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Dispersion of Love waves from the 2010 Efpalio earthquake in the Corinth Gulf region, Greece

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Abstract A moderate size earthquake occurred close to the town of Efpalio in the Corinth Gulf region on January 18, 2010. In the present paper we have used this earthquake to study the dispersion of short-period Love waves along nine paths in a broader vicinity of the Corinth Gulf. The observed group velocities require very low *S*-wave velocities in the uppermost 2-km depth. This structural feature differs distinctly from most previous models that were derived from arrival times of body waves in the region. Moreover, it was recognized that the observed group velocities were relatively low for western paths, medium velocities were observed in the central part of the region, and the highest velocities in its eastern part.

Keywords Upper crust · Love waves · Dispersion · Greece · Corinth Gulf

1 Introduction

Greece is characterized by a complex geology formed by a large number of well-defined isopic zones (Fig. 1). These zones were deformed under a recent stress field following the last orogenic phase and new structures, like the Corinth Gulf, emerged. The Corinth Gulf is one of the most active seismic regions of Europe and a very active intra-continental rift.

On January 18, 2010, a moderate size M_w 5.3 earthquake occurred near the town of Efpalio on the northern coast of the western Corinth Gulf (Fig. 1). This main shock was followed by a sequence of weaker earthquakes which lasted almost six months. Data on the geological and tectonic settings of the Efpalio region, its seismicity, focal mechanisms of the Efpalio sequence and examples of seismograms can be found in Sokos et al. (2012). According to that paper, the centroid of the main shock had a latitude of 38.422°N, longitude of 21.941°E, depth of 4.5 km, and origin time of 15:56:09.8. We adopt these values as reference parameters in the present study.

Fig. 1 shows the positions of the selected seismic stations in a broader vicinity of the Corinth Gulf that yielded quality records of surface waves from the Efpalio main shock. The stations lie at epicentral distances between 36 km (station MAM) and 188 km (station AOS). After correcting for instrumental responses, the velocity records were integrated to displacement. The horizontal components were then rotated according to the geometrical azimuths to obtain the radial and transverse components. It was found that the transverse components contained well-developed and dispersive groups of Love waves (thick lines in Fig. 2). As Rayleigh waves on the vertical and radial components were much weaker and more complex, we have restricted our analyses to Love waves only.

To estimate group velocities, we used the time-frequency analysis (Kocaoglu and Long 1993; Kolínský 2004). Each time-frequency diagram was filtered along the 'ridge' of the fundamental mode, and the inverse transform was performed. In this way we obtained filtered seismograms of the fundamental mode which are shown in Fig. 2 with thin lines. In most cases the filtered seismograms approximate the observed ones quite well.

The aim of the present paper is to use the observed Love waves for estimation of the shallow crustal structure in the Corinth Gulf region.

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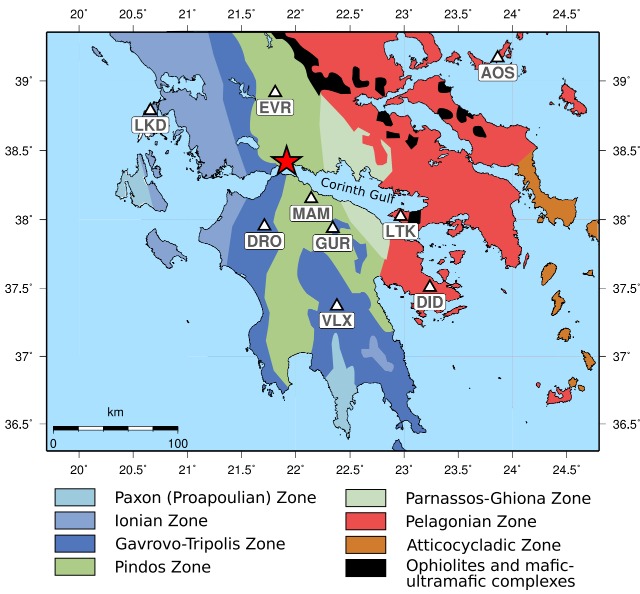

