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Active seismic layers and crustal structure in some Italian regions

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Abstract. Following a general comparison between the hypocentres and crustal structure derived mainly from deep seismic soundings, a more detailed analysis is attempted in some areas of the Italian region. This study is undertaken by defining a region as homogeneous when the same type of velocity-depth functions has been obtained. All available carthquake hypocentres within the area are used for the correlation with the V_p distribution in order to obtain a sufficiently large number of data. Where available, low magnitude events and aftershock sequences have also been used. The conclusions already derived (Cassinis et al., 1984) are more clearly confirmed. In the south-eastern Alps (Friuli), the main earthquake sources are in the upper crust, between 5 and 15 km, at the top of the first velocity inversion zone. In Tuscany (thin continental, transitional crust), the seismicity is weak and very shallow. In the Calabrian arc, the main seismicity seems to originate in the lower crust or at the top of the lithospheric "lid" which appears to be decoupled from the overlying, highly mobilized formations of the crust. Tentative explanations of these different regimes are given with regard to the influence of factors that can determine the transition from brittle to quasi-plastic behaviour.

Key words: Seismogenesis – Brittle-plastic transition – Velocity functions – Inversion zones

Introduction

In a previous paper (Cassinis et al., 1984) the available hypocentres of earthquakes with $M > 3.5$ were correlated with individual functions of V_p obtained using wide angle reflection seismic profiles (deep seismic soundings = DSS).

The correlation is made with respect to hypocentres contained in a vertical cylinder with a radius of 40 km around the average position of the calculated velocity function. The seismic activity is described by histograms of the distribution with depth of the number of foci, the maximum magnitude and the released energy during the historical period (1000–1981 A.D.).

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There are strong limitations and constraints in this general analysis due to the inequality of the data. The number of available hypocentres is often too small to allow reliable statistics and the accuracy of the focal depth is very poor for the old data (before 1970). Also, the interpretation of the velocity functions is sometimes debatable. However, the results led to some preliminary considerations, namely (Fig. 1):

- In north and central Italy the most active seismic layer seems to be located in the upper crust, near the top of the first zone of velocity inversion. The seismic activity in the middle-lower crust is very low everywhere or, at least, high magnitude earthquakes do not originate there.
- In the south, especially along the Calabrian arc, the main activity seems to originate at the "*M*" discontinuity, while the low seismicity in the upper crust could correspond to a low strength materiel and, possibly, a decoupling could exist between the crust and upper mantle.
- In other words, in north and central Italy the upper crust seems more brittle than in the southern section. Here, in turn, the upper mantle appears to be stronger or, at least, it seems that stresses are not transmitted to the upper crust.
- Another point is that, considering the interpretation of the crustal structure according to the DSS results (Fig. 2), the main seismicity in northern and central Italy takes place on the lifted edge of the foreland crust (Adriatic), while in south Italy and in Sicily it occurs along the lifted edge of the hinterland crust (South Tyrrhenian) (see also Cassinis, 1983; Cassinis et al., 1979; Giese and Morelli, 1975).

In view of these preliminary results (Cassinis et al., 1984), in this study we have considered the need for more detailed surveys in particular areas where adequate data are available. The models of the crust, designed according to the behaviour of the seismic velocity V_p, have been improved by the introduction of hypotheses on other parameters (temperature and type of material). On the other hand, aftershock series and microseismicity have also been used, when available, for the correlation with the velocity functions in order to increase the number of foci. The behaviour of the main shocks has been compared to that of the microseisms and aftershocks.

