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# SEISMIC WAVE PROPAGATION AND CRUSTAL STRUCTURE IN WESTERN CORINTH GULF (GREECE)

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## ABSTRACT

The velocity structure in the western Corinth Gulf (Greece) is investigated using earthquakes located during the July-August 1991 seismological experiment. The events are recorded by a network consisting of 60 one or three component digital stations deployed approximately at 5km intervals. Interpretation of the data, including time-term analysis and ray tracing identified $P_g$, $P_m$, $P_n$ and intracrustal reflections. The length of the selected profiles is less than 80km and intersects several major tectonic and geological units. Several profiles were constructed in various azimuthal directions, in order to study the seismic wave propagation (identification of direct, intracrustal reflected and refracted phases) and the existence of lateral heterogeneities.

The interpretation of the data suggests the COR91 tectonic velocity model. The intractrustal reflections are best explained by a layer of 1km thick at the top of the model with velocities varying between 4.8 and 4.95 kmsec$^{-1}$ and by a layer of 6km thick at a depth of 26km with a low velocity of 6kmsec$^{-1}$. The body wave velocities increase discontinuously at 4 and 12km. This depth variation consists the seismogenic layer as show studies of local seismicity. The upper mantle velocity of 7.6kmsec$^{-1}$ is reached at a depth of 32km. The upper layers of the proposed velocity model are characterized by high velocities, the lowest layer by a low velocity zone while the $V_p/V_s$ ratio is estimated 1.84.

## INTRODUCTION

The gulf of Corinth, is a WNW-ESE trending, 120 km long and 25 km wide active asymmetric graben, with a history of repeated large earthquakes and a high level of background seismicity (Makropoulos and Burton, 1981). The graben is surrounded by faults, the larger of them located along the southern shore of the gulf dipping towards the north while the northern shore is a southward faulted flexure (Keraudren and Sorel, 1987), which affects Quaternary and recent formations, activated since lower Pliocene time (Mercier et al., 1976; 1989). Tectonic observations across the Gulf indicate normal faults trending approximentely E-W with N-S extension (King et al., 1985; Melis et al., 1989; Rigo et al., 1993). In contrast west of the Gulf, in Kefallinia-Zakynthos area, seismic profiling surveys show large-scale compressional regime in NW-SE direction (Makris 1978; Monopolis and Bruneton 1982; Ferentinos et al., 1985). During the summer 1991, a seismological experiment was carried out in the western Corinth gulf, with the aim to record the local seismicity. The dense network of 60 digital stations, covered an area of about 50x50km. The seismographs were equipped by one or three component short period seismometers. The average, in land, distance between the stations was less than 5km. During the experiment, an important seismic activity was recorded and located with an accuracy better than one kilometer. The high quality of the recorded seismograms, the precision of the source parameters determination and the spatial distribution of the network, allowed the use of seismic profiles, in order to study properties and composition of the crust, as a basis of understanding its nature, dynamics and evolution.

