J. Geophys. 45, 319–327, 1979

Journal of
Geophysics

Crustal and Upper Mantle Structure Beneath the Apennines Region as Inferred From the Study of Rayleigh Waves*

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Abstract. Rayleigh wave phase velocities were obtained in the period range 12.5–83.3 s using the almost linear array of long period seismic stations installed at Bari (BAI), Grosseto (GSO), Bologna (BOL) and Torino (TNO).

The Hedgehog inversion gives a crustal thickness in the range 25–37 km.

The presence of a low-velocity layer in the crust is allowed, while low-velocity material within a few kilometers of the Moho is required. The shear-wave velocities below 60 km are rather higher than usual channel values. If crustal thicknesses of the order of 37 km are rejected as suggested by other geophysical data then the low-velocity layer in the crust is required in order to satisfy the observed dispersion relation.

Key words: Rayleigh waves – Crust – Upper mantle – Italy.

Introduction

Among other benefits to be derived from such investigations, the study of the upper mantle structure under Italy and the regions adjoining it is important for deriving an understanding of the stress history of interactions at the boundary between the European and African plates. But more interesting in this investigation is the study of crustal properties, in view of the possibility to determine the distribution versus depth of *S*-wave velocity, which gives information about the presence or not of partial melting in crustal layers. The detection of extended areas with soft crustal layers is indeed relevant for the understanding of the crustal deformation, mainly in mountaineous regions. These informations cannot be obtained from the study of body waves because of the non-uniqueness in the inversion to models with a low-velocity zone unless travel-times for reflected waves or for those of a deep seismic source are available (Gerver and Markusevitch, 1966).

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* Publ. No 104 of P.F. Geodinamica-CNR-Roma

0340-062X/79/0045/0319/$01.80

Further interest for such research is represented by the relevant differences in the models of the Apennines region constructed up to now. Caloi (1957, 1958), using earthquake travel-times, has indicated that the North-Central Apennines have a crustal thickness of about 45 km. Close to this result is the value found by Caputo et al. (1976) using surface waves at periods larger than 23 s and an oversimplified crustal model. Much to the south, Colombi et al. (1973) have found a crustal thickness of the order of 35 to 45 km under the westernmost part of the Apennines, thinning to the east, by refraction methods. Based on sketchy information, Giese and Morelli (1973, 1975) have attributed a crustal thickness to this region of 25 to 30 km. More recently Nolet et al. (1978) have found a crustal thickness of 35 km, essentially based on single station group velocity data and over paths east of the area sampled by our data. The consideration of the geological frame (e.g. Elter et al., 1975) makes the difference in location of the aforementioned profiles significant, thus we don't feel confident in combining these two data sets to reduce the possible range of solutions in the inversion process.

Data

The results reported here are obtained from the processing of seismograms recorded for an earthquake which occurred on April 4, 1975, at 38.1 N, 22.0 E, origin time 05:16:16.2, $M_b = 5.4$, focal depth = 53 km.

We have used recordings of fundamental mode Rayleigh waves made on the vertical component long-period seismographs (WWSSN equivalent) installed at Bari (BAI), Bologna (BOL), Grosseto (GSO) and Torino (TNO) (Fig. 1) as part of the Italian Long Period Seismographic Network. The station coordinates are given in Table 1, as well as the ones for Aquila (AQU).

The records were analyzed by a standard time-windowing and frequency-filtering technique (e.g. Pilant and Knopoff, 1964; Biswas and Knopoff, 1974; Panza, 1976).

The station spacing is too short to provide any possible resolution of the structure underneath any subregion spanned by a station pair. The overall span of the array means that the data for all four stations are best used in a joint array-processing analysis. This means that we shall only be able to determine the 'average' dispersion characteristics for the entire region and hence, after inversion, obtain an 'average' structural cross-section for the entire region.

To calculate the 'average' phase velocity without bias toward the phase data for any particular subset of stations we fit a phase versus distance diagram at a given period by a linear regression. Examples of these plots are given for two periods in Fig. 2. The phase slownesses are the slopes of the phase-distance curves. The results are given in Table 2 for selected multiples of the folding frequency.

