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*Original Investigations*

# Crustal Structure of the Central Aegean Sea and the Islands of Evia and Crete, Greece, Obtained by Refractional Seismic Experiments

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**Abstract.** In 1973 and 1974 deep seismic sounding experiments were performed in the Aegean area by German and Greek geophysicists, from which two crustal sections were established. The one strikes along the islands of Amorgos-Mikonos-Andros and Evia and the other along Crete, in the E-W direction. The main results obtained are:

Along the Amorgos-Mikonos-Evia section the crust is updipping from 32 km below Evia to 26 km below Amorgos in the southern Aegean Sea.

The crust is of the continental type with $V_{pg}=6.0$ km/s for the crystalline basement and 7.7 km/s for the upper mantle.

The average velocity of the crust computed from $P_mP$-reflections has a value of 6.21 km/s.

The sedimentary cover is very unevenly distributed with maximum thickness at North Evia. The crystalline basement outcrops at the southern part of the island and the Cyclades.

Along Crete the crust is somewhat thicker, than that below South-Evia with 34 km at the western part and about 30 km at the eastern part of the island.

At the western part of the island the nappes have their greatest thickness, the Messara Basin in the east containing the largest neogene sequences on Crete.

The crustal structure of Crete is also of the average continental type.

**Key words:** Seismic experiments – Crustal structure of Aegean Area, Greece.

## 1. Introduction

In March and April 1973, 28 MARS-66 seismic recording stations (Berckhemer, 1970) provided by various German universities[1] were used to record 8 shots fired east and west of Crete, south of Mikonos and at North Evia (see Fig. 1). The shots with charges ranging between 0.4 and 3 tons, were fired at sea and at a quarry of the "Skalistiris Mining Co." at Mandouthi, North Evia. Thus, by reversing both profiles we intended to obtain true *P*-wave velocities along the seismic sections and the crustal structure and thickness. Extrapolations from the Ionian Sea-Peloponnese seismic section (Makris, 1975, 1976) indicated that the crust below Evia had an approximate thickness of the order of 28–32 km. This result, however, was liable to errors, since no true velocities for the Peloponnese crust had been obtained by the Ionian Sea-Peloponnese seismic experiment of 1971 (Makris, 1972).

The measurements reported below were supported by the National Institute of Geological and Mining Researches (NIGMR) Athens, which provided the explosives and was responsible for permits and for contacting the various Greek authorities.

## 2. Location of the Profiles and Some Technical Details

The Evia-Mikonos profile has a length of 270 km and was observed at 46 locations with an average separation of 5.5 km. Two shots of 3 t. were fired at Mandouthi (SP2), North Evia, and gave excellent recordings over the whole length of the section. The reversed observations were obtained by firing two shots of 0.4 and 0.8 tons at sea south of Mikonos. The first shot (0.4 tons) also gave very good recordings for the first part of the seismic section and partly reversed the observations of the "Mandouthi" shots. The second shot (0.8 ton) could not be recorded due to very bad weather conditions. We repeated our observations on Evia in 1974 by firing a 0.8-ton shot south of Amorgos (see Fig. 1) and completed this section to a fully reversed profile (see Fig. 6).

[1] The following institutes and persons participated in the seismological program:

(1) Bundesanstalt für Geowissenschaften und Rohstoffe, Hannover: Sudhir Garde, Heinz Schröder, Hermann Schubart.

(2) Institut für Geophysik, Technische Universität, Clausthal-Zellerfeld: Tilmann Rothfuchs, Günter Wagenbreth, Michael Leppin, Dietrich Schöneck, Klaus Kaiser, Roland Vees, Friedrich Schildknecht, Bernd Kölln, Hermann Hemschmeier.

(3) Institut für Geophysik, Freie Universität, Berlin: Franz Renger, Peter Röwer, Detlev Damaske, Karl Georg Schütte, Christian Reichert.

(4) Institut für Geophysik, Technische Universität, Karlsruhe: Klaus Schulz, Peter Thorner, Sonja Faber.

(5) Institut für Meteorologie und Geophysik, Universität Frankfurt: Wolfgang Lorig, Timann Spohn.

(6) Institut für Geophysik, Universität Hamburg: Walter Grevsmühl, Hermann Richter, Walter Homann, Helmut Gertig, Jürgen Klußmann, Jürgen Rohde, Klaus Müller, Jannis Makris, Dorothee Ruhle von Berg, Monika Hamann.

(7) National Institute of Geological and Mining Researches, Athen: Agellos Stavrou, Nikos Tsambourakis, Spiros Prionas, Georg Laios