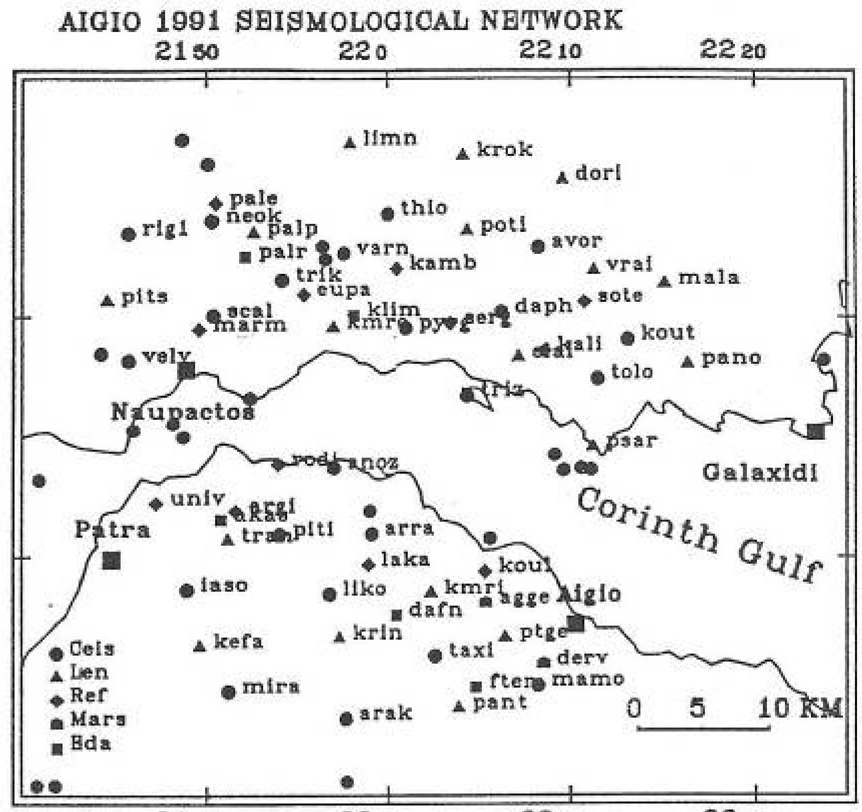


Figure 1. Map of the temporary network.

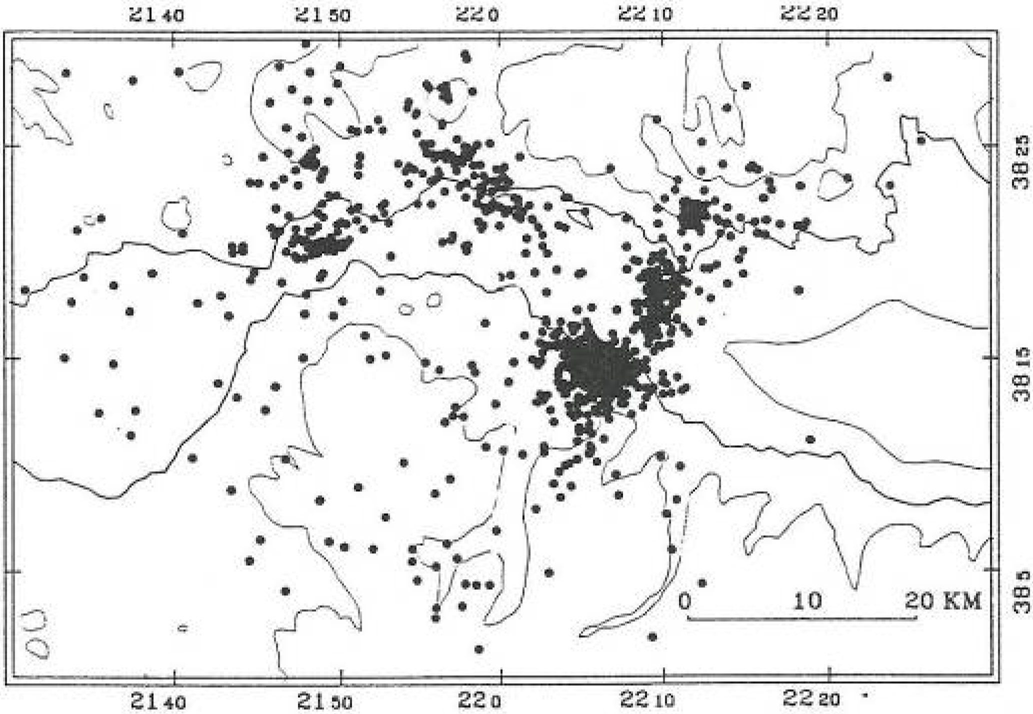


Figure 2. Seismicity map of 1200 microearthquakes located during July 1991.