Figure 3 gives a comparison of the data with dispersion values available for Italy (Caputo et al., 1976; Calcagnile et al., 1979). Significant differences can be observed at periods larger than 50 s, reasonably reflecting the presence of large lateral variations also in the upper mantle in the Italian region.

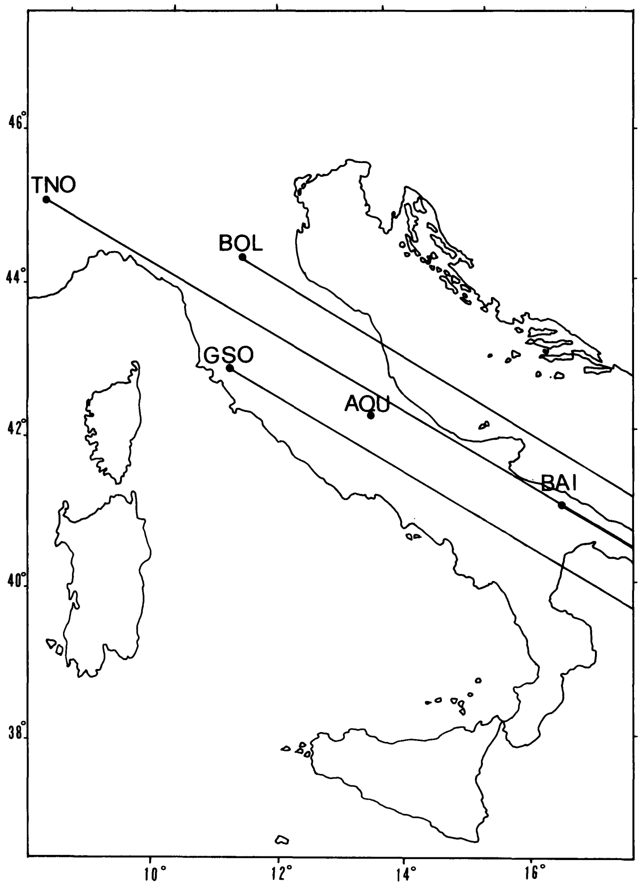

Fig. 1. Wave paths to the stations used in the determination of Rayleigh wave dispersion

Table 1. Seismic stations

L'Aquila	(AQU)	42°21′14.0″N	13°24′11.0″E
Bari	(BAI)	40°52′40.0″N	17°12′13.0″E
Bologna	(BOL)	44°29′12.0″N	11°19′44.4″E
Grosseto	(GSO)	42°45′08.1″N	11°06′58.6″E
Torino	(TNO)	45°03′31.5″N	07°41′49.0″E

Table 2. Phase velocities used in the inversion

Period (s)	Phase velocity (km/s)	Error ε (km/s)
83.3	4.03	0.07
62.5	4.00	0.05
50.0	3.91	0.05
41.7	3.81	0.05
31.3	3.64	0.05
25.0	3.51	0.05
17.9	3.11	0.05
12.5	2.84	0.07

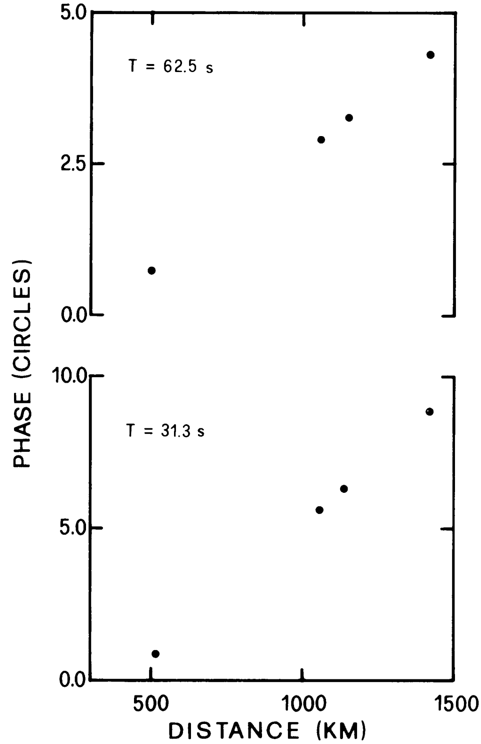

Fig. 2. Examples of plots of phase versus epicentral distance used to determine the average phase velocity. At each period phases are referred to an arbitrary integer

