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Northwestern Turkey Earthquakes and the Crustal Structure Inferred from Surface Waves Observed in Western Greece

by Oldřich Novotný, Jiří Zahradník, and G-Akis Tselentis

Abstract Records of several earthquakes occurring in Turkey in 1999 obtained at broadband seismic stations in western Greece have been used to study the dispersion of surface waves, mainly Love waves. The observed group-velocity dispersion curves have been inverted into horizontally layered models of the Earth's crust by a modified method of the single-parameter variation. As compared with a previous model for the territory of Greece, the dispersion data require significantly lower velocities in the uppermost crust and smaller crustal thickness. In particular, the resulting model displays *S*-wave velocities between 1.3 and 2.4 km/sec in the upper 2 km and a crustal thickness of about 33 km.

Introduction

Broadband seismic systems consisting of Guralp CMG-3T velocity sensors have operated in the western part of the Corinth Gulf, Greece, since November 1997. These seismic stations were used to study the focal mechanisms of local earthquakes and the damaging Athens earthquake of 1999 (Tselentis and Zahradník, 2000; Zahradník *et al.,* 2001). These stations also recorded prominent surface waves with well-developed dispersion for a series of strong earthquakes in northwestern Turkey in 1999. The interpretation of these dispersion data is the subject of this article.

Observed Group Velocities

The parameters of the three earthquakes in Turkey used in the present study, according to the U.S. Geological Survey National Earthquake Information Center (USGS NEIC) bulletins, are listed in Table 1. We used the records from seismic stations Sergoula (SER, 38.413° N, 22.057° E) and Clauss (CLA, 38.198° N, 21.772° E), located in the Corinth Gulf, western Greece (Fig. 1). The distance between these stations was 34.5 km. Station SER recorded all three earthquakes, but station CLA recorded only the 17 August 1999 event. The seismograms of the mainshock of 17 August 1999, recorded at SER, are shown in Figure 2. Love waves were studied from the records of transverse (T) components and Rayleigh waves from vertical (Z) components. Rayleigh waves from radial (R) components were used to check the results from vertical components.

Several peaks and troughs of surface waves for the 17 August event exceeded the dynamic range (Fig. 2). Such clipped records cannot be processed by spectral methods, but the standard graphical method can be applied (Båth, 1979; Kocaoglu and Long, 1993). Therefore, a modification of the peak and trough technique based on smoothed travel times of zero crossings was applied, for consistency, in processing all seismograms. Group velocities, U, were estimated by the formula

$$U(T) = \Delta/(t + \Delta t - t_0). \tag{1}$$

where T is the instantaneous period measured on the seismogram at time t, Δ is the epicentral distance, t_0 is the origin time, and $\Delta t(T)$ is the instrumental time shift. The group velocity determined by equation (1) characterizes an average structure of the crust and upper mantle between the source and receiver.

A harmonic motion recorded by a CMG-3T seismograph precedes the true ground displacement (for periods greater than about 0.03 sec). The true arrival time of a wave with period T can be obtained from the seismogram by adding the time shift $\Delta t = \gamma T/(2\pi)$, where γ is the phase of the transfer function between the ground-motion displacement and the velocity record. For example, the time shifts equal 2.7 and 9.2 sec for periods of 10 and 30 sec, respectively. Compared with the shortest travel times (of about 200 sec) these time shifts cannot be neglected; see an analogous discussion by Bonner and Herrin (1999).

The group velocities of fundamental modes, determined from the dominant wave groups, are shown in Figures 3 and

Table 1
Earthquake Source Parameters

Date	Origin Time	Latitude (N)	Longitude (E)	Depth (km)	*M*
08/17/99	00:01:39.1	40.75	29.86	17	7.8
08/19/99	15:17:45.0	40.62	29.14	10	5.2
08/31/99	08:10:49.5	40.71	29.95	10	5.2

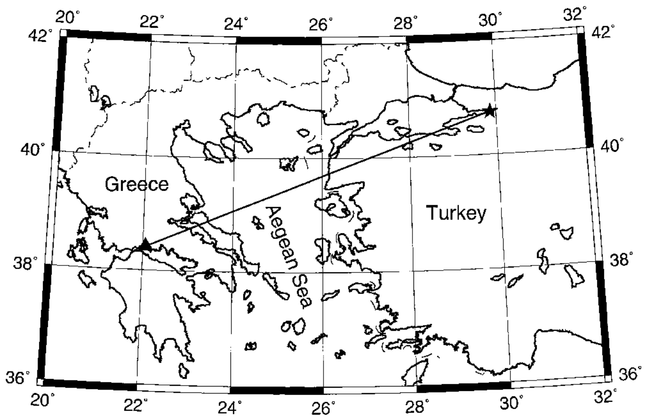

Figure 1. The epicenter (star) of the mainshock of the Turkey earthquake of 17 August 1999 and seismic station SER (triangle).