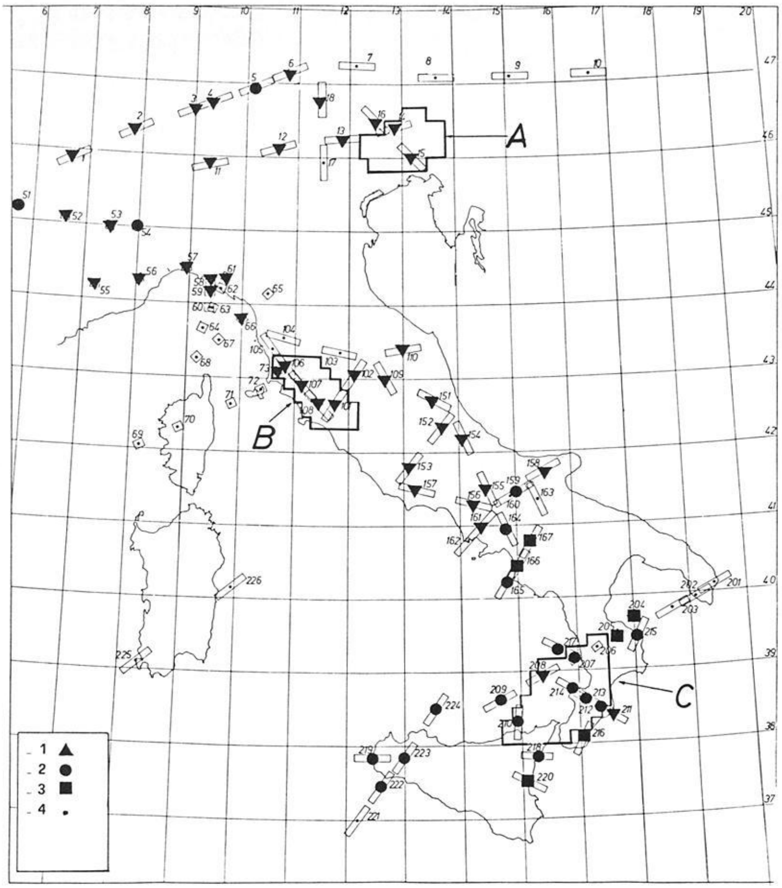

Fig. 1. Position map of velocity functions derived from seismic crustal profiles. The small *dots* represent the position of the half-critical distance for the "M" surface; the *rectangles* (oriented according to the strike of the seismic profile) indicate the linear range of velocity distribution; the *squares* correspond to a vertical distribution. Comparison between seismic activity and crustal structures. The most active seismic layers are located: 1) at the top of the first zone of velocity inversion (upper crust); 2) near the "M" discontinuity; 3) near the upper transition to the "M" (in the case of "double" transition); 4) lack of information on seismicity (Cassinis et al., 1984). The three areas examined in detail in the present study are indicated: A) Friuli; B) south Tuscany; C) Calabrian arc

