On the data integration of seismological observation and deep seismic exploration. The crustal modeling for the Messina Straits Area as an example

A. Bottari* - B. Federico*

Received on June 6th, 1981

ABSTRACT

Given an earth crust model having homogeneous layers and spherical symmetry, the travel-times equations of the direct and refracted longitudinal waves are first deduced. These linear equations are determined from the seismic ray parameter p_i and from the known term a_i. The parameter p_i is connected with the velocity of the elastic waves that are refracted on the i-th layer; a_i depends on the velocity of the layers crossed by the seismic ray and on their thickness.

Subsequently, the equations of the first arrivals, observed in quite a number of stations surrounding the Straits of Messina during the seismic crisis of the Gulf of Patti (1978), are utilized for working out a crustal model for the Messina Straits area. Given the inadequatly of the passive seismology data both in number and in characteristics the most superficial layers of the model are defined using the results of deep seismic exploration carried out in areas contiguous to the Straits. The thickness and velocity of the longitudinal waves in the underlying layers are, instead, determined

* Istituto Geofisico e Geodetico, Università di Messina (Italy).

using a method based on the δT, $\delta\Delta$ allowance calculation (travel time, angular distance) from the relation $a = 2\delta T - 2p\delta\Delta$.

This allows the observed travel time to be reduced to the level of the discontinuities revealed.

When compared with the Jeffreys and Bullen standard crustal model, the one proposed here is appreciably "slower" for the first six kilometers of depht; whereas it is in satisfactory agreement with the Jeffreys and Bullen model for the longitudinal wave velocities in the lower crust, while being on the whole thicker (39 Km as opposed to 33 Km).

With respect to the other models deduced from deep seismic exploration in contigous areas — both Tyrrhenian and Ionian — the greatest differences are found for the Etna area.

RIASSUNTO

Dato un modello di crosta terrestre a strati omogenei e simmetria sferica, sono preliminarmente dedotte le equazioni delle dromocrone delle onde longitudinali dirette e rifratte. Tali equazioni, lineari di primo grado, sono determinate dal parametro del raggio sismico p_i e dal termine noto a_i. Il parametro p_i è legato alla velocità delle onde elastiche che si rifrangono sullo strato i-esimo; a_i dipende dalla velocità degli strati attraversati dal raggio sismico e dalla potenza degli stessi.

Successivamente, le equazioni delle dromocrone dei primi arrivi, osservati in un buon numero di stazioni circostanti lo Stretto di Messina in occasione della crisi sismica del Golfo di Patti (1978), sono utilizzate per l'elaborazione di un modello di crosta per l'area dello Stretto di Messina. Gli strati più superficiali del modello, in mancanza di dati di sismologia passiva adeguati sia per numero che per caratteristiche, sono definiti mediante risultati di esplorazioni sismiche profonde eseguite in aree contigue allo Stretto. Gli strati sottostanti sono invece caratterizzati, in potenza e velocità delle onde longitudinali, mediante una metodologia che, in base al calcolo dei contributi δT, $\delta\Delta$ (tempo di tragitto, distanza angolare) presenti nella relazione

$$a = 2\delta T - 2p\delta\Delta$$

consente di ridurre la dromocrona osservata al livello delle discontinuità rilevate.

Il modello qui proposto (denominato MSM2) confrontato con quello di crosta standard di Jeffreys e Bullen, risulta apprezzabilmente più « lento » per i primi 6 Km di profondità ed in soddisfacente accordo per le

velocità delle onde longitudinali nella crosta inferiore. Risulta comunque nel complesso più spesso (39 invece di 33 Km).

Rispetto ad altri modelli dedotti in base ad esplorazione sismica profonda in aree contigue — sia tirreniche che ioniche — le maggiori diversità si riscontrano per la zona dell'Etna.

Introduction

A good spatial and temporal characterization of the seismic activity of a certain region depends very much on an appropriate choise for the siting of seismic stations as well as on adopting a crustal model sufficiently appropriate for the area.