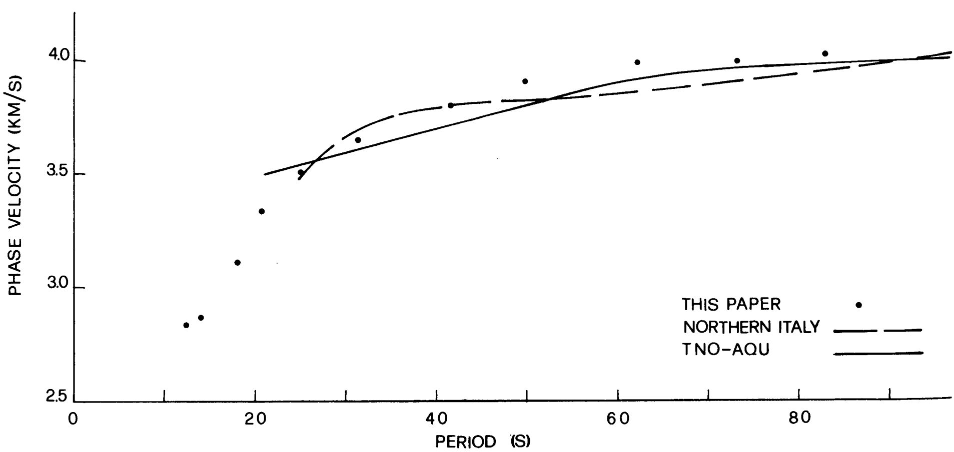

Fig. 3. Phase velocities for the investigated region; dispersion values for adjacent areas are shown for comparison; TNO-AQU from Caputo et al. (1976), Northern Italy from Calcagnile et al. (1979)

Dispersion values analogous to those given in this paper have been obtained by Biswas and Knopoff (1974) in the Colorado plateau region.

Inversion

As indicated, the small spacing between stations prevents us from dividing the span of the array into smaller regions. Accordingly, to invert the dispersion results, we assume that a homogeneous horizontally layered structure underlies the region from Bari to Torino. The assumption of a layered model does not imply a commitment on our part to the presence of sharp discontinuities in properties; they simply point toward a rapid change in the gradient of elastic parameters. The inversion method applied is the well-known Hedgehog procedure. In the inversion, if the difference between computed and observed phase velocities exceeded the limit $\pm\varepsilon$ (see Table 2) at any individual period, the model was rejected. Models which pass these tests were further tested to determine whether the root-mean-square (rms) deviation σ was appropriately small. All models with $\sigma \leqq 0.03$ km/s were finally accepted. These estimates of error are suggested by the scatter of the data in the phase versus distance diagrams.

The cross-section used in the inversion is listed in Table 3. Six model parameters were allowed to vary, namely the thickness of the middle and lower crustal layers and their shear-wave velocity, the thickness and the shear-wave velocity of the lid. Furthermore we tested five values for the upper asthenosphere shear

Table 3. Cross-section used in the inversion

Layer thickness (km)	β (km/s)	α (km/s)	ρ (km/s)
5.0	1.90	3.50	2.00 sediments
10.0	3.10	5.50	2.50 upper crust
P4	P1	6.10	2.75 middle crust
P4	P2	6.90	3.00 lower crust
P5	P3	8.10	3.45 lid
303-2·P4-P5	P6	8.20	3.50 upper asthenosphere
∞	4.85	8.60	3.65

Confidence limits $\sigma = 0.03$ km/s, single point rejection if $|\Delta c| > \varepsilon$ (see Table 2)

Parameter	Range and step	Starting value	Grid variable x_n
P1 (km/s)	2.7 (0.2) 3.9	3.5	$-4 \div +2$
P2 (km/s)	3.1 (0.2) 4.1	3.9	$-4 \div +1$
P3 (km/s)	4.2 (0.1) 4.8	4.3	$-1 \div +5$
P4 (km)	5 (2) 15	9	$-2 \div +3$
P5 (km)	15 (20) 75	35	$-1 \div +2$
P6 (km/s)	4.3 (0.1) 4.7	4.4	$-1 \div +3$