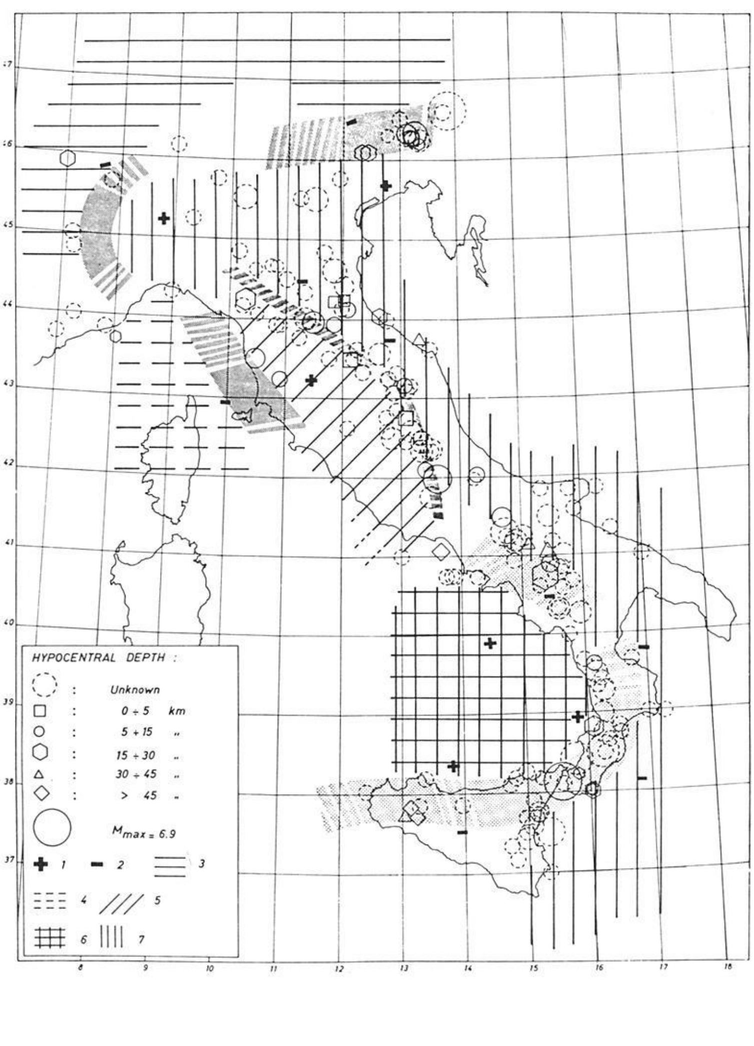

Fig. 2. The major lateral discontinuities and crustal domains compared with the map of strong earthquakes from 1000 to 1975 A.D. ($M > 5.5$). 1) and 2) Lifted or depressed "M". Domains: 3) European-Alpine; 4) Corsican-Ligurian; 5) "Tuscan" intermediate crust; 6) South Tyrrhenian; 7) Adriatic-African. The *broad bands* indicate the zones of "double" "M" transition; the *narrow strip* corresponds to a "flexure" of the "M" (Cassinis et al., 1984)

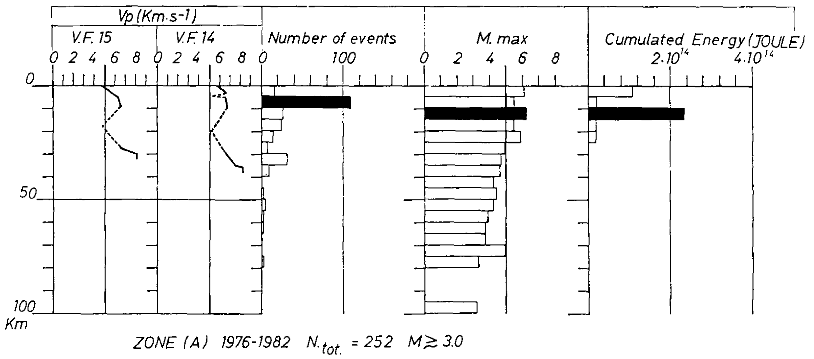

Fig. 3. Velocity functions of zone A (Friuli) (Italian Group for explosion seismology) compared with the seismic activity during the period 1976–1982 (hypocentres with $M>3.0$ recorded by the Istituto Nazionale di Geofisica), 252 events. From left to right: a) velocity functions; b) number of events in steps of 5 km; c) highest magnitude recorded in each step; d) cumulated energy in each step (Joule), linear scale. *Black bars:* depth interval with maximum values. Energy values calculated after the Bäth formula

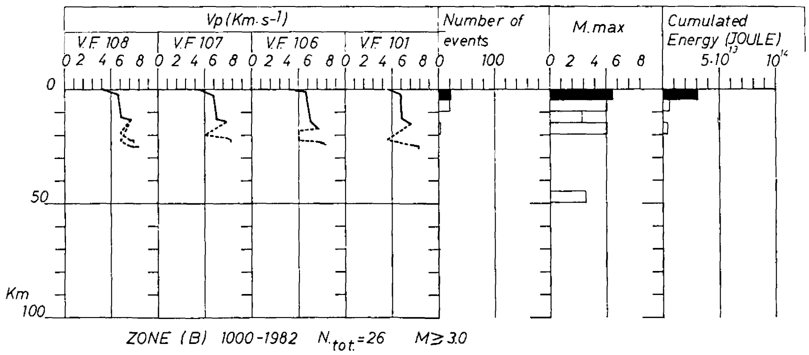

Fig. 4. Same as in Fig. 2 for zone B (south Tuscany). Period of observation: 1000–1982 A.D. Total number of events: 26

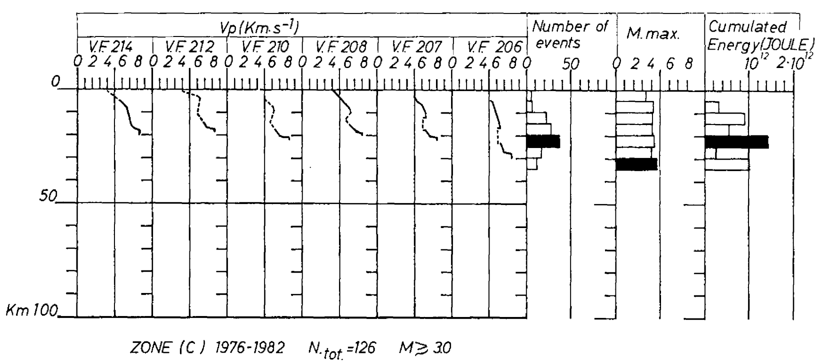

Fig. 5. Same as in Fig. 2 for zone C (Calabrian arc). Period of observation: 1976–1982. Total number of events: 126