In July-August 1991, a seismological network was installed in the western Corinth gulf with the collaboration of the Universities of Athens, Paris and Grenoble, in order to study the local seismicity. The network, presented in figure 1, consists of 60 digital stations of different type. Black circles represent one vertical component instruments (1hz seismometers) recording at 100 samples/sec. The other symbols represent three-component instruments (Lennartz, Reftek and Eda) equipped with seismometers varying from 1 to 5 seconds recording at 60 - 200 samples/sec. The locations of the studied earthquakes are represented by solid circles. The mean distance, in land, between the stations is approximately 5km. The geometry of the network allows high precision determination of arrival times and therefore, accurate location of the hypocenters.

During the experiment, about 5000 microearthquakes were recorded and analyzed (Rigo et al., 1995; Papadimitriou et al., 1995). Figure 2 shows the 1200 located events, recorded during July 1991 with uncertainties less than 5km. From these events 600 are located with uncertainties less than one kilometer. The located seismicity covers approximately the whole area, but it is mainly concentrated around Aigio and Naupaktos with a depth distribution varying from surface to 12km. An important seismicity is also located between 12 and 30km. The location of the events is established by a $V_p/V_s$ ratio equal to 1.84, which is determined by Chatelain's method (fig. 3a), and a velocity model determined by minimizing the mean rms (fig. 3b), consisting of three discontinuities at 4, 12 and 30km.

In this study we attempt an approach to the determinetion of the velocity structure in the area, using seismic profiles of events located at different sites and at various depths. The tasks of this study are the determination of a P and S velocity model, the calculation of the $V_p/V_s$ and Poisson ratio and the detection of lateral heterogeneities.

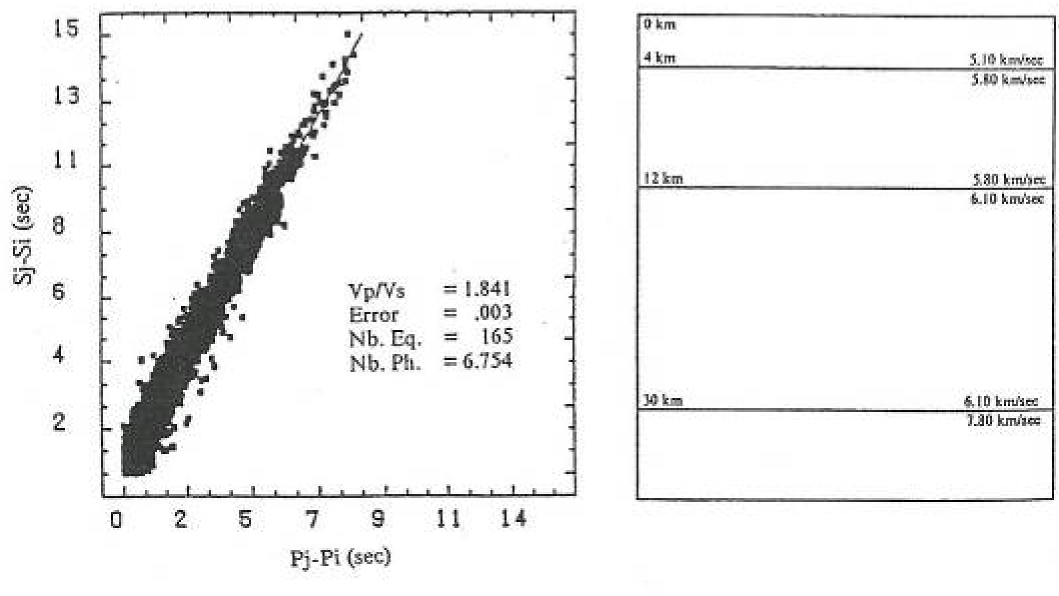


**Figure 3**. (a) $V_p/V_s$ ratio which is determined by Chatelain's method (1978). (b)velocity model, which is determined by minimizing the mean rms.

