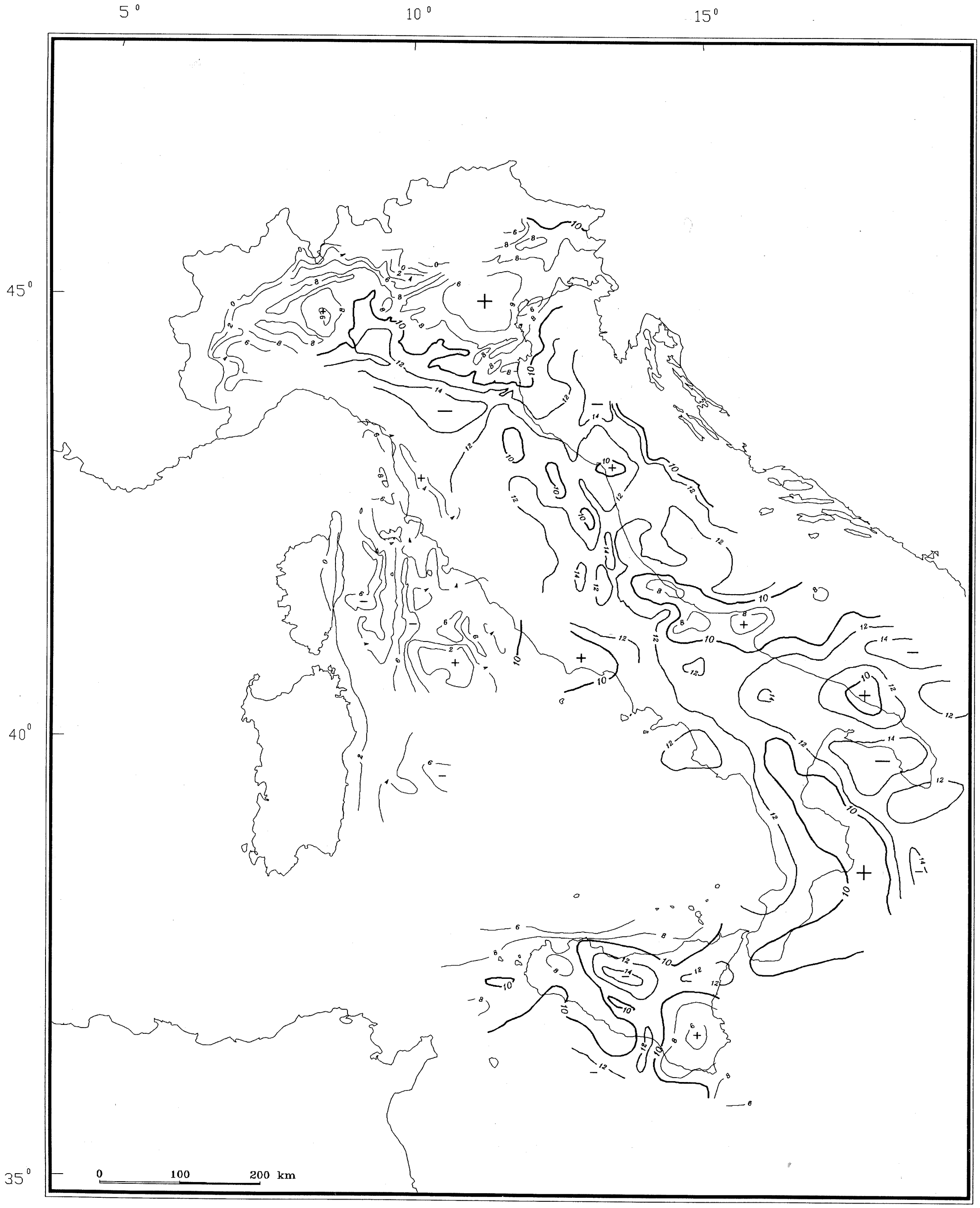

Fig. 7. Isobaths (in km) of the magnetic basement (Cassano *et al.*, 1986).