This was the problem faced by the AA when they set up a short-period seismic network financed by the National Research Council (CNR) for the Messina Straits area (Italian Project "Geodinamica").

In particular they assumed a plane and layered crustal model to be characteristic of the Straits area (Bottari et al., 1979) and worked out a calculation program for the siting of the hypocenters cf the seismic events occurring in this area (Bottari and Neri, 1980).

The main purpose of this paper is to offer a methodology with examples for the characterizing of a crustal model using both passive seismology data (shortly PS) and active seismology data (shortly AS).

In the examples given, the PS data are those for earthquakes sited by using standard travel times which are considerably approximate for any region having propagation characteristics as particular as those of the Straits of Messina; therefore, they need to be appropriately corrected. On the other hand, the AS data are from deep seismic explorations carried out in the area in question.

CHARACTERIZATION OF A CRUSTAL MODEL

A spherically symmetrical crustal model, layered and having seismic wave velocities uniform in every layer is characterized by the *P* and *S* direct and refracted phase travel-times of the type

$$T = p\,\Delta + a \qquad [1]$$

where T is the travel time (sec), Δ is the epicentral distance expressed in degrees, p is seismic ray parameter expressed in sec.deg^{-1} and a is a constant depending either on the structure of the model or, if the structure is fixed, only on the hypocentral depth. The characterization of [1] presupposes a knowledge of seismic wave velocity distribution at various depths. In the case where the function $v(r)$, expressing the behavior of the velocity with distance r from the center of the model, can be represented in the form $v = a \,.\, r^b$ (with a and b constants), the well-known relations (Bullen, 1953):

$$\Delta_{12} = \int_{r_2}^{r_1} p\, r^{-1} (\eta^2 - p^2)^{-1/2}\, \mathrm{d}\, r \qquad [2]$$

$$T_{12} = \int_{r_2}^{r_1} \eta^2\, r^{-1} (\eta^2 - p^2)^{-1/2}\, \mathrm{d}\, r \quad , \qquad [3]$$

that define the allowances Δ_{12} and T_{12} for a seismic ray segment of parameter p, with ends at levels r_1 and r_2 (with $r_1 > r_2$). They can be integrated and offer the following solutions

$$\Delta_{12} = (1 - b_{12})^{-1} \left(\cos^{-1} \frac{p}{\eta_1} - \cos^{-1} \frac{p}{\eta_2} \right) \qquad [4]$$

$$T_{12} = (1 - b_{12})^{-1} [(\eta_1^2 - p^2)^{1/2} - (\eta_2 - p^2)^{1/2}] \qquad [5]$$

where

$$\eta_1 = \frac{r_1}{v(r_1)} \quad , \quad \eta_2 = \frac{r_2}{v(r_2)}$$

For a seismic ray having its deepest point in a given spherical layer, and for which the condition $\frac{dv}{dr} < \frac{v}{r}$ is met, one has:

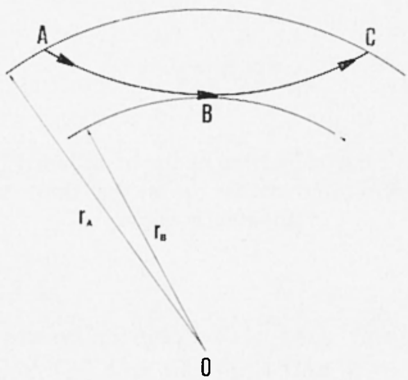

Fig. 1 - Outline of seismic ray in a spherically stratified model.

$$\Delta_{AC} = 2\Delta_{AB} = 2(1 - b_{AB})^{-1} \cos^{-1} \frac{p}{\eta_A} \qquad [6]$$

$$T_{AC} = 2T_{AB} = 2(1 - b_{AB})^{-1} (\eta_A^2 - p^2)^{1/2} \qquad [7]$$

where A and C are the ends of the portion of ray contained in the layer and B is the deepest point of the ray. By means of [6] and

[7] Δ and T (angular distance and travel time) are calculated for the whole path of the seismic ray.