The events selected fulfill the following criteria: sufficient number of recorded traces; good quality of traces; high precision of the determined location; azimuthal coverage by the network; variety in depth distribution; variety in the range of epicentral distances. The location of the selected events spatially varies, providing ray travelpaths for which the lowest turning point could be associated with upper mantle velocities, allowing the study of the lower crustal structure. Among the studied earthquakes, we present four, located at 3.2km, 4.1km, 16.4km and 26km depth respectively. The analyzed profiles are generally along a NNE-SSW direction, while for a certain number of profiles a reduced velocity of 6 $kmsec^{-1}$ is applied. The method used includes: (a) application of a ray-tracing system in a radially stratified spherical medium to calculate the theoretical travel times; (b) fitting of the theoretical travel times with the first arrival picked phases; (c) construction of seismic profiles along various azimuths and fitting between the observed direct, reflected and refracted phases with the corresponding theoretical travetimes by time-term analysis.

In order to determine the velocity structure, we considered an initial model determined by seismic investigation in Kenya Rift valley (Henry et al., 1990). In this study, the authors proposed three velocity models that reflected the crustal structure of the area We selected the model #2, which is characterized by a low velocity zone.

## UPPER CRUSTAL STRUCTURE

In our analysis we events located at different depths were selected, in order to detect the existance of discontinuities within a range from surface to Moho. In each step, we consider the direct, and the reflected waves on the layers below. In addition, we consider multiple reflections on the uppermost layers.

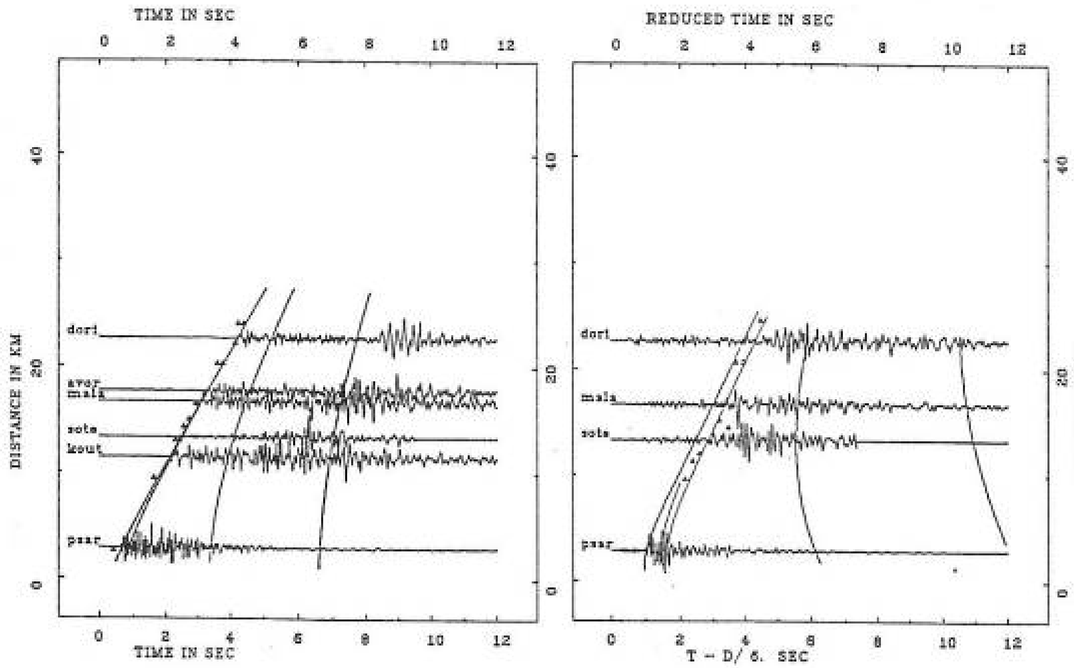


Figure 4. (a) P-wave record section (b) S-wave record section

direction, we observe an agreement of the first P or S theoretical arrival times and the corresponding observed ones. The differences between the two profiles are: (a) the first P arrival phase is impulse for the event located at 4.1km depth, while it is emergent for the event located at 3.2km depth, even though the magnitudes and focal mechanisms of both events are similar (b) the apparent velocity corresponding to the event located at 4.1km depth is higher than the other located at 3.2km depth. In addition, the dispersion of the S waves phases (fig. 5b) is very small. We conclude that an interface exists between 3.2 and 4.1km depth.