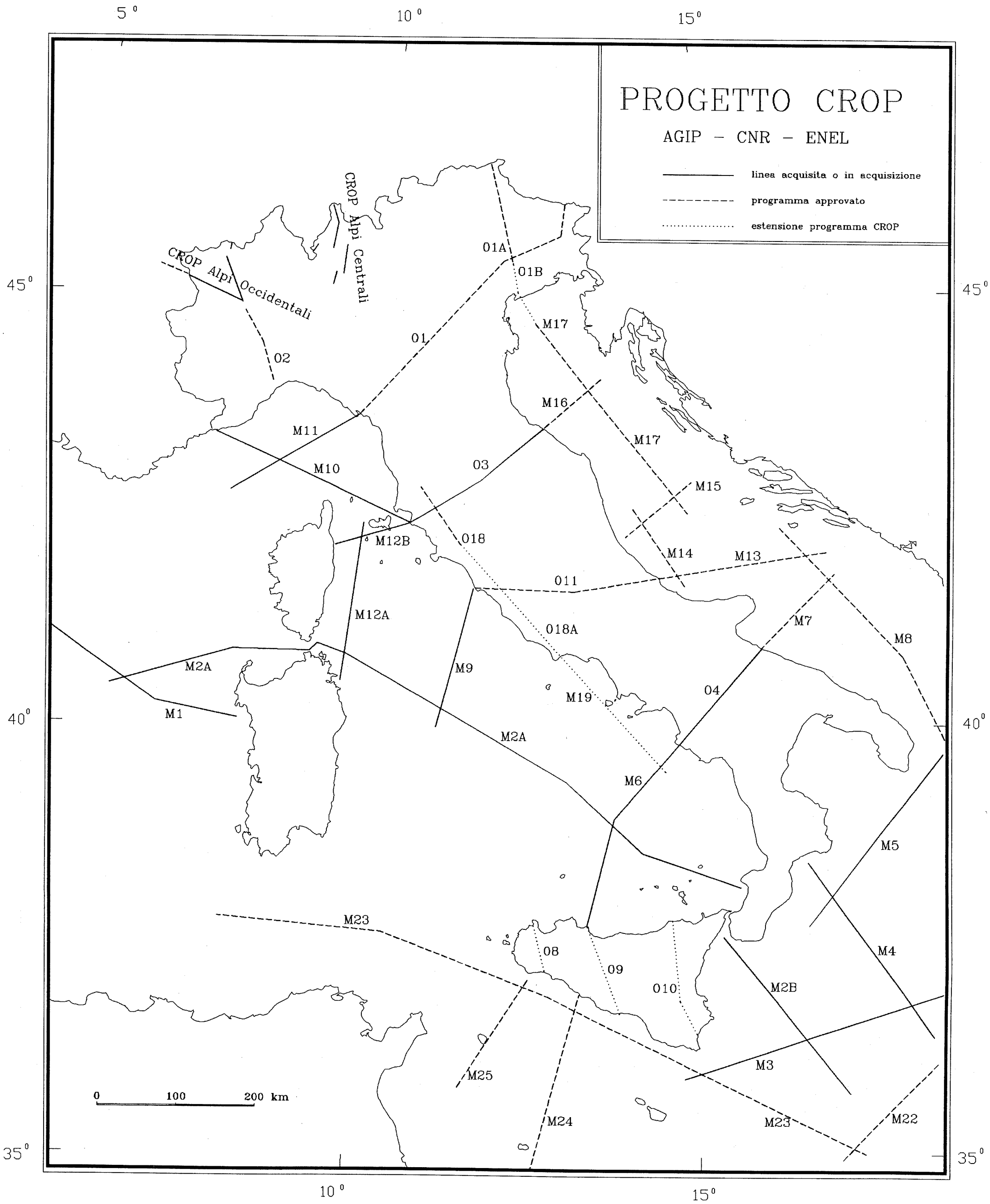

Fig. 8. The deep reflection seismic project CROP in Italy.