4 by isolated points. The points lie in bands whose widths are less than about 0.2 km/sec.

Initial Models

A simple model M1, consisting of three layers on a half-space, has frequently been used for locating earthquakes in western Greece (Tselentis *et al.,* 1996). Its *P*-wave velocities and layer thicknesses are given in Table 2. As for the *S*-wave velocities and densities, we assume the following relations (Melis *et al.,* 1989; Červený *et al.,* 1977, respectively):

$$\beta_i = \alpha_i/1.78, \; \rho_i = 1.7 + 0.2\alpha_i, \qquad (2)$$

where α_i is the *P*-wave velocity (in km/sec), β_i is the *S*-wave velocity (km/sec), and ρ_i is the density in the *i*th layer (g/cm^3).

In some focal mechanism studies, low-velocity layers had to be introduced at the top of model M1 in order to explain the observed long duration of seismograms (Zahradník *et al.,* 2001). We shall denote this slightly more complicated model as M2 (Table 2). The *S*-wave velocities and densities are again assumed to satisfy equation (2). Models M1 and M2 were used as initial structural models in our surface-wave dispersion analysis. However, it is to emphasize that only the M1 model has been used routinely in western Greece.

Interpretation of the Observed Data

The theoretical phase velocities were computed for horizontally layered models by the matrix methods described by Proskuryakova *et al.* (1981). The methods represent real-valued modifications of the Thomson-Haskell matrices for Love waves and of delta matrices for Rayleigh waves (Haskell, 1953; Watson, 1970). For computing group velocities we used analytical formulas based on the application of the implicit function theorem to the matrix form of the dispersion function (Novotný, 1970); for the analogous formulas, see also articles by Harkrider (1970) and Schwab and Knopoff (1972).

The theoretical group-velocity dispersion curves for the initial models M1 and M2 are shown in Figures 3 and 4 by solid lines. It can be seen that, at short periods, the observed velocity values are located systematically below the theoretical curves M1. The theoretical dispersion curves for the initial model M2 better satisfy the observed data. To further improve the agreement between the theoretical curves and observed data, we calculated new models using a modification of the single-parameter variation method.

In the usual version of the single-parameter variation, we vary the first parameter until the minimum deviation is achieved, then we proceed to the second parameter, and so

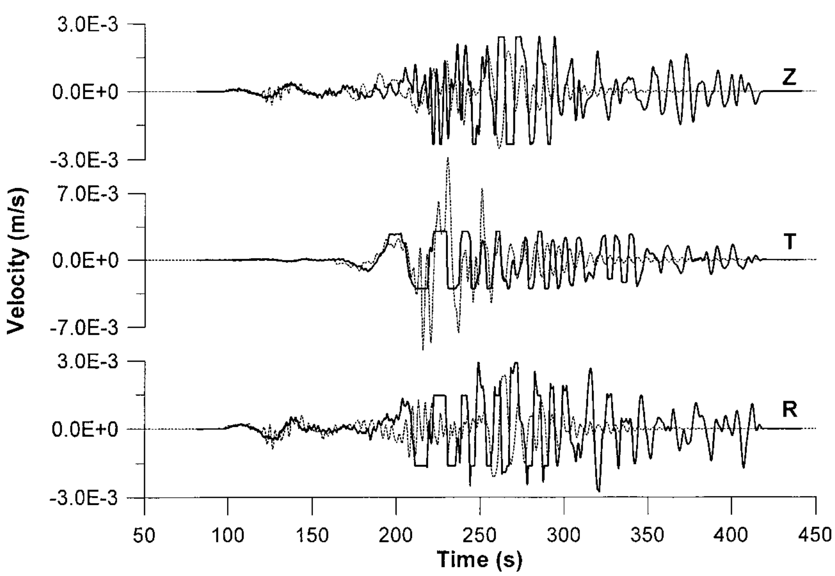

Figure 2. The mainshock of 17 August recorded at the SER station. The observed (thick) and synthetic (thin) seismograms are compared. Unequal scaling is used on the vertical axes. Zero of the horizontal axis corresponds to the origin time. The transverse (T) and radial (R) components were obtained by rotating the NS and EW components according to the geometrical azimuth.