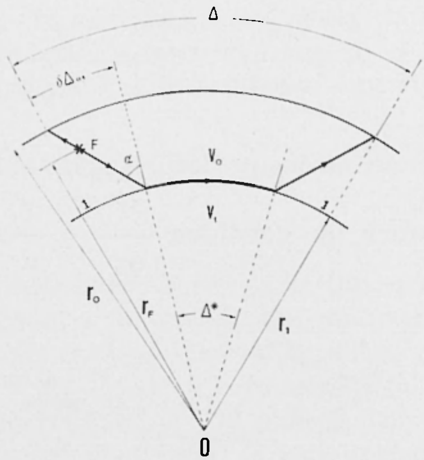

Fig. 2 - Seismic ray refracting at the boundary 1,1 of two horizontal homogeneous layers: α is the limit value of the incidence angle.

In the particular case of a spherically symmetrical layered crustal model, $b = 0$, and thus [4] and [5] become:

$$\Delta_{12} = \cos^{-1} \frac{p}{\eta_1} - \cos^{-1} \frac{p}{\eta_2} \qquad [8]$$

$$T_{12} = (\eta_1^2 - p^2)^{1/2} - (\eta_2^2)^{1/2} \qquad [9]$$

More generally, [8] and [9] allow the "stripping" to be applied to reduce the travel-time revealed on the surface ($r = r_0$) to (the corresponding one for) the level $r = r_h$, where the known term of the travel time vanishes. Thus once the crustal model is fixed, the problem of how to determine the P and S waves travel-time (direct, refracted and reflected) can be solved, as can also the inverse problem.

To solve the direct problem, it must first of all be borne in mind that when the incidence angle of the seismic ray on the discontinuity surface is equal to the limit angle α (Fig. 2), one has

$$\frac{v_0}{v_1} = \text{sen}\ \alpha \quad , \quad p = \frac{r_0}{v_0}\ \text{sen}\ \alpha = \frac{r_1}{v_1} = \eta_1$$

If Δ is the distance at which the seismic ray refracted on the discontinuity 1,1 emerges, and Δ^* is the angular distance on the refractor horizont, one obtains:

$$\Delta = \Delta^* + 2\,\delta\,\Delta_{01}$$

$$T = T^* + 2\,\delta\,T_{01} = \frac{\Delta^*}{v_1} + 2\,\delta\,T_{01}$$

$$T = \frac{\Delta - 2\,\delta\,\Delta_{01}}{v_1} + 2\,\delta\,T_{01} = p(\Delta - 2\,\delta\,\Delta_{01}) + 2\,\delta\,T_{01} \quad .$$

Thus

$$T = p\,\Delta + 2\,\delta\,\Delta_{01} - 2\,p\,\delta\,T_{01} = p\,\Delta + a$$

a being immediately calculable by means of relations [8] and [9]. Thus, the equation of the first refracted phase is

$$T_1 = p_1\,\Delta + a_1 \quad .$$

Similarly, taking into account the allowances $\delta\Delta$ and δT for the layers underlying the layer in question, one obtains for the i-th layer:

$$T_i = p_i \Delta + a_i \ .$$

When the travel times for superficial foci are deduced in this way, the corresponding ones for foci having $h > 0$, can be obtained from them by subtracting the allowances [8] and [9].

When the relations $T = T(\Delta)$ are known for a certain number of phases P_o, P_I ..., P_n, the inverse problem can be solved in a spherical model with concentric layers and velocities v_o, v_1, ..., v_n

i) $$T_0 = p_0 \Delta$$

ii) $$T_1 = p_1 \Delta + a_1$$

$$\cdots\cdots\cdots$$

$$T_n = p_n \Delta + a_n$$

From i) $v_o = r_o \pi / 180\ p_0$ ($r_0 = 6471$ Km). *From* ii) the apparent velocity $v_1 = r_o \pi / 180\ p_1$, can be deduced, and thus the "stripping" is performed for various values of r until