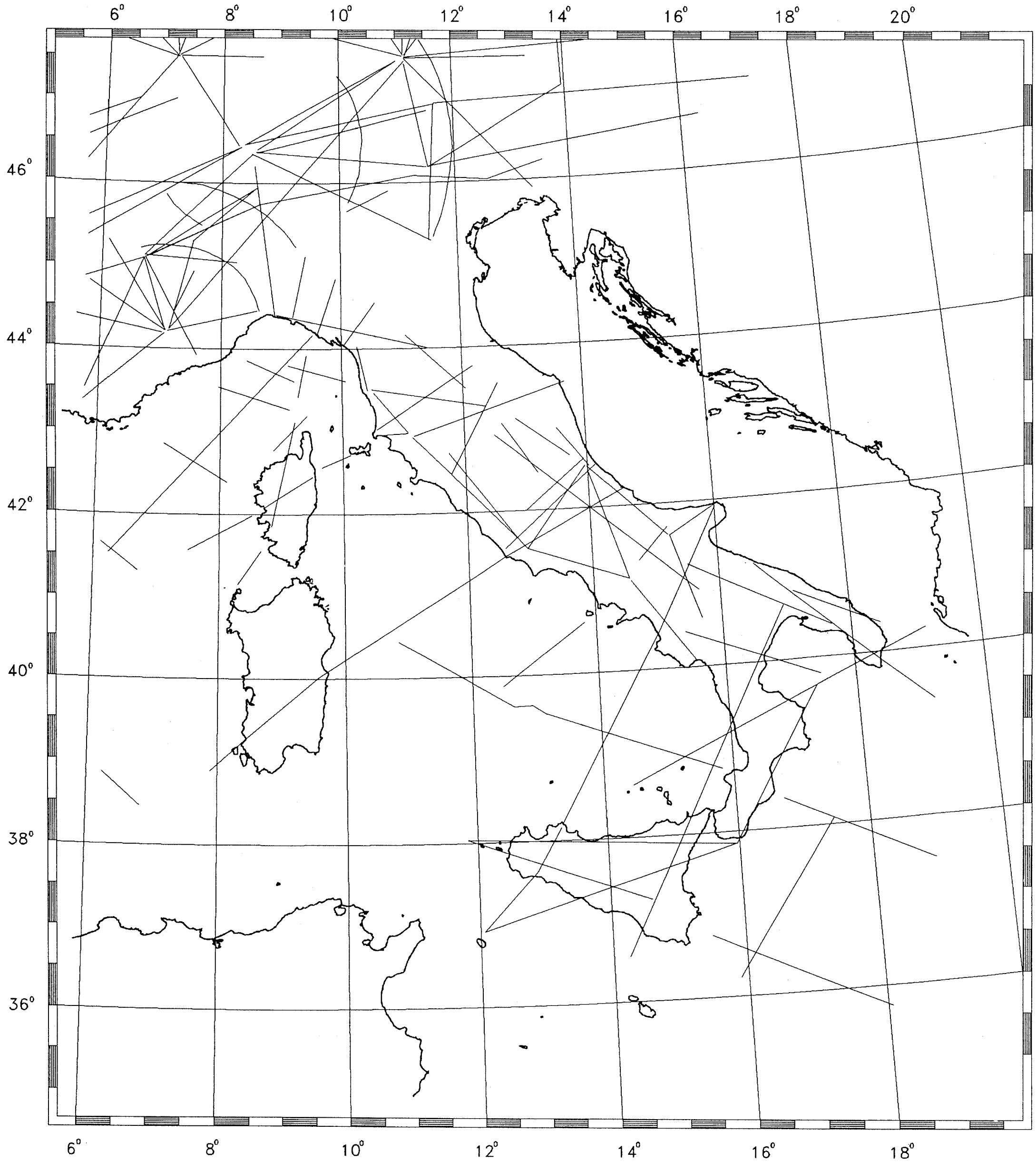

Fig. 9. DSS seismic profiles in Italy 1956-1981 (CNR, Explosion Seismology Group).