Fig. 1 Geological sketch map of the Corinth Gulf region, modified after Koglin et al. (2009). The triangles show the seismic stations and the asterisk is the epicentre of the main shock

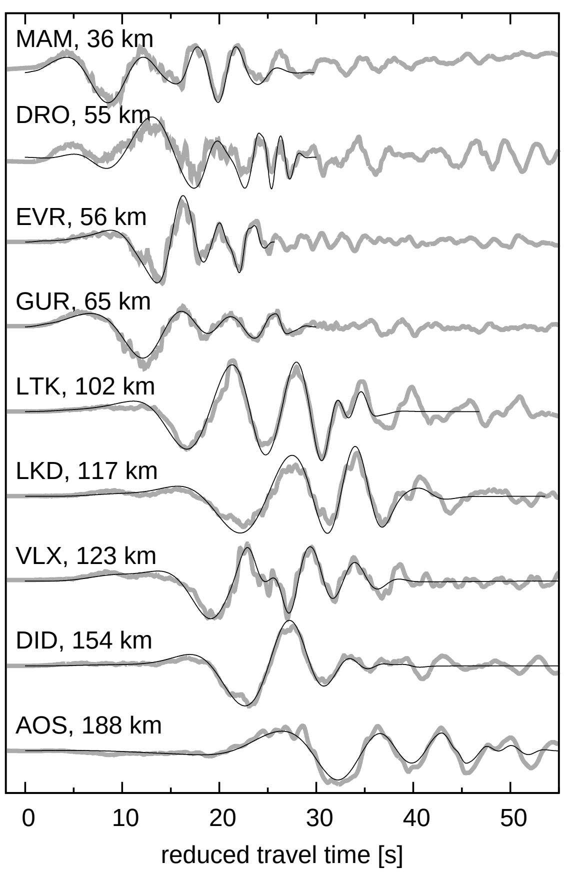

Fig. 2 Seismograms of the transverse components of displacements: observed seismograms (thick lines) and filtered ones to follow the branch of the fundamental mode (thin lines). The travel times were reduced using the reduction velocity of 6.0 km/s. All the seismograms are normalised to their maximum values and are arranged according to the increasing epicentral distance

2 Previous models

Let us mention four previous layered models of the Earth's crust that are relatively simple and represent the main geological structures in the Corinth Gulf region. Their *S*-wave velocity models are shown in Fig. 3a. Rigo et al. (1996) performed a dense seismological experiment on the south-western coast of the Corinth Gulf and derived a *P*-wave velocity model. We denote this model as RI. For determining its *S*-wave velocities we adopted the velocity ratio of $v_P/v_S = 1.81$; see below. Haslinger et al. (1999) measured local seismicity in the area around the Gulf of Arta in the Ionian region (approximately around station LKD). We denote their model as HA.

Novotný et al. (2001) studied the dispersion of surface waves generated by several earthquakes occurring in north-western Turkey in 1999, and recorded at seismic stations in the western part of the Corinth Gulf. These authors observed Love waves in the period range between ~ 7 and 40 s, and Rayleigh waves between 9 and 15 s. Their final model is denoted as MF and its parameters are also given in Table 1. In the present study, we use model MF as the reference and initial model. Our surface-wave data from the Efpalio earthquake extend the observations of Love waves to shorter periods of $\sim$3 s.

Novotný et al. (2012) studied the upper crustal structure in the vicinity of Efpalio using arrival times of the 2010 earthquake sequence. They proposed a model, denoted as LA81, using an initial model from Latorre et al. (2004) with the velocity ratio $v_P/v_S = 1.81$. Model LA81 was derived to a depth of 10 km only, but here we prolonged it beneath this depth by model MF to be able to compute the dispersion curves.