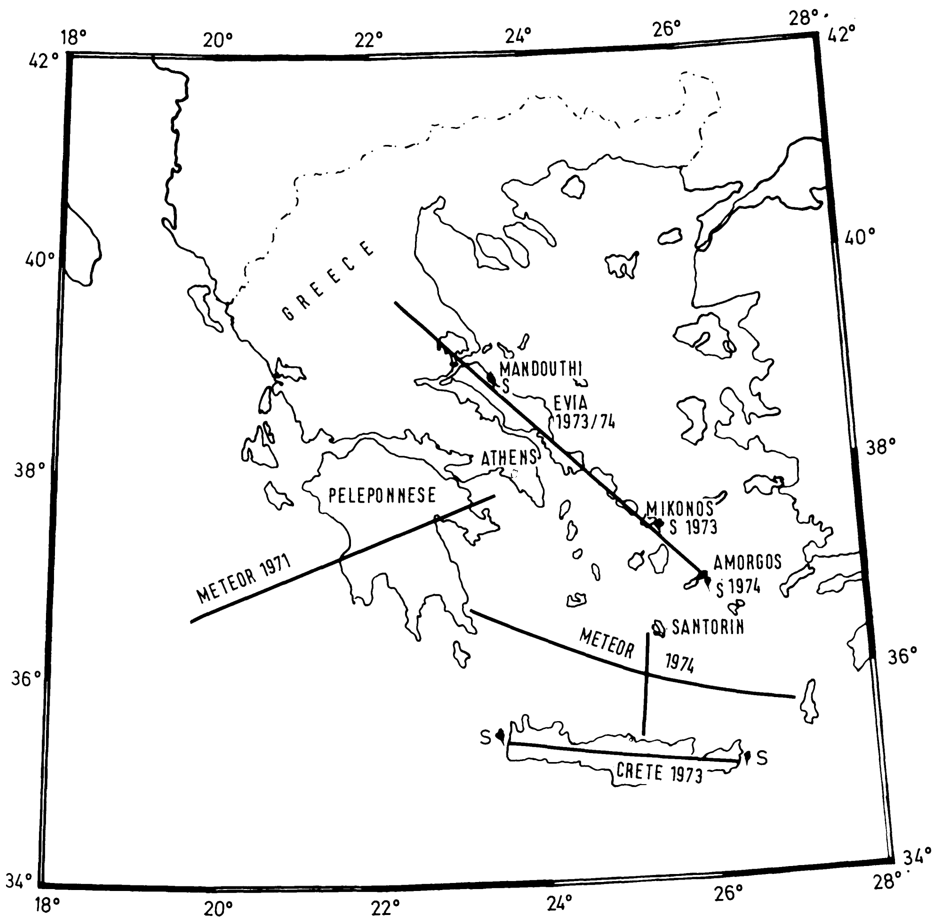



Fig. 1. Locations of profiles and distribution of recording stations and shots

The Cretan section has a length of 260 km. The mean separation of the seismic stations is app. 5.0 km. The section was also observed in two parts by firing four shots at sea; two at the eastern and two at the western coast of Crete. The explosions at the eastern coast gave good recordings all along the seismic section. The reversed profile, however, was not completed (see Fig. 8), since very bad weather conditions did not permit us to adjust the shot depth properly. A great part of the explosion energy was lost in the explosion fountain and no observations could be made at the distant stations.

Time signals and information to seismic stations were transmitted through "Radio Pallini" by Athens at 3.15 MHz. This Radio station had been put at our disposal by OTE (Organization of Telecommunications of Greece) and could be modulated through any telephone of the Greek telephone system. The OTE-organization supported our program very efficiently and contributed considerably to the success of this experiment.

### 3. Processing and Evaluation of Seismic Data

Most analogue tape recordings of both seismic sections were digitized at the CDC-1700 analogue-digital computer of the Institute of Geophysics, University of Hamburg. The data were band-pass filtered between 0.2 and 20 Hz and were plotted in time-distance curves. The time scale of the plots was reduced by 6 km/s, so that signals travelling with this velocity correlate to curves parallel to the distance axis. No elevation corrections were applied, since the locations of the stations do not differ significantly in altitude. The seismic arrivals that were used for the interpretation are the $P_g$, $P_n$ and $P_mP$-phases. The velocity of the sediments was assumed to be 4 km/s. This was necessary, since the sedimentary cover was too thin and the 5 km spacing of the stations too large to permit first-arrivals through the sediments to be recorded. The $P_g$-phase gave true velocities of 6 km/s for all sections. The $P_n$-velocity appears with various apparent values from section to section according to the dip of the Moho-discontinuity. The true value of the $P_n$-phase was computed from the Evia sections and the Amorgos-shot of 1974 to 7.7 km/s. Below Crete the apparent velocity of the $P_n$-phase is 7.54 km/s. Since no reversed observations for this signal group were obtained, it is not possible to give a true $P$-wave velocity of the upper mantle below Crete. The fact, however, that the seismic section was positioned parallel to the strike of the morphological units and the very small gravity gradient along the section (Makris, 1976), indicate that the Moho-morphology in the east-west direction does not change significantly. Therefore, the $P_n$-velocity obtained must be very near to the true one. In any case, both profiles, that of Evia as well as that of Crete show $P$-wave velocities which are smaller than 8 km/s.

Strong PmP-phases were recorded at the Mikonos, Evia and Amorgos sections and permit a more reliable interpretation along the Amorgos-Mandouthi line than at Crete. The evaluation of the PmP-reflections according to the $t^2$-$\Delta^2$-method gave a mean crustal thickness of 29 km and a mean $P$-wave velocity of 6.21 km/s for Evia.