$$2\,\delta\,\Delta_{r_o r} = 2\left(\cos^{-1}\frac{p}{\eta_{r_o}} - \cos^{-1}\frac{p}{\eta_r}\right) \qquad [10]$$

$$2\,\delta\,\Delta\,T_{r_o r} = 2\,(\eta_{r_o}^2 - p_1^2)^{1/2} - (\eta_r^2 - p_1^2)^{1/2} \qquad [11]$$

make the known term a_1 of ii) vanish. If r_1 is the level where this condition or the equivalent condition $2\delta T_{r_o r} - 2p\delta\Delta_{r_o r} = a_1$, is satisfied, then $r_o - r_1 = h_1$, is the thickness of the first layer with a velocity of v_o and $v_1 = r_1 \pi / 180\ p_1$ is the real velocity in the underlying layer.

An analogous procedure can be applied for the other travel times, and moreover, the travel-time valid at the surface can be reduced to a level below which it can be assumed that the velocity varies with r according to a law of the type $v(r) = a\,r^b$. As is known, when the travel time reduced to such a level the Herglotz-Wiechert method can be applied; once the condition $\frac{dv}{dr} < \frac{v}{r}$, has been satisfied this method allows the distribution $v = v(r)$ to be calculated in the underlying zone.

A Crustal Model for the Messina Straits Area

The results of an elaboration of a crustal model for the Messina Straits area area reported (cf. Appendix) as an example of active and passive seismological data integration.

The Gulf of Patti seismic crisis in the spring of 1978 involving north-eastern Sicily gave rise to several hundreds of recordings at the stations of the Messina Straits network and of the Aeolian Islands.

The epicentral distances cover an interval of about 0.7 degrees; this has been extended to 2 degrees, by utilizing data of Calabrian earthquakes as well as observations from the stations of the Etna network.

A first review of the data has concerned the first arrivals which correspond to differing with different epicentral distance. The distance and travel-time values are deduced by utilizing the I.N.G. hypocentral calculations; these utilize a two-layered crustal model with parameters $h_1 = 25$ Km, $v_1 = 5.3$ Km sec^{-1}; $h_2 = 27$ Km, $v_2 = 6.6$ Km sec^{-1}.

This kind of model is substantially in disagreement with the results of seismological research and those of deep prospecting carried out in area in question (Cassinis et al., 1963; Colombi and Scarascia, 1973; Bottari and Girlanda, 1974; Giese and Morelli, 1975; Morelli et al., 1975). However, it does not appreciably alter the estimated positions of the epicenters if the spatial distribution of the stations is favorable — as it is in the case in point — since they are azimuthally well displayed around the source.

The data are divided into two groups for focal depth 10 ± 5 Km and 20 ± 5 Km respectively. These, in fact, are the mean values deducible from the I.N.G. determinations. For each of these two groups two distinct alignments appear, revealing distinct refractor horizons.

Since the crustal model conditions the origin time and the travel times but not the slope of the travel-time segments, the apparent velocities of the refracted waves can be characterized along the horizons revealed.

For the first group one has

$$T = (1.26 \pm 0.30) + (19.032 \pm 0.536)\,\Delta, \quad v = 5.84 \text{ Km}\cdot\text{sec}^{-1} \quad [12]$$

$$T = (3.86 \pm 1.02) + (17.136 \pm 0.584)\,\Delta, \quad v = 6.49 \quad \text{»} \quad [13]$$

and for the second group

$$T = (2.51 \pm 0.11) + (17.337 \pm 0.212)\,\Delta, \quad v = 6.41 \quad \text{»} \quad [14]$$

$$T = (5.97 \pm 0.29) + (14.226 \pm 0.270)\,\Delta, \quad v = 7.82 \quad \text{»} \quad [15]$$

First of all it can be noted that the parameter p uncertainty in [13] and [14] makes only the first decimal figure of the respective velocities significant (6.49 and 6.41 $\text{Km} \cdot \text{sec}^{-1}$). Therefore, the refractor horizon velocities revealed by the linear approximations of the two data groups are respectively 5.8 and 6.5 for group I and 6.4 and 7.8 for group II.