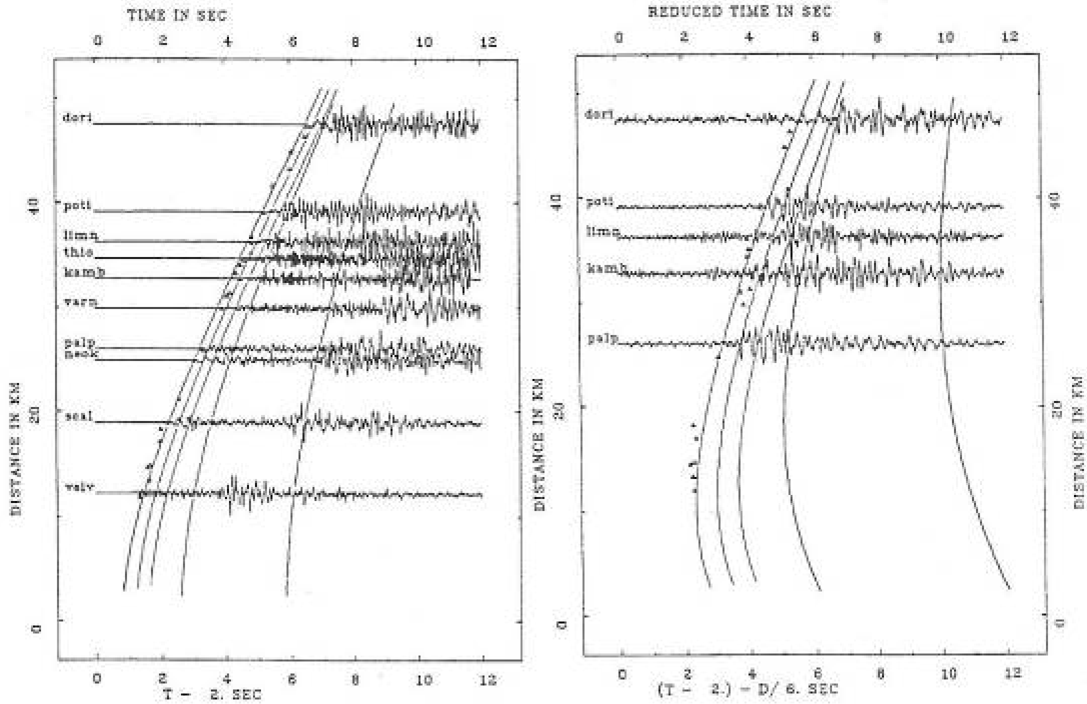


Figure 6. (a) P-wave record section (b) S-wave record section

The third profile concerns an earthquake located at 16.4km depth. We consider the reflections at 22km and 32km depth and also multireflections at 1km depth (a simple and a double). The first P or S arrival theoretical phases fit very well the corresponding observed ones (fig. 6a and 6b). We also remark the good agreement concerning the reflected phases, essentially those produced by the 1km thick determined sedimentary layers, as it can be seen at *neok*, *limn* and *dori* stations.

The last profile concerns an earthquake located at 26km depth. For this case, we considered the reflection produced at 32km depth, the intracrustal reflection produced between the 26-32km discontinuities and the multireflections at 1km depth, as assumed in the previous example. It is important to notice that in this case, the only possibility to fit the observed phases is the interpolation of a low velocity zone between 26km and 32km depth.