Investigation in particular areas

Three areas have been selected for more detailed analysis (Fig. 1): Friuli (A) on the margin of the southern Alps; central and southern Tuscany (B); the Calabrian arc (C). The velocity functions as well as the foci falling within the area are used. For the analysis. Each area is defined as a homogeneous structural unit, when the velocity functions within its boundaries show a similar behaviour.

In Fig. 3, two velocity functions and histograms of the observed seismic activity during the period 1976–1982 (252 events) are plotted for zone A (Friuli). The depth interval is 5 km. The hypocentres considered here ($M \geqq 3$) do not include aftershock sequences or small magnitude events. The correlation seems clear; the number of events shows a maximum between 5 and 10 km while the M_{max} and the cumulated energy reach the highest value one step below, at the top of the velocity inversion. The seismic activity is very weak in the middle and lower crust. The shift of 5 km between the first and the other two maxima can be explained by the high number of small energy events recorded in the period (Siro and Slejko, 1982).

In Fig. 4, four velocity functions in central-southern Tuscany (zone B, a geothermal area) are plotted. All data indicate a thinned continental crust (total thickness about 22 km) with a low velocity gradient in the upper crust and a velocity inversion in the lower crust. Unfortunately, local seismographic networks do not operate there, so only the catalogue and bulletins of the low density national observatories are available. The number of plotted events is very small. However, the focal depths of the observed events agree with the macroseismic data of the area, indicating very shallow earthquakes (ENEL catalogue, unpublished). The most active layer seems to be located within the uppermost 5 km.

Finally, Fig. 5 illustrates six velocity functions of zone C (Calabrian arc); all functions show a large thickness of low-velocity material overlying a zone of strong velocity gradient. The velocity inversion seems

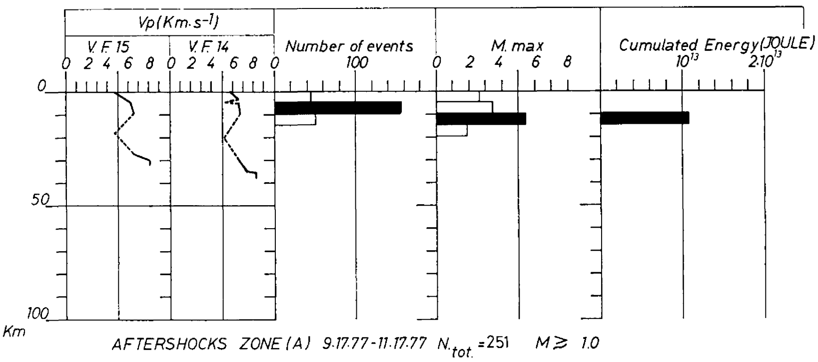

Fig. 6. Histograms of an aftershock sequence in zone A, following a main shock of $M=5.4$ (251 events). Same symbols as in Fig. 3 and the following figures. Compare with Fig. 3

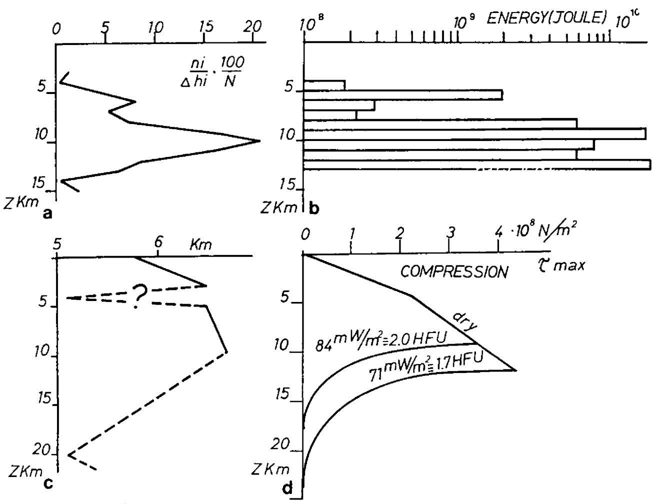

Fig. 7a–d. Aftershock sequence of 16 September–17 November, 1977 (same as Fig. 6). **a** normalized frequency-depth curve (depth interval: 1 km); **b** cumulated energy in each interval (logarithmic scale); **c** velocity function 14 (upper part – Fig. 6); **d** maximum stress curves (after Meissner and Strehlau, 1982). Dry quartz rheology, compression mechanism