Table 4

Solution number	Middle crust (P1/P4)	Lower crust (P6/P4)	Lid (P3/P5)	r.m.s. σ
(a) Upper asthenosphere P6 = 4.50 km/s				
1	3.3/11	3.9/11	4.3/35	0.028
2	3.5/11	3.7/11	4.3/35	0.023
3	3.3/9	3.7/9	4.2/35	0.024
4	3.3/9	3.7/9	4.3/55	0.026
5	3.3/11	3.9/11	4.3/55	0.029
6	3.5/11	3.7/11	4.3/55	0.027
7	3.3/11	4.1/11	4.2/35	0.023
8	3.3/11	4.1/11	4.3/55	0.027
9	3.5/11	3.7/11	4.4/55	0.026
10	3.5/13	4.1/13	4.3/35	0.023
11	3.5/13	3.9/13	4.4/35	0.025
12	3.5/13	3.9/13	4.3/55	0.029
13	3.5/13	4.1/13	4.2/35	0.029
14	3.5/13	3.9/13	4.4/55	0.022
15	3.5/13	4.1/13	4.2/15	0.030
16	3.5/9	3.5/9	4.2/35	0.022
17	3.5/9	3.5/9	4.3/55	0.023
18	3.3/9	3.5/9	4.3/55	0.028
19	3.5/11	3.7/11	4.4/75	0.024
20	3.5/13	3.9/13	4.4/75	0.024
21	3.5/15	4.1/15	4.4/15	0.027
22	3.5/15	4.1/15	4.5/35	0.028
23	3.5/15	4.1/15	4.5/15	0.030
24	3.5/15	4.1/15	4.5/55	0.028
25	3.5/15	4.1/15	4.5/75	0.028
26	3.5/9	3.3/9	4.3/35	0.028
27	3.7/9	3.3/9	4.3/55	0.027
28	3.3/7	3.3/7	4.2/55	0.027
29	3.5/7	3.1/7	4.2/35	0.026
30	3.7/9	3.1/9	4.4/55	0.027
31	3.5/7	3.1/7	4.2/55	0.027
32	3.7/9	3.1/9	4.4/75	0.024
33	3.5/7	3.1/7	4.3/75	0.028
(b) Upper asthenosphere P6 = 4.60 km/s				
	3.3/9	3.7/9	4.3/75	0.030
	3.5/9	3.5/9	4.3/75	0.028
	3.3/11	4.1/11	4.4/75	0.030
	3.5/13	4.1/13	4.5/55	0.030
	3.5/13	3.9/13	4.5/75	0.028
(c) Upper asthenosphere P6 = 4.40 km/s				
	3.5/15	4.1/15	4.5/75	0.030

velocity obtaining solutions indicating its range of variability. The allowed range of variation for these parameters are given in the lower part of Table 3.

The constant values in the crustal model are suggested by results of DSS and indirectly relevant gravity data (e.g., Giese and Morelli, 1975). Small differences from the constant values in density and *P*-wave velocity are not expected to have a significant influence on the result of the inversion presented in Table 4. Reasonable changes in the first 15 km of the crust do not affect significantly the inversion result. This is not a surprising result if variational parameters in this layer and the error allowed at short periods are considered. Thirty-three successful solutions x_n were obtained, values of x_n are listed as well as the values of the rms deviation between the model curve and the observations over the eight selected periods. The test on the upper asthenosphere shear velocity indicates that its range is limited to values in the range 4.4–4.6 km/s.