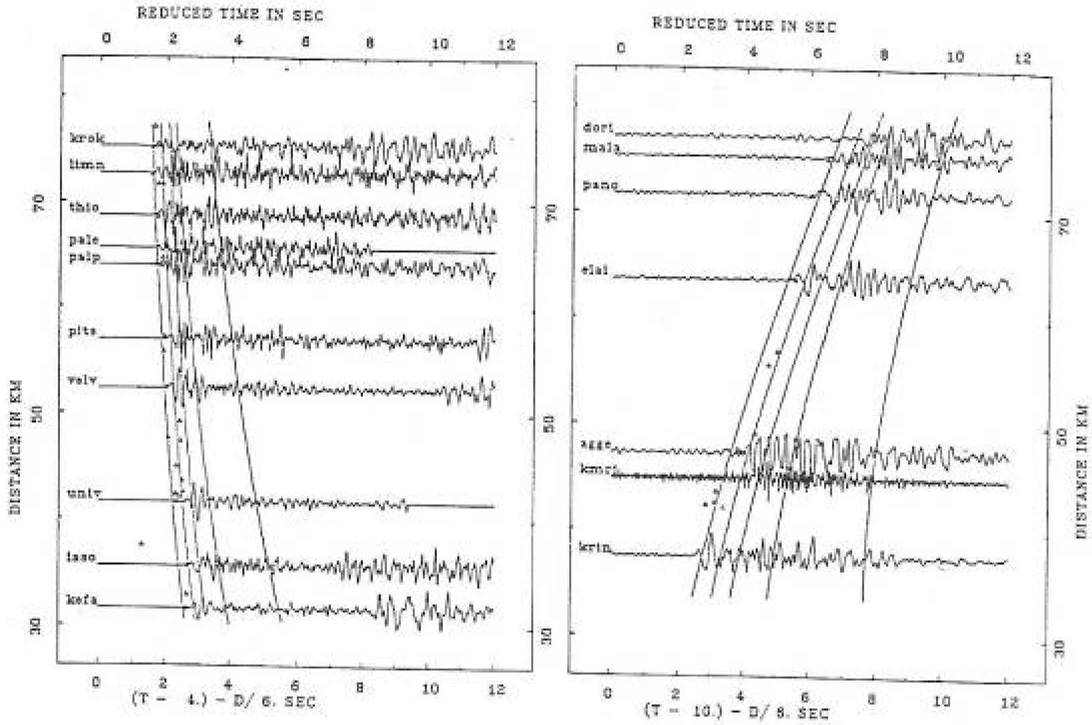

Figure 7. (a) P-wave record section (b) S-wave record section

## DISCUSSION

The body wave recorded profiles were combined with time-term analysis and ray tracing yields to construct the COR91 velocity model presented in figure 7. This model consists of 6 layers: The first layer is situated between 0-1km depth with velocities varying from 4.8 to 4.95kmsec$^{-1}$. The discontinuity situated at 1km depth generates phases observed in several profiles. The second layer is situated between 1-4km, characterized by gradient velocity varying from 5.4 to 5.7kmsec$^{-1}$. The discontinuity located at 4km is well determined by the two profiles presented in figures 2 and 3. This upper crustal structure is characterized by high velocities. The seismic activity recorded during the 1991 experiment, is located beneath approximately after 2km depth (Rigo et al. 1995). The third layer is situated between 4 and 12km, the velocity ranging from 5.9 to 6.1 kmsec$^{-1}$, constituting the seismogenic layer where the major part of the seismic activity of small and large earthquakes is located (P. Papadimitriou et al. 1995). The fourth layer is situated between 12 and 22km, with velocities varying from 6.15 to 6.20kmsec$^{-1}$. The 1000 well located local earthquakes during the 1991 Rio-Antirrio experiment, have mainly occurred above the 12km depth. So, the discontinuity at 12km depth could be interpreted as a brittle-ductile transition zone. The fifth layer is situated between 22-26km with a constant velocity of 6.40kmsec$^{-1}$. This value constitutes the highest velocity within the crust and it is well documented by the profile presented in figures 6a and 6b. The sixth layer is situated between 26-32km with a constant velocity of 6kmsec$^{-1}$. This layer constitutes an important