The seismic data were evaluated in the following way: First, the directly correlated apparent velocities and intercept-times were used to construct a model of dipping first order seismic discontinuities. The model was then checked by means of gravity data and its parameters varied until the best fit between computed and observed Bouguer gravity anomalies was achieved. Finally, theoretical travel-times were computed and compared with the observed data. The method used for the computation of the travel-time curves is a modified version of the "trapezoid-method" (Stein, 1968). According to this method the model is divided into equidistant sections along the x-axis which are limited by first order discontinuities at depth. These discontinuities also limit intervals in which a prescribed velocity-depth function is valid. In all computations linear velocity-depth functions, $v^i(z) = v^i{}_0 + a^i z$, are used. The $a^i$ velocity gradients are either automatically computed below the shot-point by defining velocities at the upper and lower limit of each interval, or, in the case of strong lateral variations, are given as constants for each section of the model involved. The travel-times are computed by summing up the time intervals, $\delta t^j_k$ needed by the $j$-seismic

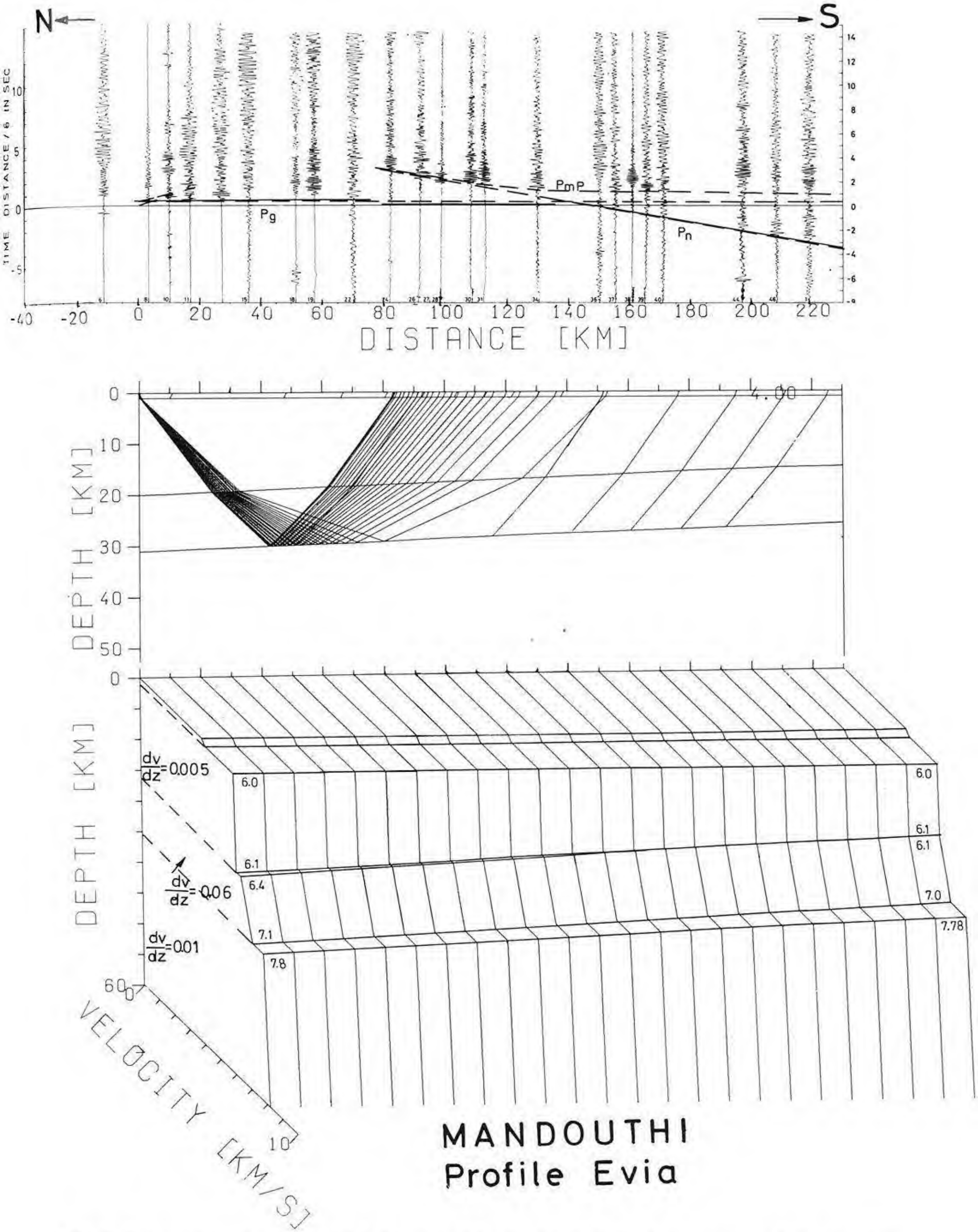
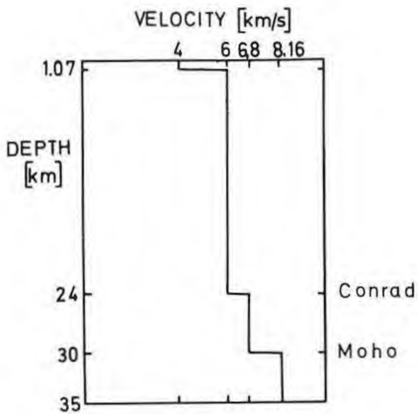



**Fig. 2.** Evia-section observed in N-S direction (Mandouthi-shot). In the upperpart of the drawing the travel-time diagram is given. The time-axis is reduced with 6 km/s. This is valid for all travel-time diagrams presented in this paper. —— lines give the direct correlations. ---- lines give the theoretically computed travel-times according to the model presented at the middle part of the drawing. At the lower part, the $v(z)$-curves along the model and the $P$-wave velocities and gradients $dv/dz$ are given. By introducing a velocity gradient at the lower part of the crust, the best fit between computed and observed travel-times of the $P_g$-$P_n$- and $P_mP$-phases was obtained