to be located in the lower crust and is generally weak. The transition to the Moho is fast and is found at depths ranging between 18 and 25 km. The main seismicity seems to be concentrated between 20 and 25 km, i.e. at the "M" transition zone or in the uppermost mantle. The foci with depths >30 km (subduction slab) were not considered.

Study of Aftershocks in zone A (Friuli)

An aftershock sequence containing 251 events, recorded by the local seismographic network (O.G.S., Trieste) (Suhadolc, 1982) during the 60 days following a moderate event ($M=5.4$) that occurred on 16 September 1977 at a depth of 14 km (±2 km), was analysed. The sequence took place in the same area of the destructive shock of 16 May 1976 with $M_s=6.4$–6.5 at an estimated depth of about 10 km. The minimum recorded magnitude is about 1.0. Only those events with an estimated accuracy of local depth better than ±3 km were used for the analysis.

In Fig. 6 the correlation between the velocity functions and the seismic activity is shown again, using the aftershock sequence. The peaks of the histograms correspond exactly to those of Fig. 3 (1976–1982).

The number of hypocentres has also been represented (Fig. 7) with a normalized frequency-depth curve ("defreq" diagram, Meissner and Strehlau, 1982) using depth intervals of 1 km. The peak of the curve is found at a depth of 10 km, corresponding to the top of the inversion zone.

The depth range of the aftershocks is between about 14 (main shock) and 4 km. According to Shebalin (1976) and other authors, the thickness of the active layer increases very slightly with magnitude. A magnitude increase from 6.0 to 6.5 corresponds to an increase in thickness of about 2 km (from 10 to 12 km).

Models of the continental crust and the seismic regime

The tentative models of the crustal structure in Italy are based mainly on the velocity distribution of P waves as interpreted from DSS profiles. Other complementary data, such as gravity or magnetic anomalies, are considered when significant. Unfortunately, one key factor, the heat flow value, is generally not available, especially inland, except in central and southern Tuscany.

For the following considerations, it is assumed, in a general sense that, in the continental crust, the upper part of the velocity functions (gradient decreasing with depth) corresponds to the upper crust (the top of the "granitic" layer corresponding to $V_p=5.8$–6.0 km/s). In general, the first velocity inversion is considered as marking the transition from the upper to a "middle" crust. The last part of the curve, where an increasing velocity gradient is generally observed, corresponds to the lower crust and the transition to the upper mantle, where V_p reaches about 8 km/s.

The Italian region is a transitional area where the structure and, consequently, the velocity distribution with depth are subjected to strong lateral variations. Therefore, the crustal type is often very different from the model proposed for the "continental" crust (see, for instance, Mueller, 1977).

In all the proposed models, the upper and middle crust is considered as "granitic" or having a behaviour similar to quartzitic rocks. On the basis of this assumption, and taking experimental and theoretical results into account, several authors (Byerlee, 1968; Brace and Kohlstedt, 1980; Sibson, 1982; Meissner and Strehlau, 1982) have calculated the limits of stresses in the continental crust. The variables are temperature and pressure; their increase determines a transition from a frictional regime to a quasi-plastic flow that induces stresses in the rigid formation above.

Crustal strength-depth curves are calculated for different quartz rheologies and considering both dry and wet conditions (no pore pressure or hydrostatic pore pressure). Also, the different types of faulting (compression, extension or strike slip) are taken into consideration. According to these model parameters, the peak of the maximum stress marks the transition between the brittle and ductile behaviour. The peak range is quite wide: in the case of a wet crust, heat flow $=2$ HFU and extension faulting, the peak is at 5 km and $\tau_{max}=10^7$ N/m^2. For a dry crust, heat flow $=1.2$ HFU and compression, the peak is at 32 km and $\tau_{max}=9\times10^8$ N/m^2 (see for example Meissner and Strehlau, 1982).

Let us apply these "stress max" curves to the model of the upper-middle crust in zone A (Fig. 7).