The first problem is that of setting the hypocentral depth in the new proposed model and of characterizing that model by means of the travel times observed. Given the insufficiency of data for foci where $h < 5$ km, the most superficial parts has been "modeled" by introducing two layers

$$h_1 = 2\ \mathrm{Km}, \quad v_1 = 3\ \mathrm{Km} \cdot \mathrm{sec}^{-1}; \quad h_2 = 4\ \mathrm{Km}, \quad v_2 = 4.5\ \mathrm{Km} \cdot \mathrm{sec}^{-1}$$

that are compatible with some deep seismic profiles carried out in the area (Cassinis et al., 1969; Colombi and Scarascia, 1973; Giese and Morelli, 1975). On the basis of such an integration, the direct wave equation for a superficial focus is

$$T = 37.065\,\Delta \quad , \qquad v_1 = 3.0\ \mathrm{Km} \cdot \mathrm{sec}^{-1} \quad . \tag{16}$$

The refracted wave equation at the base of the first layer ($v = 4.5$ Km $\cdot$ sec^{-1}) is of the form

$$T_1 = 24.710\,\Delta + a_1 \quad . \tag{17}$$

On the basis of known elements (η_0, η_1, p) the known term of [17] is directly calculable from the relation

$$2\,\delta T - 2\,p_1\,\delta\Delta = a \quad .$$

One obtains (Appendix, 1)

$$T_1 = 24.710\,\Delta + 0.99 \quad . \tag{18}$$

Similarly, the travel-time of the waves refracted on the upper surface of the third layer is

$$T_2 = 19.032\,\Delta + a_2 \quad , \tag{19}$$

where (Appendix, 2)

$$a_2 = 2\,(\delta\, T_{01} - p_2\, \delta\, \Delta_{01} + \delta\, T_{12} - p_2\, \delta\, \Delta_{12}) = 2.277 \quad .$$

Therefore, the travel-time equation

$$T_2 = 19.032\, \Delta + 2.28 \qquad [20]$$

shows a known term higher than that of equation [12]. Since, for the reasons said, the term a depends both on the depth of the hypocenter and on the velocity of the layers, the focal depth value that is compatible in the new model with the term $a = 2.28$ sec must be calculated. To this end, it is enough to show that the difference between the known terms of [20] and [12] is equal to the allowance $\delta T_{02} - p\delta\Delta_{02}$. That occurs (Appendix, 3) when the focus is at the level $r_1 = 6365.86$ Km, corresponding to a depth of $h_1 = 5.14$ Km.

Therefore, the equation [12] corresponds to the longitudinal waves refracted on the horizon having a velocity of 5.85 Km $\cdot$ sec^{-1} and originating at a depth of 5.14 Km.

For calculating the thickness of a layer having a 5.85 Km$\cdot$sec^{-1} velocity and overlying one with a 6.5 Km$\cdot$sec^{-1} velocity, equation [13] is used which, as has been shown, corresponds to a focal depth of $h = 5.14$ Km. Discrepancies appear in the coefficients of equations [13] and [14] relative to the same phase; they involve the difference between the values of the apparent velocities of the waves refracted at the upper limit of the fourth layer. A mean value of 6.5 Km$\cdot$sec^{-1} has been given to the velocity of the fourth layer, and therefore to the corresponding seismic ray parameter a value of 17.107 sec$\cdot$grad^{-1}. It follows that the travel-time equation [13] can be written

$$T = 3.86 + 17.107\,\Delta$$

and the corresponding one for a superficial focus is obtained by adding to the known term 3.86 the allowances (Appendix, 4)

$$a = 3.86 + 1.10 .$$

Finally

$$T = 17.107\,\Delta + 5.00 \quad . \qquad [21]$$

The 6.5 Km · sec^{-1} discontinuity level can be calculated from putting the condition (Appendix, 5)

$$2\,\delta T_{01} - p_3\,\delta\Delta_{01} + 2\,\delta T_{12} - p_3\,\delta\Delta_{12} + 2\,\delta T_{23} - p_3\,\delta\Delta_{23} = 5.00 \quad .$$

The condition is satisfied for $r = 6348.09$ Km corresponding to a depth of $h = 22.9$ Km. Thus the thickness of the third layer is 16.9 Km.