In order to compute the dispersion curves effectively using matrix methods, we have restricted ourselves to horizontally layered models composed of homogeneous and isotropic layers. Moreover, we assume that the *S*-wave velocity and density in each layer are related to *P*-wave velocity as follows (Novotný et al. 2012, 2001)

$$v_P/v_S = 1.81, \; \rho = 1.7 + 0.2 v_P \, , \qquad (1)$$

where v_P is the *P*-wave velocity in km/s, and ρ is the density in g/cm^3. The theoretical phase velocities of Love waves in the layered media were computed by the matrix method described by Proskuryakova et al. (1981), and analytical formulae were used for computing group velocities. The curves in Fig. 3b show the theoretical Love-wave group velocities for the models in Fig. 3a.

3 Observed group velocities

The observed group velocities of the fundamental mode of Love waves are shown in Fig. 3b as points. Each dispersion

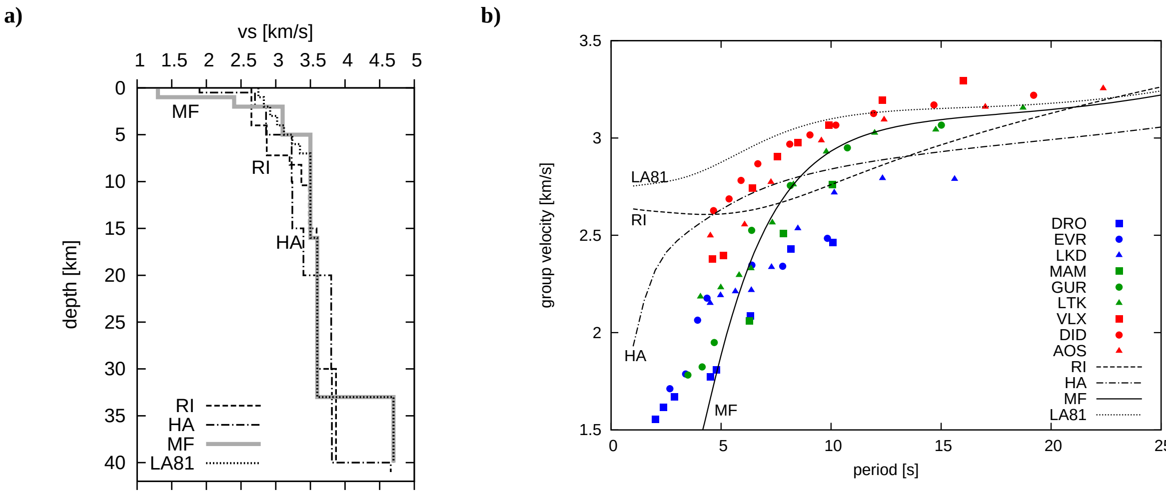

Fig. 3 Previous models and Love-wave dispersion. **a)** *S*-wave depth-velocity functions for some previous models: RI – Rigo et al. (1996); HA – Haslinger et al. (1999); MF – Novotný et al. (2001); LA81 – Novotný et al. (2012). **b)** Group-velocity dispersion curves. The lines are the theoretical dispersion curves for the previous models. The values observed at western, central and eastern stations are shown as the blue, green and red points, respectively

curve characterizes an average crustal structure between the source and receiver. The observed data cover the periods between $\sim$ 3 and 20 s.

To estimate the accuracy of the group velocities, a modification of the peak and trough technique based on smoothed travel times of zero crossings was also applied to all seismograms (Novotný et al. 2001). The group velocities, determined by this graphical method (not shown here), differed from the values in Fig. 3b by less than 0.1 km/s.

All points in Fig. 3b lie in a band whose width is about 0.5 km/s. Although this width is rather large, some comparisons with previous structural studies can easily be done. First of all, we should mention the striking difference between low values of the observed group velocities at short periods (around 5 s and shorter), and much higher theoretical group velocities for models HA, RI and LA81. Note that these models were derived from many hundreds of body-wave arrival times. Nevertheless, it is evident that the superficial *S*-wave velocities in these models are too high and should be reduced. On the other hand, model MF satisfies the observed dispersion data quite well.