Considering the independent information on the focal mechanism of the earthquakes in this area, we can assume a compressive stress. It is more difficult to choose the other parameters involved (wet or dry rheology, creep rate and heat flow).

Only on the basis of the depth of the peak of the depth-frequency diagram (Fig. 7a) and of its value (consistent with the largest known earthquakes in the area) do we assume a dry quartz rheology and a creep rate $\dot{e}=10^{-17}\,s^{-1}$.

As far as the heat flow is concerned, and also considering the following section, we estimate a range between 1.7 and 2.0 HFU (corresponding to 71 and 84 mW/m^2, respectively) as reasonable. With these assumptions, the peak of τ_{max} ranges between about 3.8×10^8 N/m^2 at $Z=10$ km and 4.5×10^8 N/m^2 at $Z=14$ km (Fig. 7d).

To make only a rough estimate, we can transform these two values of the peak of τ_{max} into the maximum distortional strain energy per unit volume, E_d (Sibson, 1982):

$$E_d=\frac{|(\sigma_1-\sigma_3)^2\cdot(1-K+K^2)|}{6\mu}, \tag{1}$$

where

$$0<K=\frac{(\sigma_2-\sigma_3)}{(\sigma_1-\sigma_3)}<1.$$

If we take

$$\sigma_2=\frac{\sigma_1+\sigma_3}{2},$$

then $E_d=\dfrac{\Delta\sigma^2}{8\mu}$, where μ is the rigidity.

According to the Mohr formula (see Terzaghi and Peck, 1948)

$$\tau_{max}=\frac{\sigma_1-\sigma_3}{2}=\frac{\Delta\sigma}{2},$$

we obtain

$$E_d=\frac{\tau_{max}^2}{2\mu}. \tag{2}$$

Assuming an average density for granitic rocks of $\rho=2.7$ gr/cm^3 and $V_p=6.2$ km/s, we have the rigidity $\mu=3.45\times10^{11}$ dyne/cm^2. Substituting these values in Eq. (2) we find, for the two peaks:

$E_{d_1}=2.93\times10^6$ Joule/m^3 for $\tau_{max}=4.5\times10^8$ N/m^2

and

$E_{d_2}=2.09\times10^6$ Joule/m^3 for $\tau_{max}=3.8\times10^8$ N/m^2.

In order to estimate the radiated energy, we have to account for the seismic efficiency η. Assuming a coefficient η of 0.1 (Lomnitz, 1974) and taking the average value between E_{d_2} and E_{d_2} we arrive at $E_{d\,rad}=2.51\times10^5$ Joule/m^3. This value seems to be compatible with the size of the largest earthquake recorded in historical times (16 May 1976). Taking a focal volume for this earthquake:

$V_f=A\cdot\bar{u}=400$ km$^2\times33$ cm $=1.32\times10^{14}$ cm^3

(Cipar, 1980) we calculate a total energy of:

$E_{tot}=3.31\times10^{13}$ Joule $=3.31\times10^{20}$ erg,

a value which corresponds fairly well to the published values (Cipar, 1980).

In conclusion, the choice of the "stress max" curve seems well supported by three independent sets of data, namely: the depth of the bottom of the active layer, the energy produced by the largest known earthquakes and the depth of the velocity inversion. Of course, the above considerations do not allow us to forecast the M_{max} to be expected in the area; however, they contribute to the definition of the active layer and of the most probable depth of the nucleation of the strongest earthquakes. According to both the V_p functions and to the hypocentres, this depth should be between about 10 and 14 km.

Heat flow and velocity inversion in zone A

The selection of a heat flow ranging between 1.7 and 2.0 HFU (71 and 84 mW/m^2) is supported by the velocity-depth functions. The interpretation of velocity inversions based on the DSS data contains many ambiguities. However, the top of the first inversion is generally fairly well determined by the termination of the P_g travel-time curve, while the thickness and velocity distribution within the inversion zone are both unknown. Christensen (1979) has calculated the critical thermal gradient (i.e. the thermal gradient that balances the increase of velocity due to the increase of pressure) for various types of rocks.

If we consider a homogeneous granitic upper crust, we see from Christensen's values that at 10 km depth a very high heat flow is required to produce a velocity inversion, unless other factors such as pore pressure or change of material properties are involved. Thus, it seems reasonable to consider a heat flow higher than the average in order to explain, at least partially, the low-velocity zone. Also, geological considerations support these high values, the area being located on the margin of the Alpine orogenic belt (Mongelli, 1983).