The mean focal depth value of the second group of earthquakes must be calculated in the new model in order to utilize the relative travel-times in a similar way. Meanwhile [14], in view of what has been said about the apparent velocities, can be written

$$T = 17.107\,\Delta + 2.51 \quad .$$

To calculate this focal depth one must put (Appendix, 6)

$$\delta T_{11} - p_3\,\delta\Delta_{11} = 2.49$$

corresponding to the difference between the known terms of the same phase originating at the surface and at hypocentral level. For the focus one obtains the level $r = 6348.24$ Km corresponding to a depth of 22.76 Km, and the lower limit of the layer having a velocity of 6.5 Km · sec^{-1}, is r = 6348.09 Km, that is to say h = 22.9 Km. The hypocenter of the second group is therefore a little above the 6.5 Km·sec^{-1} discontinuity. That is confirmed by the behaviour of the experimental travel-time [14] that appears linear even for the shortest epicentral distances where the direct longitudinal wave phase has not been observed (Bottari et al., 1978).

By means of equation [15] and of the knowledge of the focal depth of the second group of earthquakes, the thickness of the fourth layer can be finally calculated; its lower limit reveals an apparent velocity of 7.82 Km · sec^{-1}. In fact, [15] is reduced to a surface level r = 6371 Km, by adding to the known term 5.97 the time resulting from the corrections (Appendix, 7). Therefore, the equation for the longitudinal waves originating at the surface and refracting on the Moho is

$$T = 14.226\,\Delta + 9.22$$

Similarly to the case of the third layer the thickness of the fourth layer can be calculated by applying the relation

$$2\,\delta\, T_{04} - 2\, p_4\, \delta\, \Delta_{04} = 9.22$$

and using the allowances given in detail in Appendix 8.

In conclusion the crustal lower limit is h = 38.8 Km corresponding to a spherical level of r = 6332.2 Km, together with a real velocity of 7.77 Km · sec^{-1} and thickness of h = 15.9 Km for the fourth layer.

The essential parameters of the deduced model (shortly MSM2) are reported in Fig. 3.

Compared with the Jeffreys and Bullen (1948) standard model, MSM2 is appreciably "slower" in the superficial part; whereas the difference is less marked in the lower part. On the whole MSM2 is thicker (39 Km) than that of Jeffreys and Bullen model (33 Km).

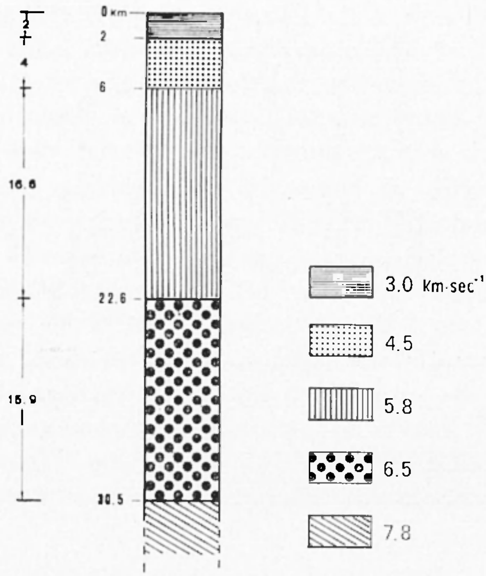

Fig. 3 - Thickness of layers and longitudinal wave velocities of the crustal model MSM2 proposed for the Messina Straits area.

Comparison with a previously worked out model (MSM1) for the same area utilizing the same data but a different procedure (Bottari et al., 1978) shows appreciable differences only for the fourth and final layer, which in MSM2 is faster than 0.1 Km · sec and thicker than 2.4 Km (Fig. 4).