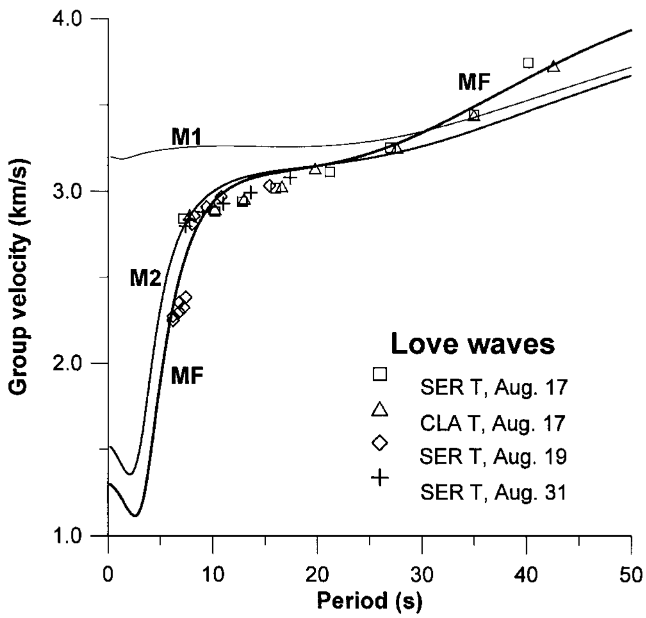

Figure 3. Dispersion curves of Love-wave group velocities. The isolated points represent the observed values, the lines are the theoretical dispersion curves for the initial models M1 and M2, and for the final model MF.

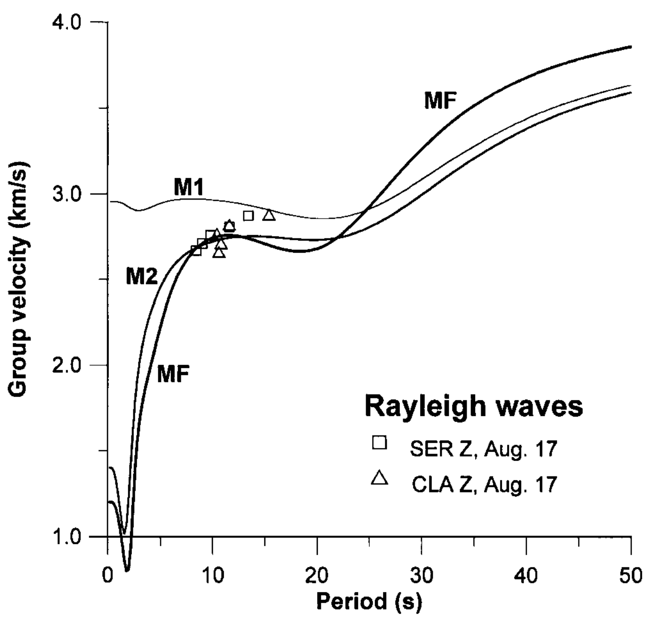

Figure 4. Dispersion curves of Rayleigh-wave group velocities. Denotation is similar to Figure 3.

Table 2
Parameters of the Initial Models M1, M2, and of the Final Model MF

	M1			M2			MF		
i	α_i	β_i	d_i	α_i	β_i	d_i	α_i	β_i	d_i
1	5.70	3.20	5	2.70	1.52	1	2.31	1.30	1
2	6.00	3.37	13	4.50	2.53	1	4.27	2.40	1
3	6.40	3.60	21	5.70	3.20	3	5.52	3.10	3
4	7.90	4.44	∞	6.00	3.37	13	6.23	3.50	11
5				6.40	3.60	21	6.41	3.60	17
6				7.90	4.44	∞	8.37	4.70	∞

α_i, *P*-wave velocity (km/sec); β_i, *S*-wave velocity (km/sec); d_i, the thickness (km) of the *i*th layer.

forth. However, the final model obtained in this way can differ significantly from the initial one. In order to emphasize the importance of the initial model we suggest the following modification (Holub and Novotný, 1997). In a loop from the first to last parameter, we allow a given parameter p to be changed only by one step, Δp; more precisely, we select from values $p - \Delta p$, p, and $p + \Delta p$ the value that yields the least deviation, while keeping all the other parameters fixed. Then we immediately proceed to the next parameter. After arriving at the last parameter we return to the beginning of the loop and continue in the one-step changes. Hence, our inverse method can be characterized as an iterative one-step modification of the single-parameter variation.

It is well known that surface-wave dispersion is controlled predominantly by the distribution of *S*-wave velocities, but less by *P*-wave velocities and densities (*P*-wave velocities have no effect on Love waves) (Brune and Dorman, 1963). Consequently, in order to simplify the interpretation, we considered the *S*-wave velocities and layer thicknesses to be the only independent parameters of the medium subjected to the search. In particular, we used identical velocity steps (e.g., 0.1 km/sec) and thickness steps (e.g., 1 km) in all layers. The *P*-wave velocities and densities were assumed to satisfy equation (2). The interpretation was performed in several modes, for example, by varying the velocities and thicknesses simultaneously, by keeping the layer thicknesses fixed during several first velocity iterations, by splitting of some layers, by using norm L_1 instead of L_2, or by selecting subsets of the input dispersion data (e.g., Love waves only). The main features of the resulting models remained unchanged, thus indicating enough robustness of the inversion.