Fig. 3. Evia-section observed in N-S direction (Mandouthi-shot) – First order discontinuity model. The velocity – depth function below the shotpoint is given at the upper left part of the drawing. The theoretically computed travel-time curves satisfy the observed $P_g$, $P_mP$ and $P_n$-travel-times. This model has to be rejected however, since no reflections from a 1.-order Conrad-Discontinuity have been observed. —— Directly correlated ---- theoretically computed travel-times · - · theoretically computed Conrad Reflections



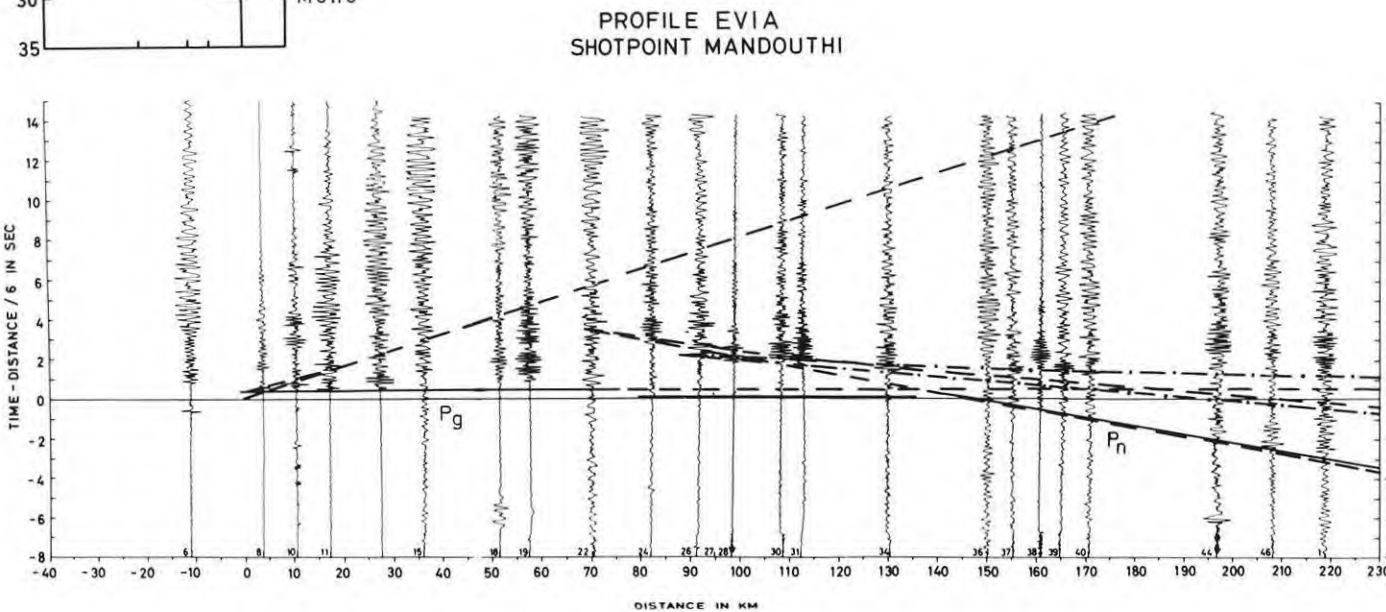
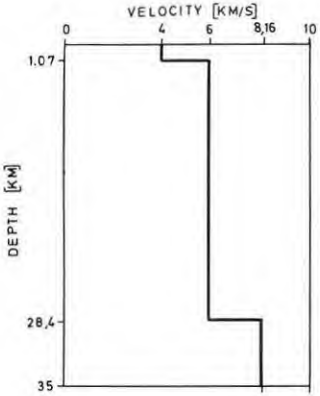
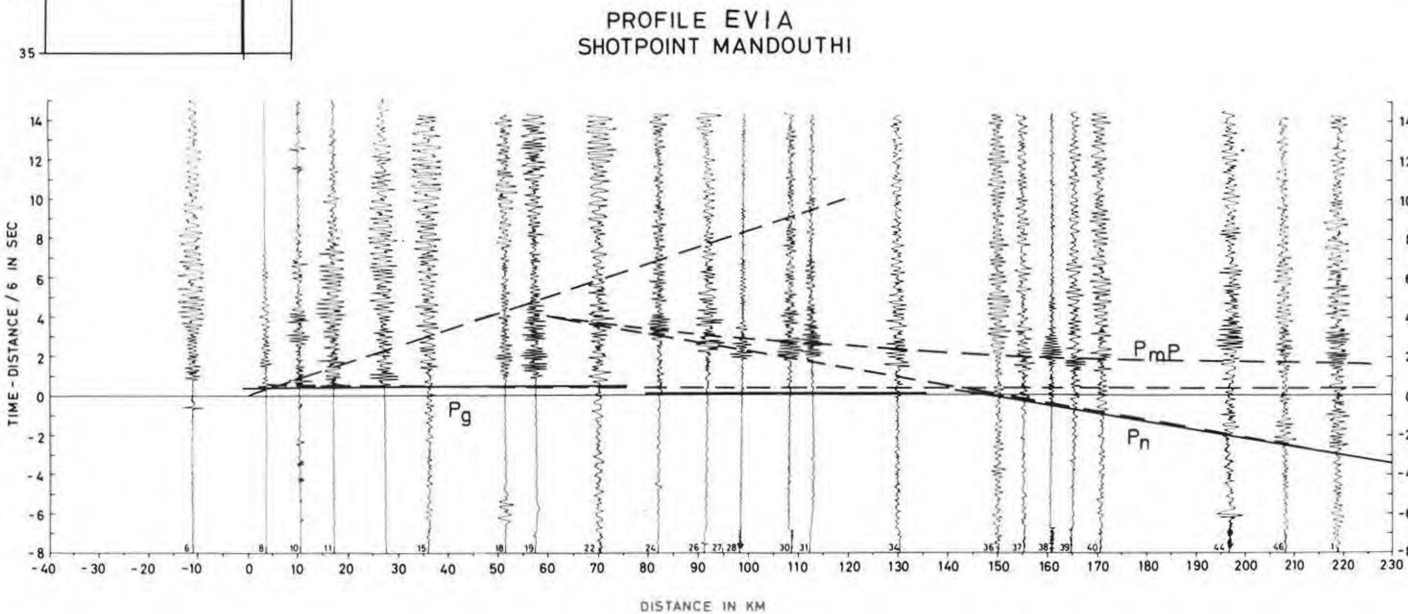