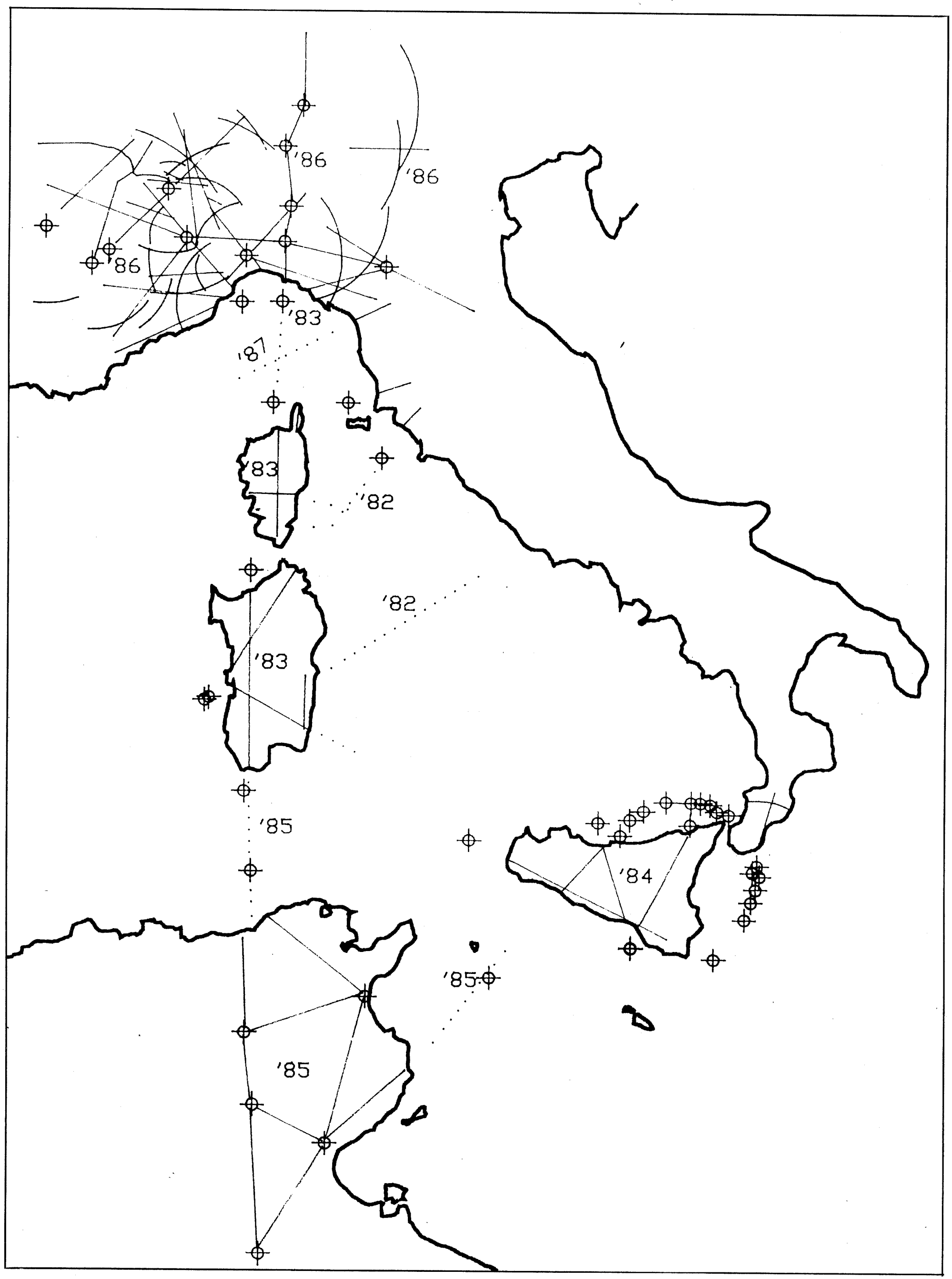

Fig. 10. DSS seismic profiles 1982-1986 for the European Geotraverse, Southern Segment (EGT-S).