Comparison with some profiles deduced from deep seismic exploration across Tyrrhenian and Ionian areas contiguous to that of the Straits of Messina (Fig. 4) suggests the following considerations.

Sections A, B and C are compatible with the results of various investigations (Cassinis et al., 1969; Colombi e Scarascia, 1973; Giese and Morelli, 1975; Morelli et al., 1975; Bottari et al., 1979) and show overall thicknesses approximately equal ($\pm$ 2 Km) to that of MSM2. In these sections the first kilometers of crust

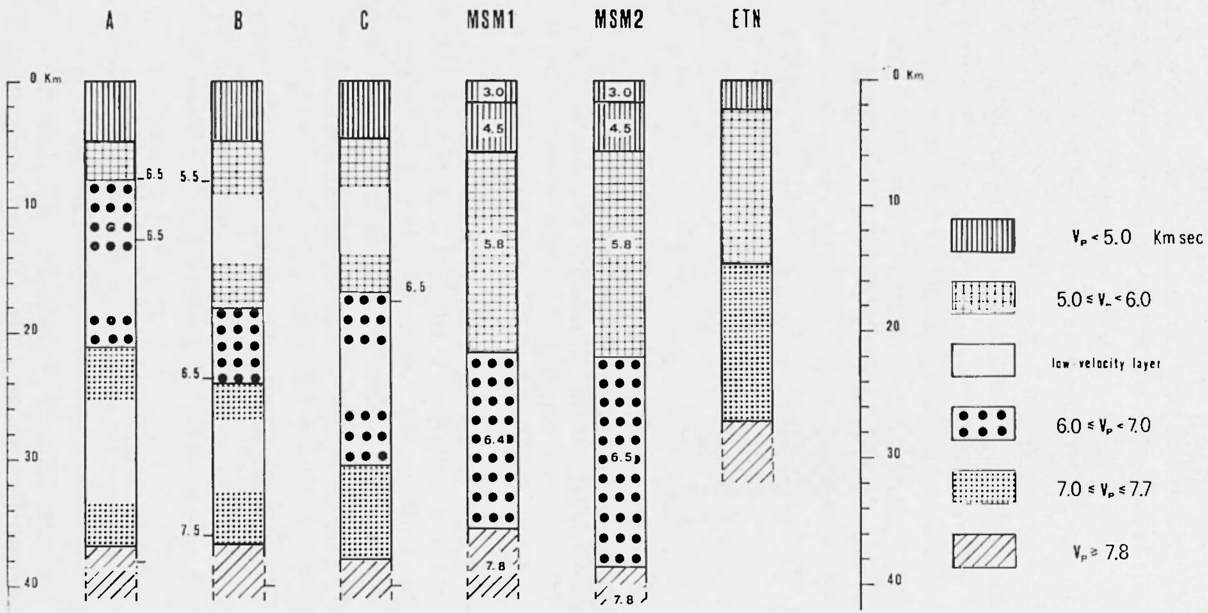

Fig. 4 - Crustal section models for the Messina Straits area (MSM1 and MSM2) and for some contiguous areas: A = Capo Calavà (Northeastern Sicily); B = southern margin of the Peloritan Range; C = Southern Calabria, Ionian Coast; ETN = Etna area.

are not easily comparable because of the different number of layers and the relative velocities of their longitudinal waves. The intermediate zone, however, is characterized by the presence of low-velocity layers and has average velocities of approximately 5.8 Km · sec, which is the value for the third layer in MSM2. Finally, the lower zone of the sections appears the most similar.

The crustal section ETN proposed by Sharp et al. (1979) for the Etna area appears very different. Comparison with it (ETN in Fig. 4) evidences very well the low velocity of the superficial layers of the volcano 2.2 Km · sec^{-1}) as well as, even if much less apparently, the slowish velocity (5.5 Km · sec^{-1}) of its central layer with its base situated at a depth of 16 Km. Finally, in the lower crust the seismic wave velocity (7.2 Km · sec^{-1}) is the lowest of all the models considered. Lastly, the overall thickness of the crust below Etna is decidedly less (27 Km) than the corresponding values proposed for the sections of the other areas.