Fig. 4. Evia-section observed in N-S direction. This single-layer crust model limited by a 1.-order Moho discontinuity satisfies the $P_g$- and $P_n$-phases but not the $P_mP$-wide angle reflections. For this reason it is rejected. —— Directly correlated --- theoretically computed travel-times



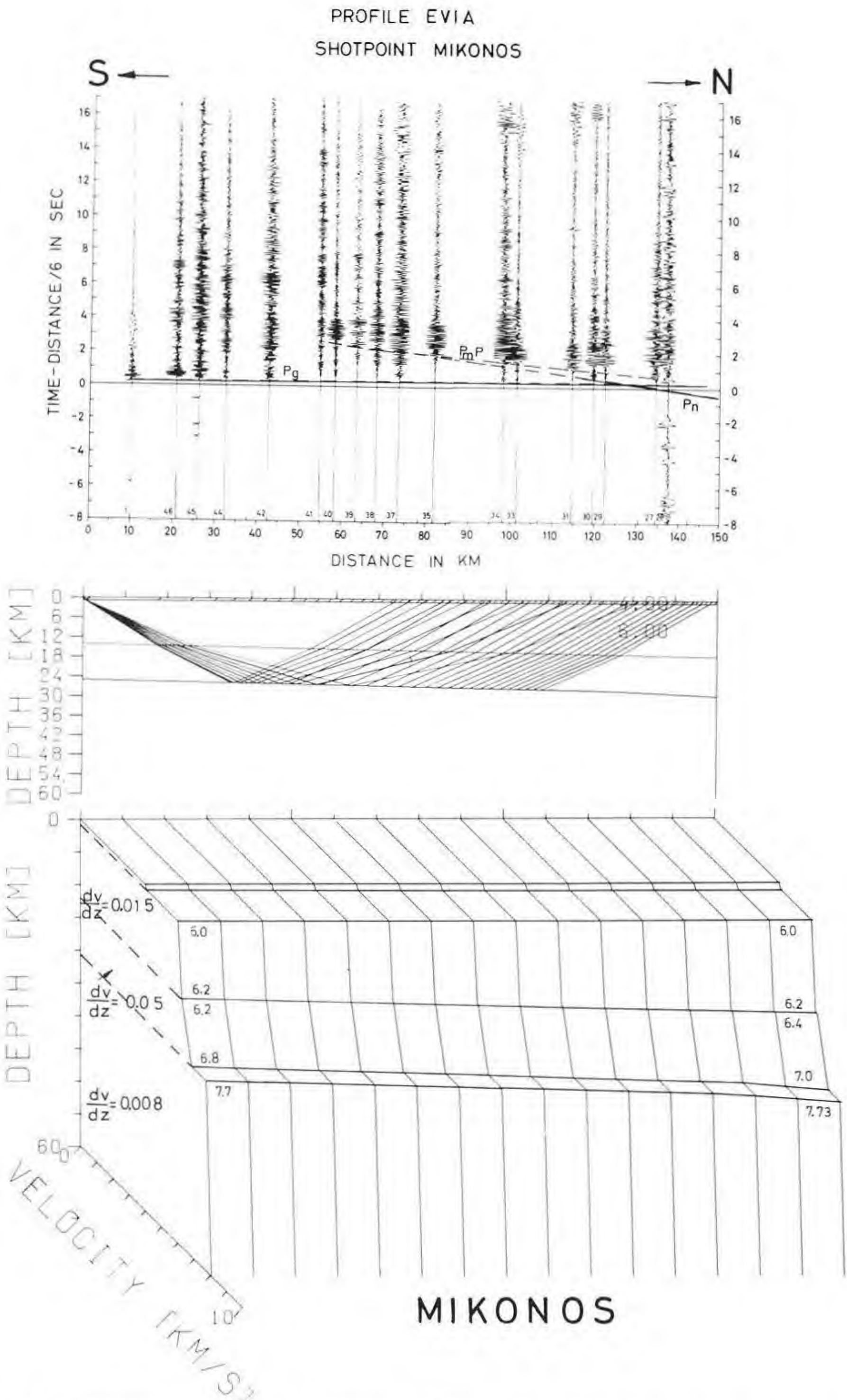



**Fig. 5.** Evia-section observed in S-N direction (Mikonos-shot). Symbols and their meaning as in Figure 2. Due to the shortness of this profile the $P_n$-phase does not exist as first arrival