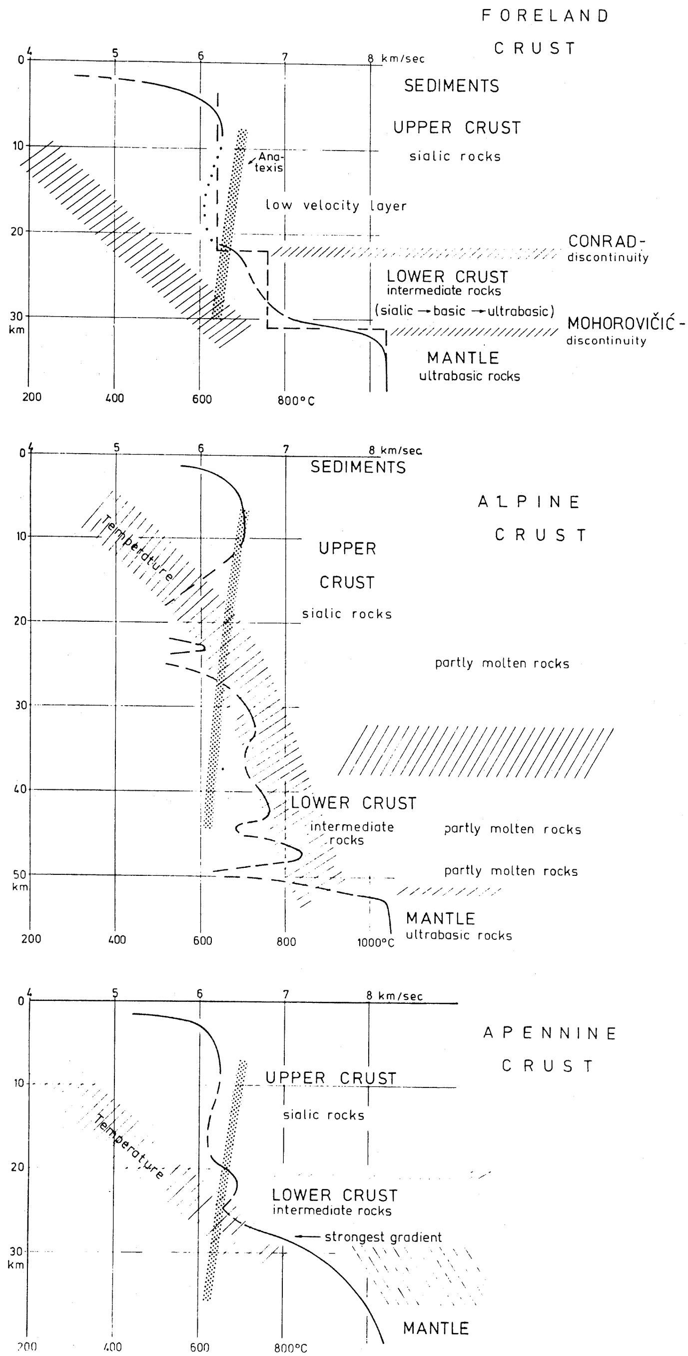

Fig. 11. Crust typologies for the northern fore-deep, Central Alps and Northern Apennines (Giese and Morelli, 1975).

The Moho map is displayed in fig. 12. In the interpretation of this map, it should be kept in mind where most of the profiles are located. For example, we have a good resolution of the Moho depth in places like Piedmont and Liguria where there is a high density of profiles whereas there is a poorer resolution in the central and oriental Po plain. Knowledge of the Moho depth has recently been extended in areas covered by the EGT such as Sardinia and the Sardinia channel. The results collected so far would indicate that three types of Moho extend within the Italian peninsula.

European Moho (thin contour lines) – The African plate collided against the European foreland in its counterclockwise rotation associated with the opening of the Atlantic. In the collision, parts of both European and African crust and slivers of mantle have accumulated above the European Moho. This scenario can be appreciated in fig. 5 where it is shown how the European ramp (*i.e.*, the lower European crust) was bent by the load increase of the northern margin of the Adriatic plate.

The same arguments apply to the Western Alps having westerly oriented vergence.