The differences between the velocity models derived from body and surface waves are due to the fact that the focal depths of earthquakes in the Corinth Gulf region are larger than about 5 km. In this case, the detailed shallow structure above this depth cannot be uniquely determined from body-wave arrival times only (Crosson 1976). To overcome this drawback, other data should be added, such as refraction measurements using surface sources, surface-wave dispersion or waveform inversions.

4 Search for new layered models

Although model MF satisfies the observed dispersion rather well, small changes in short periods are required. We calculated new models using a modification of the single-parameter variation method where model MF served as the initial one; see Novotný et al. (2001) for details.

In order to simplify the interpretation, we considered the *S*-wave velocities to be the only independent parameters subjected to the search. In particular, we used identical velocity steps (usually 0.1 km/s). Moreover, the densities were assumed to satisfy relations (1), the layer thicknesses were kept fixed, and norm L_2 was applied. Various numbers of iterations were performed, but 15 at most. The iterative process was usually stopped when the deeper layers of the model started to deviate significantly from model MF (as our data contain relatively short periods only, which are insensitive to the parameters of the lower crust).

First, model MA (Table 1 and Fig. 4) was obtained when all data from the nine stations were inverted together. Thus, this model represents an average model of the region under study. Although this model satisfies the observed data better than model MF, the final root-mean-square deviation remained rather large. The result will not improve considerably even if the inversion method is modified (another initial model, variation of velocities and thicknesses, norm L_1 instead of L_2). Therefore, for our data set, model MA is not noticeably better than model MF.

It was then recognized that the large scatter of observed data in Fig. 3b is not random, but has a systematic character, reflecting certain lateral inhomogeneities of the medium. Consequently, we have roughly separated the dispersion curves into curves with low, medium and high group velocities, as follows:

a) Low group velocities along paths to stations LKD, DRO and EVR (blue points in Fig. 3b). These stations are located in the western part of the studied region.
b) Medium group velocities for the central stations MAM, GUR and LTK (green points).
c) High group velocities for the eastern stations AOS, DID and VLX (red points).

We see that, in general, the group velocities increase from west to east, and that certain groups of dispersion curves can be distinguished. Although the division of the observed dispersion curves into three groups was somewhat formal, it can be approximately associated with the main features of the geological structure. Namely, three dominant belts can be distinguished in the geological map of Greece in Fig. 1: the Ionian-Tripolitsa zones (blue colours in the western part of the figure), the Pindos zone (central belt shown in green), and the Pelagonian zone (eastern belt in red).

The division of the observed dispersion curves into three groups made it possible to interpret the individual groups separately. The models obtained after several iterations of the modified method of the single-parameter variation are given in Table 1. We denote these models for western, central and eastern stations as MW, MC and ME, respectively. Their *S*-wave velocity models are also shown in Fig. 4a. These models satisfy the selected data sets much better than the initial model MF and the average model MA.

The main differences between the initial model MF and the new models MW, MC and ME occur in the first layer (Table 1) where the *S*-wave velocity is 1.3 km/s in model MF, but higher velocities of 1.5-1.7 km/s appear in the other models. This structural difference is probably due to the lack of dispersion data with periods below 7 s in deriving model MF, but shorter periods were used in the present study.

In order to estimate the ambiguity of models MA, MW, MC and ME, we applied delete-one jackknifing (Tichelaar and Ruff 1989). In particular, we repeated the inversions omitting one element of the dispersion data (one point in Fig. 4b) each time. The differences did not exceed 0.2 km/s.

5 Conclusions

Novotný et al. (2001) investigated the dispersion of Love and Rayleigh waves along profiles from northwestern Turkey to western Greece in the period range between about 7 and 40 s. Their resulting structural model MF displayed rather low velocities in the uppermost crust (Table 1).