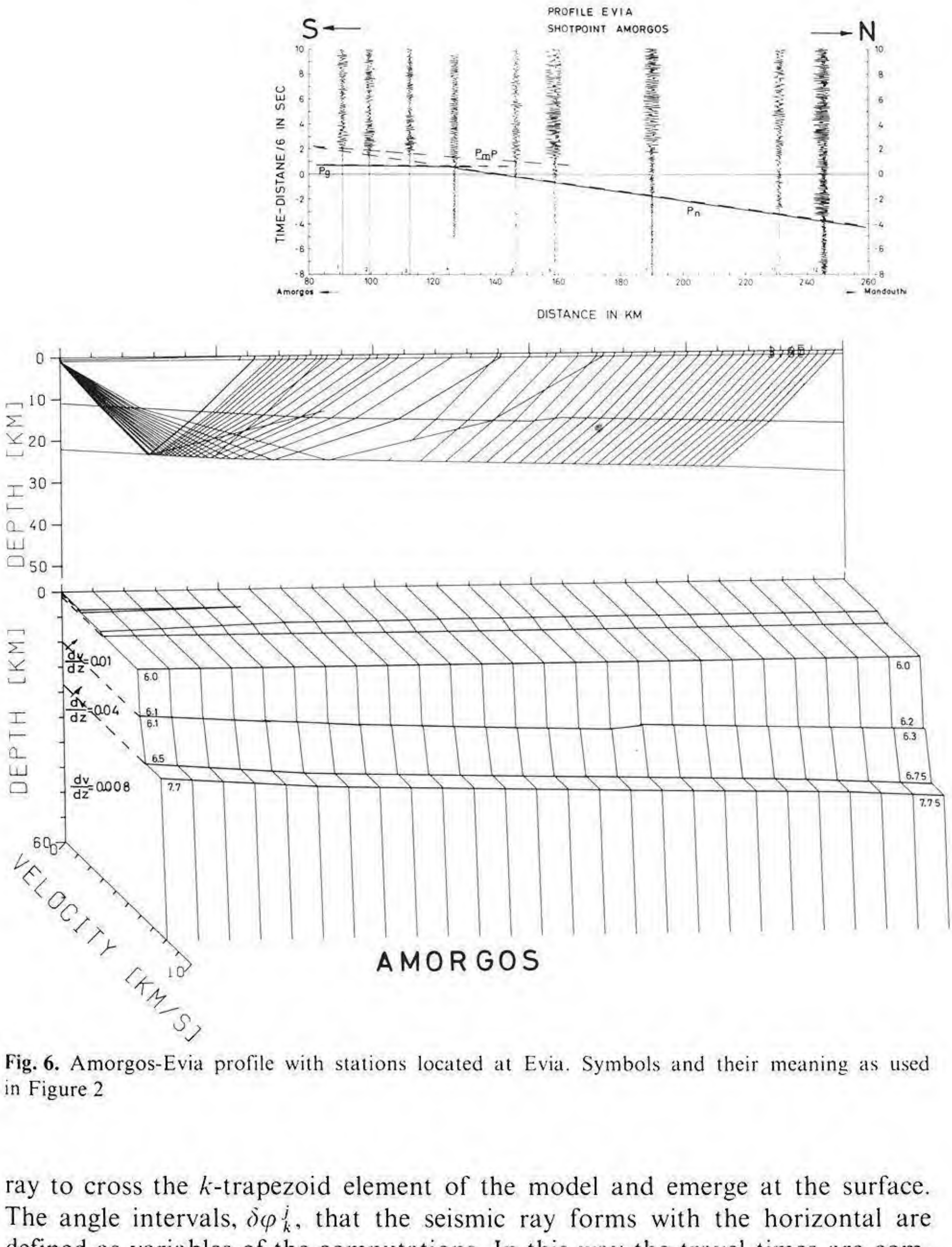



**Fig. 6.** Amorgos-Evia profile with stations located at Evia. Symbols and their meaning as used in Figure 2

ray to cross the $k$-trapezoid element of the model and emerge at the surface. The angle intervals, $\delta\varphi_k^j$, that the seismic ray forms with the horizontal are defined as variables of the computations. In this way the travel times are computed for different ray paths and compared with the observed travel times. The model parameters are iterated until the best fit between observed and computed travel times is achieved. The program is adjusted to a screen display so that a very fast parameter readjustment is possible. The results are then plotted automatically. This computational method is fairly fast, since it is only based on the refraction law and therefore only simple trigonometric functions with a short computation time are used. The accuracy and reliability of the