The maximum depth of the European Moho is approximately 70 km which is close to the maximum values reached by Tertiary orogenetic zones. This can be explained by the overall subsidence and the associated increase in pressure and temperature with depth. It follows that the basaltic lower crust at such conditions of pressure and temperature will undertake a phase transformation into eclogite. Because the seismic velocity of eclogites does not differ from that of peridotites, it appears generally quite difficult to distinguish between these two types of assemblages using geophysical methods. However, it should be remarked that the density of eclogites is consistently larger than that of peridotites and this is the reason why subsidence is generally favoured.

It also appears that the Corsican-Sardinian block is a sliver of the European continent which was once attached to Provençe. It is structurally well defined and it displays a thinned and widened western margin.

Adriatic (African) Moho – The depth of the Adriatic Moho is longitudinally nearly constant (~ 30 km) along the azimuth NNW-SSE throughout the peninsula and the Adriatic Sea. A slight deepening of the Moho occurs along the Adriatic coast (~ 35 km). The Adriatic Moho, however, undergoes some considerable bending beneath the Apennines and towards the Tyrrhenian coast it reaches a depth of nearly 50 km presumably along faults or systems of faults. This occurs at the crustal doubling which was found along the northern and western margins of the Adraitic plate (Wigger, 1984).

The Adriatic plate can be considered a back-land only along its NNW margin where it rests on the European plate. Conversely, its southwestern part should be considered a foreland. Here, its has been covered by a thin crust of Tyrrhenian origin which was driven tectonically toward NNE, E and SE in the Tertiary. These movements are also responsible for the thick sedimentary covers that characterize the Apennines.

The minimum depth of the Adriatic Moho at a regional length-scale is about 30 km beneath the central and western Po plain. This location also corresponds to the maximum mantle upwelling and the origin of the Milan positive gravity anomaly (see fig. 3). Along the flanks of this oblong W-E anomaly, the contours of the Adriatic Moho deepen toward the north (~ 55 km, European crust) and the SW (45 km, Ligurian crust) whereas they become more shallow (~ 10 km) toward the NW (*i.e.*, the Ivrea anomaly).

Peri-Tyrrhenian Moho (oceanic or thinned continental crust) – We have grouped together various types of crust which are to some extent similar but that almost certainly differ in terms of origin.

– The 10 km Moho depth-contour in the Tyrrhenian Sea delimits crust which has certain by oceanic origins. The crust is likely to be oceanic within the 15 km contour in the Provençal sea.