Zone B (South Tuscany)

The velocity functions of Fig. 4 do not show inversions in the upper crust. The velocity of 6.0 km/s is reached at a relatively small depth (between 1 and 3 km). Instead, a strong inversion is seen in the lower crust, probably at the "M" transition. This has been interpreted as evidence of high temperatures in the lower crust (Giese et al., 1980).

As already mentioned, the number of foci is too small to attempt an interpretation as for zone A. One possibility seems to be that the high temperature at the "M" transition could lower the strength of the material of the upper mantle thus producing a ductile behaviour there also. According to Calcagnile and Panza (1981), the lithospheric "lid" seems to be very thin and "weak" here.

Zone C (Calabrian Arc)

The results, already illustrated in Fig. 5, seem to support the idea of a decoupling of the upper crust which is formed by "crystalline" nappes overlying the sedimentary layers, both being highly mobilized, allochthonous formations.

The strong seismicity concentrated in the lower crust or at the top of the mantle lid, suggests a strong and relatively cold lower lithosphere; this could also be related to the subduction environment.

Discussion and conclusions

The increase of available information, both in hypocentres and crustal structure, stimulates new and more precise studies on the seismogenesis of shallow earthquakes in the continental crust. The crustal models that are proposed nowadays, in some way repeat the model of the dynamic relationship between the lithosphere and the asthenosphere, the former being considered as a "sandwhich" formed by several "slices" alternatively brittle and ductile (Chen and Molnar, 1983; Sibson, 1982).

The earthquakes should originate at the transitions between one slice and another, the primary source being located in the deepest levels, probably at the lithosphere-asthenosphere boundary.

Several factors are responsible for this behaviour. For a fairly uniform type of material (as e.g. supposing the upper crust to be formed of quartzitic rocks), the main variables seem to be temperature and pressure as well as the ratio between the pore pressure and external pressure. When the type of material changes, as in olivine rocks, the models should be modified accordingly. So far, suitable models have been proposed only for the "granitic" rocks of the upper crust: shear resistance versus depth profiles have been calculated, also considering the type of rupture (compression, extension, strike-slip mechanism).

In the Italian region this type of investigation faces many difficulties. First of all, the lithospheric structure is subjected to very strong lateral discontinuity (see Fig. 2); second, the direction of stress is often variable from the upper to the lower crust; third, accurate hypocentral locations by dense seismographic networks are available only from the past two or three years and only in some areas; in most areas the accuracy is not sufficient to observe reliable "fluctuations" of the thickness of the active layer. The use of aftershock sequences would help considerably and the results obtained in zone A seem encouraging. However, further experiments are needed.

Nevertheless, we believe that the experimental findings that have been shown are a good starting point. We consider the comparison with the distribution of *P*-wave velocities with depth to be especially useful. In spite of many constraints, the crustal structure derived from the *P* velocity functions, seems to agree with the behaviour of the seismic activity, provided that the latter is accurately determined. On the other hand, these results support the interpretation of DSS by adding proof of the existence of velocity inversions, which have in the past, been the subject of very long controversies.

It is also apparent that the negative velocity gradient within these low-velocity zones is generally larger than that predicted by models (Christensen, 1979) which consider only the increase of temperature and pressure. This is probably due, in part, to the ambiguity involved in the interpretation, but could also be additional evidence for the need to take other factors, such as pore pressure and change of material properties, into consideration.

It is clear that other types of data would be useful, as, for instance, *S*-wave velocities obtained both from the inversion of passive seismic data and from seismic active profiles.

In conclusion, the above results have confirmed that the shallow seismicity in the Italian area is strongly dependent on crustal and lithospheric structure. In zone A, the seismic active layer is clearly directly correlated to the boundary between a brittle upper crust and a weaker (or more plastic) middle-lower crust. In zone B the regime (very shallow foci) seems dependent on the high temperature of the lower crust or of the mantle itself. Finally, in zone C, the seismicity seems to originate in the lower crust or in the lithospheric lid and, therefore, it appears as more directly related to the upper mantle structure.

The study will be extended to other areas, as soon as suitable data become available, in order to verify the general trend described in Fig. 1.

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