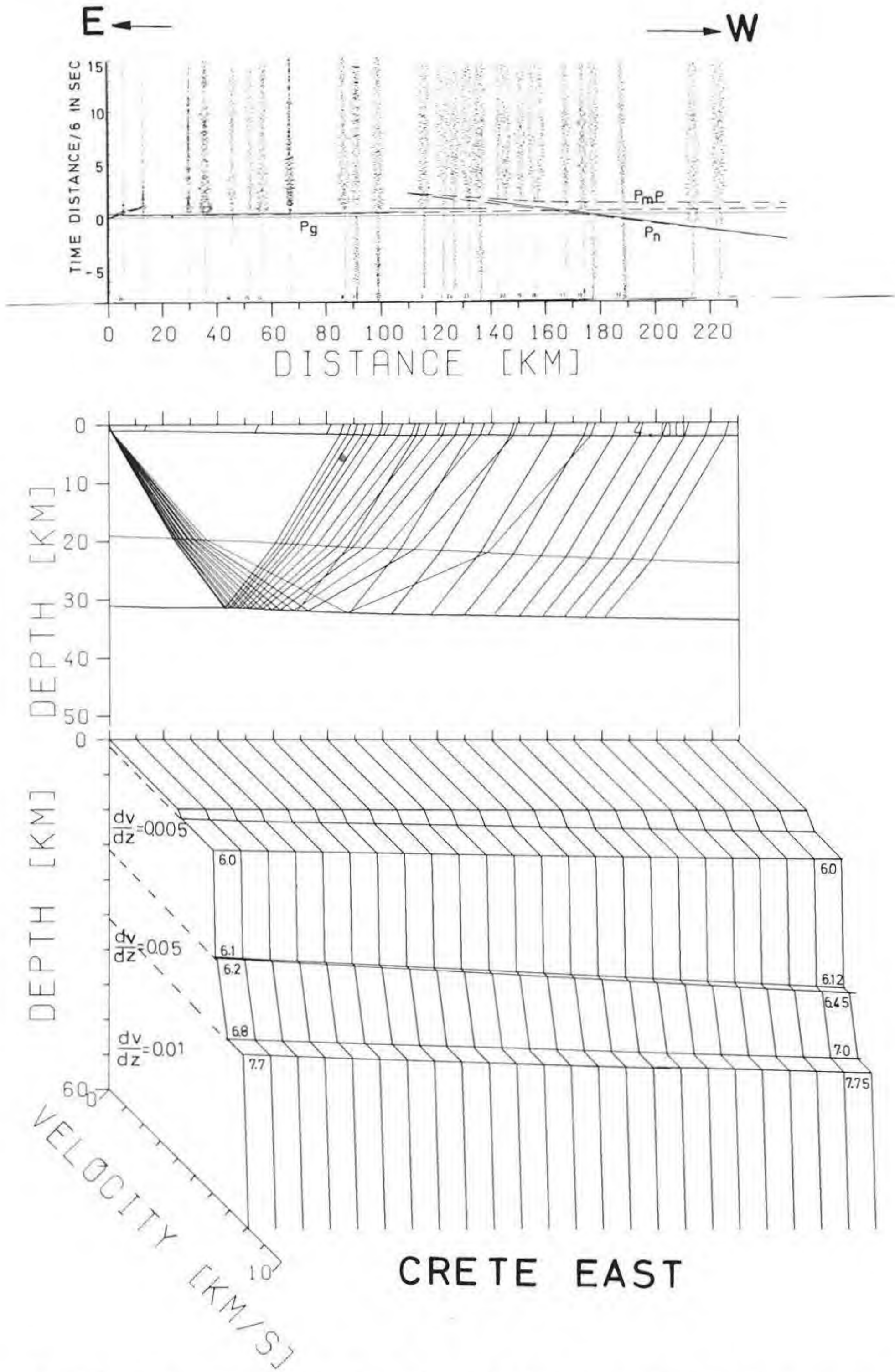



**Fig. 7.** Crete-section, observed E-W. The $P_g$-$P_n$ and $P_mP$-phases are satisfied by a gradient model as demonstrated at Figure 2 for the Evia-section. The gradient of the $P$-wave velocity at the lower crust is $dv/dz = 0.05$. The meaning of the symbols used is the same as in Figure 2