– The crust is continental as it increases its

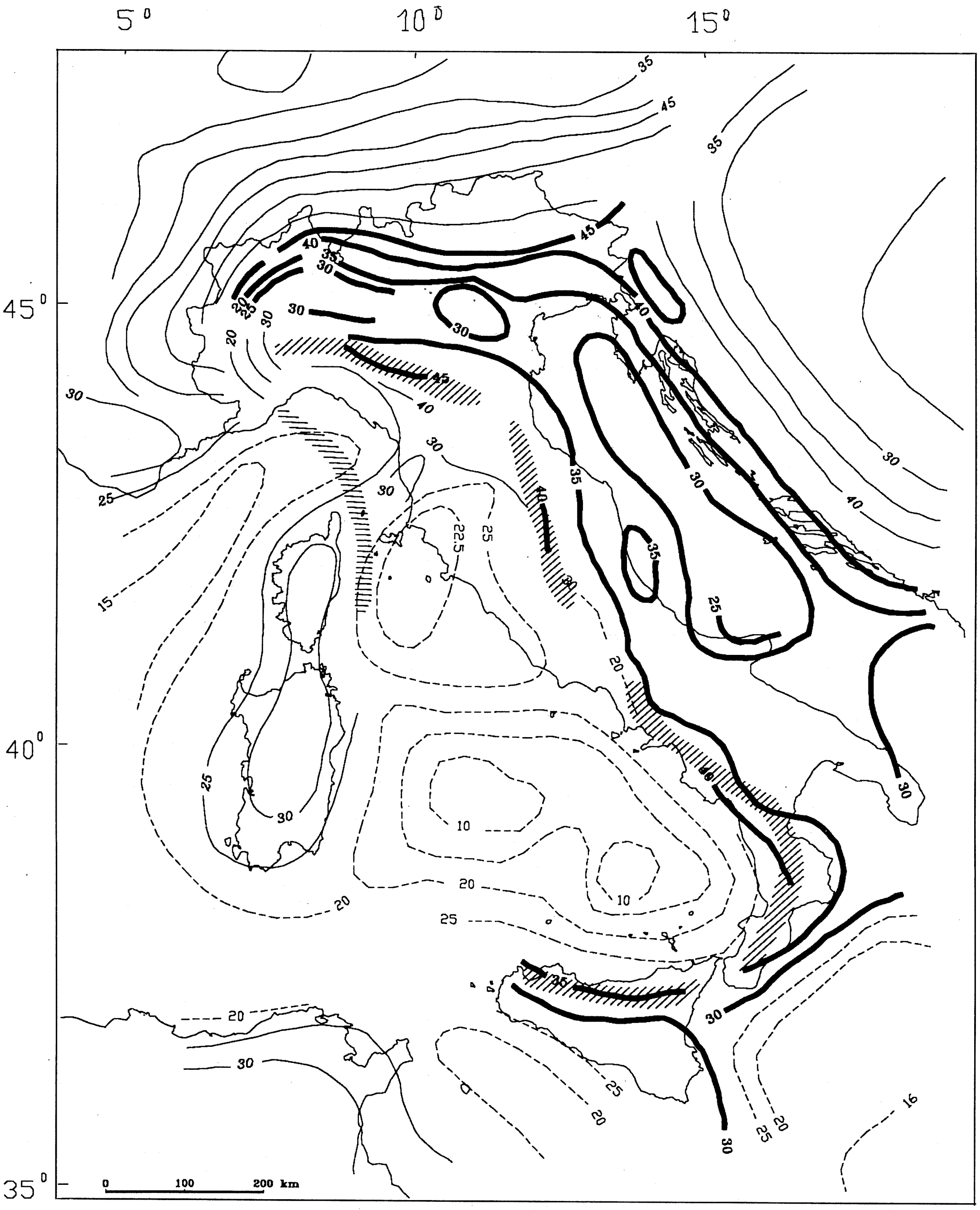

Fig. 12. Moho isobaths in Italy (Nicolich and Dal Piaz,1991; modified). Equidistance: 5 km; thick = Adriatic; thin = European; dashed = oceanic or thinned continental.