The best-fitting final model, denoted as MF, is given in Table 2 and Figure 5. In comparison with model M1, models M2 and MF reduce the misfit in the L_2 norm to about 40% and 25%, respectively. Several other models that fit the data reasonably well (misfit not worse than by 6% compared with model MF) are also shown in Figure 5. As opposed to the initial model M1, all the resulting models display rather low velocities in the uppermost crust (i.e., *S*-wave velocity between 1.3 and 2.5 km/sec in the upper 2 km), and a thinner crust (the crustal thickness between 31 and 33 km). The best-fitting model MF is very close to model M2 in the uppermost crust; from the limited dispersion data at short periods we are not able to distinguish between these two models. At long periods, a few available points of the dispersion curve indicate that model MF should be preferred. New measurements are needed to resolve these problems. Anyway, the

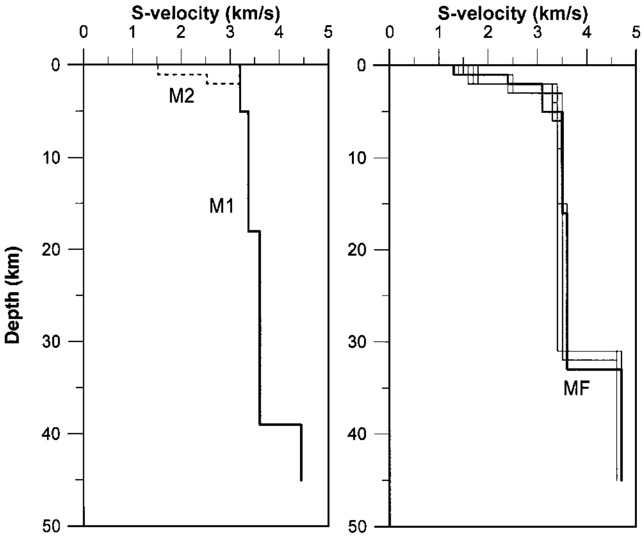

Figure 5. The initial crustal models M1 and M2 (left), and the resulting models of this study (right). The best-fitting model MF is shown by the heavy line. The other models (thin lines), also fitting the data reasonably well, are shown to demonstrate the uncertainty of the solution.

most important practical result is that both models M2 and MF are significantly better than model M1, the model currently used for earthquake locations.

To partially validate the resulting crustal model MF we compared the observed and synthetic seismograms for the mainshock of 17 August. According to the USGS moment-tensor solution, the corresponding source parameters were as follows: origin time 00:01:38.56 UTC, latitude = 40.639° N, longitude = 29.830° E, depth = 17 km, the best double-couple moment = 1.4×10^{20} N m, strike = 92°, dip = 75°, rake = 178°. As for the source–time function, we assumed a trapezium of the 20-sec duration. The synthetic seismograms for station SER, at the epicentral distance of 712.2 km, were computed by the discrete wavenumber method (Bouchon, 1981). As seen from Figure 2, the best agreement was obtained for the transverse component. This corresponds to the fact that the observed group velocities of Love waves were available in a much broader period range than those for Rayleigh waves (cf. Figs. 3 and 4) . Consequently, although the resulting model MF was derived by joint inversion of both Love and Rayleigh waves, it was more constrained by Love waves. As a consequence of the low-velocity subsurface layers in model MF, the duration of the dominant wave group is fitted reasonably well. The dispersive long-period *P*-wave group, although not used in the data inversion, is fitted too.

Conclusions

The dispersion of Love and Rayleigh waves along profiles from northwestern Turkey to western Greece has been investigated. The resulting structural model MF, constrained predominantly by Love waves, displays rather low velocities in the uppermost crust (i.e., *S*-wave velocity between 1.3 and 2.4 km/sec in the upper 2 km), and the crustal thickness of 33 km. The best-fitting model MF is very close to one of the initial models (M2), and the available data prefer model MF only formally. The most important practical result is that both models M2 and MF are significantly better than model M1, the model currently used for earthquake locations. We hope that models M2 and/or MF will improve regional waveform modeling and earthquake source inversions.

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Department of Geophysics
Faculty of Mathematics and Physics
Charles University
V Holešovičkách 2, 180 00 Prague 8, Czech Republic
(O.N., J.Z.)

Seismological Laboratory
Section of Applied Geophysics and Geology
University of Patras
Rio 261 10, Greece
(G.T.)

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