Owing to the period range we are dealing with our major concern is the estimate of average crustal properties. It must be mentioned here that, since from the record of station TNO it has been possible to obtain reliable phases only for periods larger than 25 s, the crustal models we will describe are mostly proper for the area covered by the tripartite array BAI-BOL-GSO, to which an average Moho depth less than about 31 km might be assigned according to independent geophysical results (e.g., Giese and Morelli, 1975). Our data are consistent with a crustal thickness in the range 28–37 km. A 'crustal' thickness up to 45 km is compatible with the data but since average Moho depths of 41–45 km are not realistic according to the geophysical results in the area and since the larger crustal thickness requires shear wave velocities for the lower crust which seem too high (3.9–4.1 km/s) we think that solutions n. 10-11-12-13-14-15-20, having an apparent crustal thickness of 41 km, actually correspond to models with total crustal thickness equal to 28 km, and solutions n. 21-22-23-24-25, with apparent crustal thickness of 45 km, give models with a crust of 30 km. Namely they indicate actually the presence of a transition zone rather than an abrupt jump in velocity to the underlying 'soft' mantle with a shear-wave velocity in the range 4.2–4.5 km/s. These crusts overlie mantle material characterized by strong positive velocity gradients. Thus on the basis of dispersion data only we cannot distinguish between crustal models containing or not a significant layer of what is usually called 'gabbro'. If we exclude crustal thicknesses larger than 33 km, i.e., if we exclude a Moho depth equal to 37 km or larger, as suggested by independent geophysical data (e.g., Giese and Morelli, 1975), then only crustal models where the 'gabbro' layer is practically absent are possible, except for solutions 3 and 4. Furthermore the presence of negative gradients in the velocity-depth function is allowed with a velocity contrast as high as 0.6 km/s. The absence of a clearcut 'gabbro' layer is in agreement with the model given by Giese and Morelli (1975) for the Northern Apennines and is also substantiated by the structure proposed by Colombi et al. (1973) for the southernmost part of the area. On the other hand the velocity inversion given by Giese and Morelli (1975) in the aforementioned model around 25 km, is not clearly substantiated by experimental evidence as travel-time for reflected waves or for those of a deep seismic source (Gerver and Markusevitch, 1966), while surface waves data have allowed the determination of a set of models, wherein shear-wave velocity in the low-velocity zone can be in the range 3.1–3.3 km/s, which corresponds to compressional-wave velocity in the range 5.4–5.7 km/s, if the standard relation $v_p = \sqrt{3}\, v_s$ is used. The origin of crustal low-velocity zone is not yet clearly understood, however the low-velocity values in the lower crust may indicate partial melting perhaps in combination with the presence of dehydration water.

Being aware of the fact that we cannot exclude the presence of very thin veneers with high velocity just below the Moho, on the base of our data we can state that the sub-Moho material is characterized by low velocities, never exceeding 4.4 km/s, and that below it the velocity increases with depth or is constant, being in the range 4.4–4.6 km/s.

Conclusions

The crustal structure of the Apennines region, as inferred from Rayleigh wave dispersion measurements, is in good agreement with the model given by Giese and Morelli (1975). More detail is given about the possible low-velocity zone at a depth of about 25 km, which is characterized by *S*-wave velocities in the range 3.1–3.3 km/s. If the presence of a soft layer in the crust at a depth of about 25 km is accepted, in agreement with DSS and surface wave data, then this depth can be considered as the upper limit for the focal depth of crustal shocks in the Apennines. The maximum crustal thickness consistent with the data is 37 km; all crustal models exclude the presence of crustal doubling as suggested for the Northern Apennines by Morelli et al. (1977). At the most *S*-wave velocities in the range 3.9–4.1 km/s can be considered strong indicators of the presence of some kind of transition zone, instead of the usual rapid change in elastic properties associated with the Moho discontinuity. The *S*-wave velocity in the sub-Moho material never exceeds 4.4 km/s and this is in agreement with the low P_n velocities given by Giese and Morelli (1975).

Caputo et al. (1976) from the study of the path TNO-AQU where not able to resolve between two groups of models, one with relatively low-velocity material just below the Moho, the other with a high velocity lid of thickness 30–40 km overlying a very low-velocity channel. On the basis of our results and DSS data this ambiguity seems to be resolved, and the first group of models seems to be more appropriate for the Apennines region.

The presence of high-velocity material, about 4.5 km/s, starting at depths as low as 60 km may be interpreted as evidence of a possible downbuckling process of the high-velocity lid in this region.

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Received May 16, 1978; Revised Version January 15, 1979,
April 20, 1979; Accepted May 10, 1979