APPENDIX

N°	r_i (Km)	$r_{i,i+1}$ (Km)	p sec·grad^{-1}	$\delta\Delta_i$ (grad)	$\delta T_{i,i+1}$ (sec)	a (sec)
1	6371	6369	24.710	0.016	1.789	0.990
2	6371 6369	6369 6365	19.032	0.011 0.043	0.777 1.395	1.144+ 1.133 2.277
3	6371 6369	6369 6365.86	19.032	0.011 0.034	0.777 1.100	0.572+ 0.447 1.019
4	6371 6369	6369 6365.86	17.107	0.010 0.027	0.751 0.971	0.591+ 0.505 1.096
5	6371 6369 6365	6369 6365 6348.38	17.107	0.019 0.069 0.656	1.503 2.465 13.842	1.183+ 1.282 2.482 4.947
6	6371 6369 6365	6369 6365 6348.87	17.107	0.009 0.035 0.300	0.751 1.232 6.339	0.591+ 0.641 1.204 2.436
7	6371 6369 6365	6369 6365 6348.87	14.226	0.007 0.025 0.164	0.722 1.087 4.167	0.616+ 0.727 1.833 3.176
8	6371 6369 6365 6348.38	6369 6365 6348.38 6332.50	14.226	0.015 0.051 0.338 0.436	1.444 2.175 8.587 8.894	1.231+ 1.453 3.777 2.684 9.145

References

Bottari, A., Girlanda, A. 1974 - *Some results for the Middle Western Mediterranean basin from the study of Pn Waves.* "Bull. Seism. Soc. Am." 64, pp. 427-435.

Bottari, A., Broccio, D., Federico, B., Lo Giudice, E., 1979 - *Preliminary crustal model from seismological observations at the Messina Straits network.* « Annali di Geofisica », XXXII, pp. 91-111.

Bottari, A., Lo Giudice, E., Schiavone, D., 1979 - *Geophysical study of crustal section across the Straits of Messina.* "Annali di Geofisica", XXXII, pp. 241-261.

Bullen, K.E., 1953 - *An introduction to the theory of seismology.* Cambridge University Press.

Cassinis, R., Finetti, I., Giese, P., Morelli, C., Steinmetz, C., Vecchia, O., 1969 - *Deep seismic refraction research on Sicily.* « Boll. Geofis. Teor. Appl. », 12, pp. 140-160.

Colombi, B., Scarascia, S., 1973 - *Sulla interpretazione dei profili sismici crostali. Calcolo diretto delle funzioni velocità-profondità.* « Riv. Ital. Geofis. », 22, pp. 213-226.

Finetti, I., Morelli, C., Zarudzki, E., 1970 - *Reflection seismic study of the Tyrrhenian Sea.* « Boll. Geof. Teor. Appl. », 12, pp. 311-346.

Jeffreys, H., Bullen, K.E., 1958 - *Seismological Tables.* Reprinted by Percy Lund, Humphries. The Country Press, Bradford (U.K.).

Morelli, C., Giese, P., 1975 - *Structural model of Italy.* « La ricerca scientifica », n. 90, CNR, Roma.

Morelli, C., Giese, P., Cassinis, R., Colombi, B., Guerra, I., Luongo, G., Scarascia, S., Schutte, K.G., 1975 - *Crustal structure of Southern Italy. A seismic refraction profile between Puglia-Calabria-Sicily.* « Boll. Geofis. Teor. Appl. », 17, pp. 182-210.

Scarascia, S., Colombi, B., 1971 - *Interpretazione preliminare del profilo sismico a rifrazione profonda in Calabria.* « Lab. Geof. Litosfera », CNR, pp. 1-35.

Sharp, A.D.L., Davis, P.M., Gray, F., King, G.C.P., 1980 - *Preliminary results of a seismic array study.* U.K. Res. on Mt. Etna - 1977-78-79, pp. 15-18.