In the present paper, we have extended the studies of Love-wave dispersion to shorter periods of about 3 s using the moderate size earthquake which occurred close to Efpalio on January 18, 2010. We have studied the dispersion along nine profiles in a broader vicinity of the Corinth Gulf to epicentral distances of about 190 km. The best-fitting average model, denoted as MA, is close to model MF.

Table 1 The *S*-wave velocities of the layered models. Model MF is the initial model, MA is the average model of the whole region. The other models characterize the individual subregions: MW – western, MC – central and ME – eastern subregions

Depth	v_S (km/s)				
(km)	MF	MA	MW	MC	ME
0–1	1.3	1.6	1.7	1.5	1.6
1–2	2.4	2.4	2.2	2.3	2.7
2–5	3.1	3.0	2.7	3.0	3.2
5–16	3.5	3.4	3.2	3.5	3.5
16–33	3.6	3.6	3.6	3.6	3.6
33–∞	4.7	4.7	4.7	4.7	4.7

The observation of short-period Love waves made it possible to recognize certain lateral inhomogeneities in the shallow crustal structure. In general, the observed group velocities were relatively low for western profiles, medium velocities were observed in the central part of the Corinth Gulf, and the highest velocities in its eastern part. The respective models are denoted as MW, MC and ME (Table 1). This division of the dispersion curves into three groups approximately corresponds to the geological structure of the region.

The models inferred from dispersion data (Table 1) differ significantly from several previous models that were derived from arrival times of body waves. In particular, the dispersion data require very low superficial *S*-wave velocities ranging between about 1.5 and 1.7 km/s in the upper 2 km. We hope that models in Table 1 will improve regional waveform modeling and earthquake source inversions. However, in order to make these models more precise, surface waves from more earthquakes should be analyzed.

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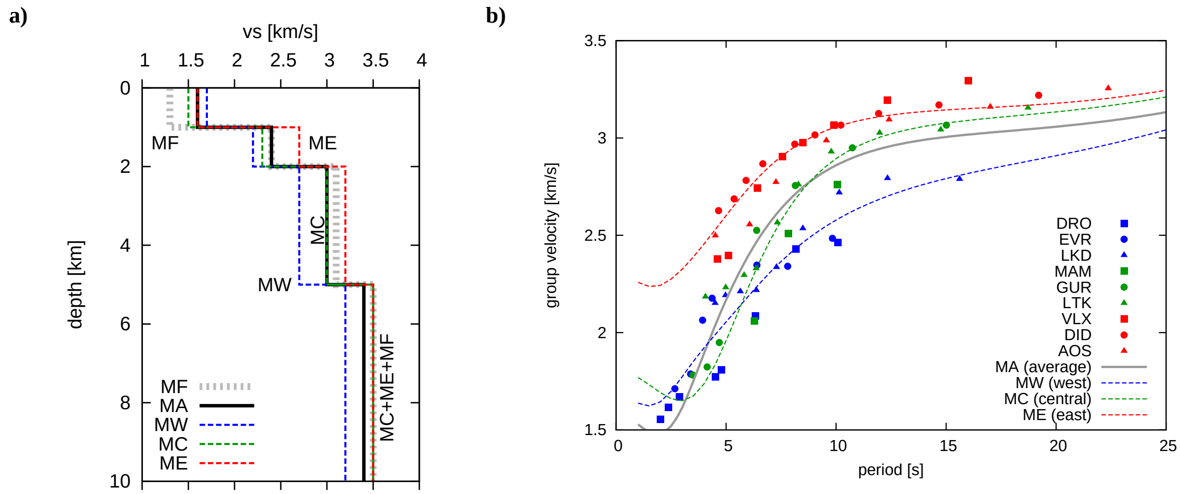

Fig. 4 New models and Love-wave dispersion. **a)** New models for the whole region (MA) and its parts (MW, MC and ME). The colours correspond to those used in Fig. 3b. Model MF is also shown for comparison. **b)** Dispersion curves. The isolated points are repeated from Fig. 3b

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