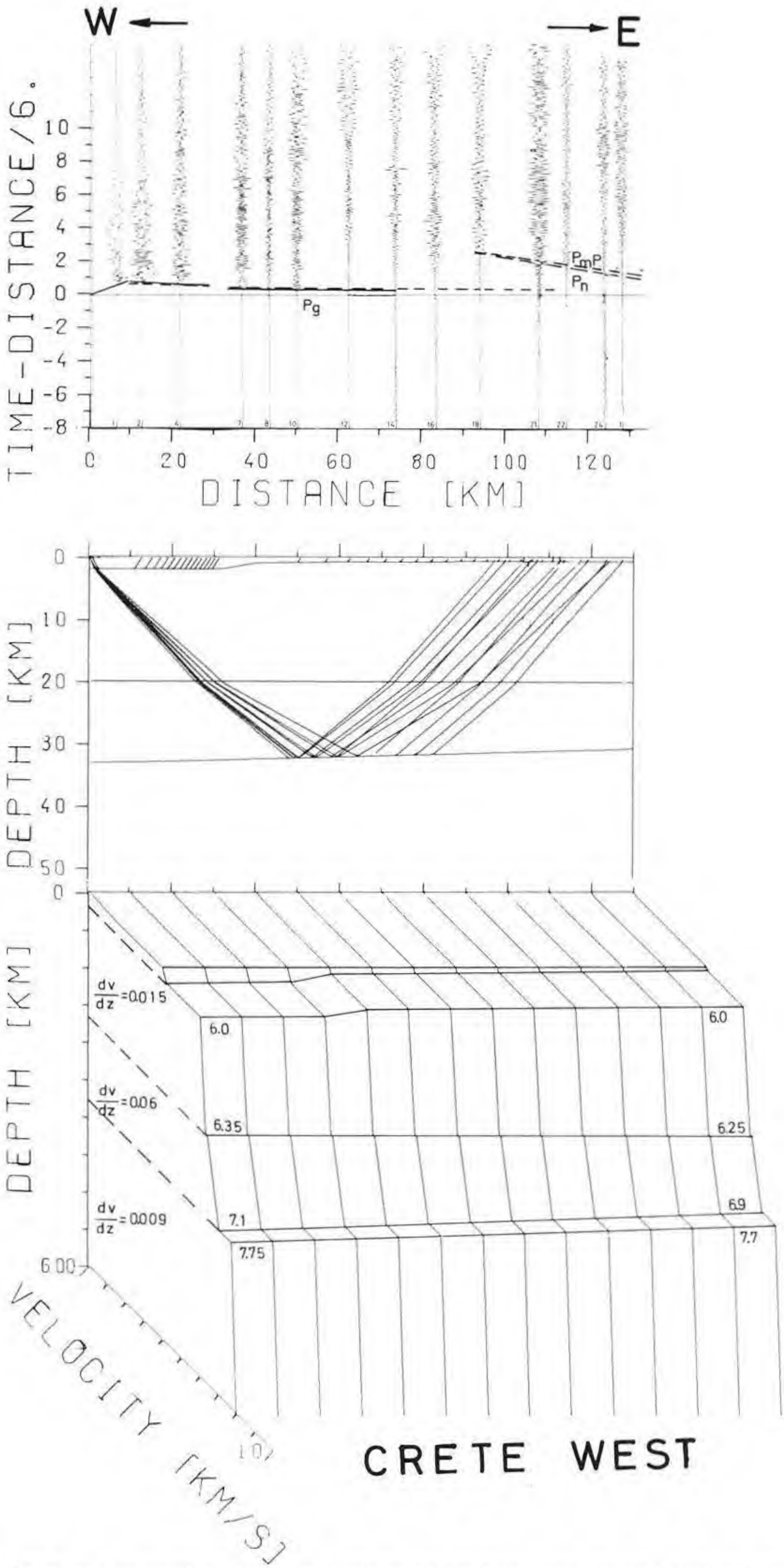



**Fig. 8.** The reversed profile to the section "Crete-East" was for technical reasons observed up to 120 km only. The significance of this section is therefore very limited

method were tested on a simple analogue model by classical evaluation methods and gave very good results.

The inversion of the simple Evia profile (Fig. 4) satisfied the $P_g$ and $P_n$ seismic arrivals. The computed $P_mP$ reflections have delayed arrivals, however, which indicates that the mean crustal velocity used for the computations is too slow. Since no reflections or refractions from within the crust were recorded, we have introduced an arbitrary velocity increase with depth so that the $P_mP$ reflected arrivals were satisfied. The limiting constraints on the $v(z)$ introduced are that the velocity at the crust-mantle boundary does not exceed 7.7 km/s and that its mean value for the crust is 6.21 km/s. Furthermore no 1st order discontinuities are permitted within the crust. In order to demonstrate the ambiguity involved in the computational process, we have given a model in Figure 4 that satisfies the observed $P_g$, $P_n$ and $P_mP$ by introducing a 1st order discontinuity at the lower part of the crust. In our opinion, the model that best satisfies the observations is that in Figure 2, which shows a velocity gradient of 6.4–6.8 km/s from 20 km depth to the Moho-discontinuity. Similar considerations have led to the model given for the Amorgos-Evia section and that at Crete. The Amorgos-Evia observations clearly demonstrate that the Moho in the Aegean area is up-dipping from north to south. The crust becomes attenuated from North Evia towards Amorgos from 32 km to approximately 28 km. The crustal velocities and thickness can be only attributed to a continental type of crust.

The Cretan crustal type is also continental, with a thickness of 30 km at the eastern part of the island increasing to approximately 34 km at the western part, see Figures 7 and 8.

The sediments are very thin and irregularly distributed, the Messara neogene basin causing a 0.2 s delay of the seismic arrivals (Fig. 7). The western limit of this basin is most probably fixed by a tectonically weak zone. In Figure 8 the traveltime irregularities observed are due to the uneven distribution of the nappes which have a maximum thickness at the western part of the island.

In both profiles the $P_g$ true velocity is 6 km/s and the crossover distance of the $P_g$-$P_n$ seismic phases is approximately 170 km.

The true velocity of the upper mantle is not higher than 7.7 km/s. It could not be obtained from the shot fired at the western part of the island, since this profile was only 120 km long. For technical reasons it was not possible to extend it any further.

## Summary and Conclusions

Deep seismic soundings along two sections in the Aegean area, Greece, have shown that the crust is of the continental type. Along the Amorgos-Mikonos-Evia line, the crust shows an up-dipping from 32 km at Mandouthi, North Evia, to 26 km at Amorgos, in the South Aegean Sea. The true velocities computed from the observations are: $V_{Pg}=6.0$ km/s for the crystalline basement and $V_{Pn}=7.7$ km/s for the upper mantle. The crust has a mean velocity of 6.21 km/s, which was computed from $P_mP$ wide-angle reflections. The sedimentary cover above the crystalline basement, which outcrops at South Evia and the Cycladic Islands, could not be determined, since it is too thin to be detected (perhaps 1–1.5 km) by the separation distance of our seismic stations. The best model computed requires a gradual velocity increase at the lower part of the crust, from 6.4–6.8 km/s.

The E-W observations at Crete showed results similar to those at the Evia-Amorgos section. The crust is also continental, increasing in thickness in the E-W direction from 30–34 km. The western part of the island, beginning at the western border of the Messara basin, is down-dipping to the west, where the nappes have their maximum thickness. The true velocity of the basement was obtained by the observations of both profiles and gave a true velocity of 6.0 km/s.

In order to explain the observed travel times of the $P_mP$ reflections a similar gradual velocity increase at the lower part of the crust to that of the Evia-Amorgos section was introduced.

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