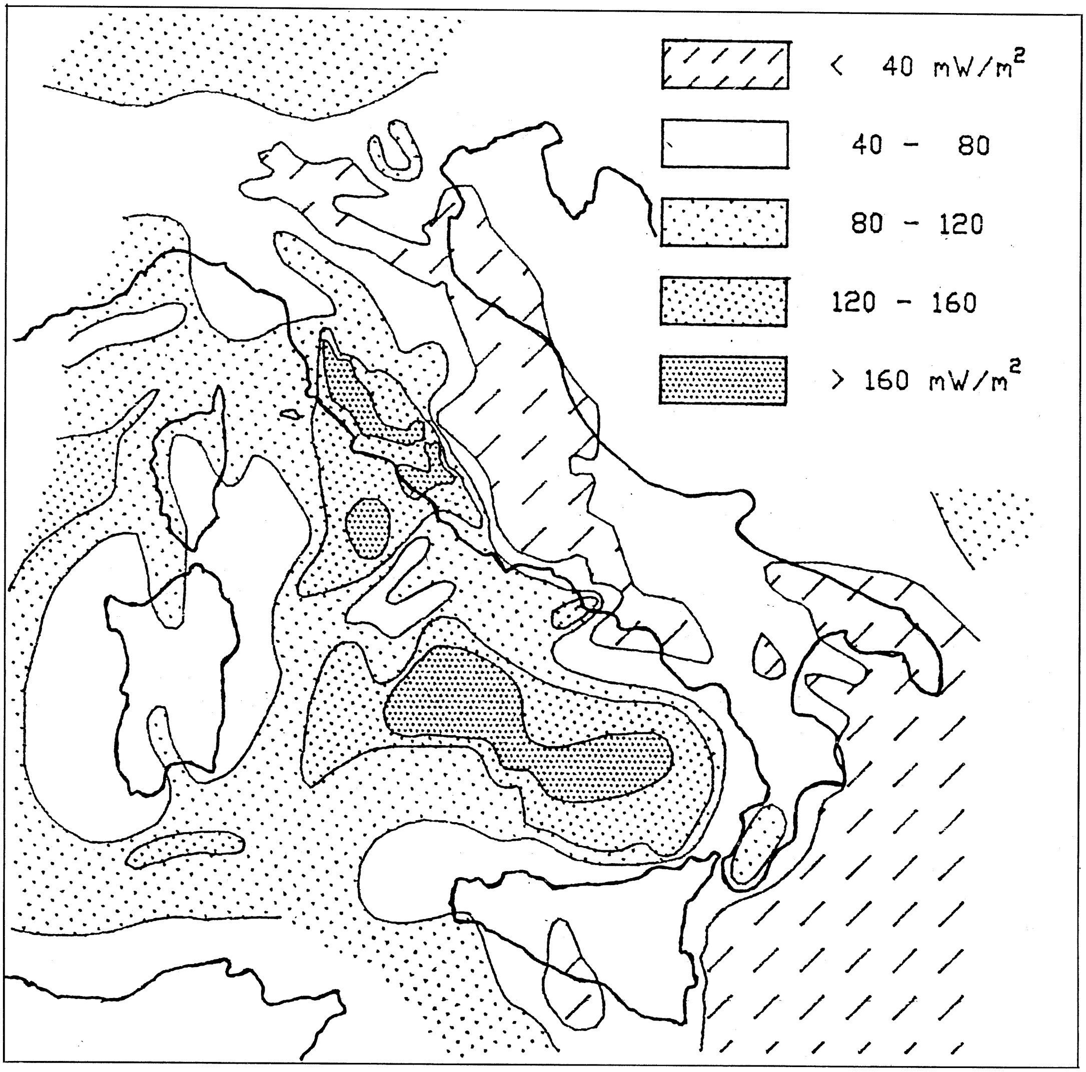

Fig. 13. The heat flow in Italy, in mW/m^2 (Mongelli *et al.*, 1992).

thickness up to ~ 25 km. It is in an opening-ridge stage in the Sicily strait.

The principal novelty consists in the land extension of this type of crust along the entire Apenninic arc either toward NE (Ligurian) or toward E (Tuscan), SE (Calabrian) and S (Sicilian) where the crust has been overthrust on the margin of the Adriatic ramp (which subsided). There is a striking similarity between this setting and the analogous one that can be found in the Alps although the time of occurrence is considerably more recent.

It is now certain that the Apennine orogeny ended in the Tortonian along the northern part of the Apenninic chain, in the Messinian along the central part of the chain, in the Pliocene along the southern part and during the Pleistocene in the Calabrian Arc. Most of the Apenninic edifice and its foredeep result from the Neogenic and Quaternary uplifts within the Tyrrhenian area. The progressive uplift of the tectonic block toward the south has been interpreted by Locardi (1988) as the subduction beneath the crust of fluid-like material of mantle provenance.

The Peri-Apenninic foredeep formed in the Pliocene when the spreading rate of the Tyrrhenian was maximum.

The Apulian basin took part in the Apennine orogeny during the Lower Pliocene. It has been found that the sedimentary covers from this basin have been folded and overthrust upon the Lower Pliocene whereas, near their frontal part, they sometimes rest upon even younger sediments.

The presence and the extension of the Tuscan-Tyrrhenian asthenoliths has also been confirmed by the anomalies in heat flow (fig. 13) that also match remarkably well the contours of thinned crust previously pointed out. This relationship (cause-effect) is of fundamental importance in understanding the geodynamic evolution of the Italian peninsula.

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