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Estimation of attenuation structure and local earthquake magnitude based on acceleration records in Greece

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ABSTRACT

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Strong-motion accelerograms are used for the estimation of attenuation coefficient in the upper crust in Greece, on the assumption that direct S-waves are responsible for the strong near-source motion. This coefficient was then used to derive a calibration formula for the estimation of local earthquake magnitude for epicentral distances $\Delta < 130$ km and $4.0 \leq M_L \leq 6.1$. Q_s values are calculated by the spectral decay of the log spectrum of the accelerograms. The resulting values exhibit the high attenuation of seismic waves in Greece and indicate the existence of a highly attenuative surface "sedimentary" layer which strongly controls the total wave attenuation.

Introduction

Local magnitude scale, M_L, although being a peak-response scale (Richter, 1935; Gutenberg and Richter, 1942), hence providing limited information on earthquake's source, is still useful for near-source ground motion studies. The main reason for this is the frequency window that it samples (around 1 Hz), which is of great interest for strong-motion seismology and earthquake engineering. The constant increase of the number of recorded accelerograms extended the estimation of local magnitude to this strong-motion data (Kanamori and Jennings, 1978; Luco, 1983). Although peak ground acceleration has been doubted to correlate particularly well with all types of structural damage (Blume, 1979) and RMS accelerations and spectral amplitudes have been used for strong-motion analysis (McGuire and Hanks, 1980), peak ground acceleration remains the main strong-motion parameter to be considered, especially for M_L estimation (Lee et al., 1990). On the other hand, attempts have been made for the calculation of an "effective peak acceleration" (Hanks and McGuire, 1981).

Spectral analysis of strong motion showed the important role of near-receiver geologic conditions and attenuation structure (Trifunac, 1976; Hanks, 1982). It has been shown (Cormier, 1982; Anderson and Hough, 1984) that the spectral decay of the log spectrum can be used for the calculation of attenuation time t^*, thus enabling one to determine the quality factor and the attenuation structure. Al-Shukri and Mitchell (1990) used this technique for a full 3D Q_s tomography in the New Madrid region in central U.S.A.

The present study is concerned with continental Greece, which belongs to a back-arc area, where the Aegean is a marginal sea with intermediate-depth seismicity on a well-defined Benioff zone (Papazachos and Comninakis, 1970) and especially high shallow seismic activity (Com-

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TABLE 1

Source and strong motion parameters of the earthquakes used in this study and results on earthquake "size" and attenuation structure

No.	Date	Time	Lat. (°N)	Long. (°E)	h	M_L	Recording codes	Δ (km)	a_g^* (cm/s^2)		log A_{f1}	log A_{f2}	t* (m s)	Q_s	M_{La}
									Long.	Trans.					
1	Jun 20, 1978	20:03:24	40.71	23.27	10	5.9	THE78-1	30	159.8	173.4	3.79	4.07	105	85	6.0
							GEV78-1	80	47.4	45.6			77	298	
2	Aug 11, 1980	09:15:59	39.30	22.82	3	4.8	ALM80-4	15	79.7	80.0		3.33	51	87	4.8
3	Sep 26, 1980	04:19:18	39.24	22.74	2	4.5	ALM80-5	7	64.4	67.0		2.89	40	55	4.1
4	Jan 17, 1983	12:41:31	37.96	20.23	6	6.1	AGR83-1	128	35.0	58.5	3.96	4.27	157	233	6.2
							LEF83-1	105	74.1	71.2			203	148	
							ARG83-1	33	141.6	121.7			79	121	
5	Jan 17, 1983	16:53:30	38.11	20.37	8	4.9	ARG83-3	13	69.1	25.8		3.42	117	37	4.5
6	Jan 31, 1983	15:27:02	38.14	20.45	5	5.2	ARG83-6	6	46.3	60.3	3.46	3.26	67	33	4.5
							ZAK83-1	56	19.2	16.8			118	136	
7	Feb 20, 1983	12:42:29	37.76	21.11	2	4.9	ZAK83-2	19	35.9	29.3		3.33	112	49	4.4
8	Mar 16, 1983	21:19:39	38.80	20.88	1	4.9	LEF83-2	16	29.8	18.2		3.27	143	32	4.6
9	Mar 23, 1983	19:04:00	38.78	20.83	2	4.8	LEF83-3	12	58.0	35.4		3.55	171	20	4.3
10	Mar 23, 1983	23:51:07	38.22	20.35	1	5.6	ZAK83-3	68	23.7	29.3	3.43	3.66	105	186	5.6
							LEF83-4	75	24.8	29.0			171	125	
							ARG83-7	13	154.6	204.4			35	115	
11	Aug 6, 1983	15:43:53	40.09	24.78	12	5.9	POL83-1	119	17.7	17.2	3.68	3.78	106	322	5.7
							IER83-2	85	30.9	28.7			67	364	
12	Aug 26, 1983	12:52:10	40.47	23.96	3	4.4	OUR83-2	16	76.7	110.8		3.45	56	82	5.0
13	Feb 19, 1984	03:47:22	40.61	23.40	8	4.2	POL84-1	25	16.5	12.6		2.82	51	204	4.0
14	Oct 4, 1984	10:15:12	37.64	20.85	14	4.5	ZAK84-1	18	40.3	88.1		3.40	56	117	5.0
15	Oct 25, 1984	09:49:16	36.83	21.72	11	4.7	KYP84-2	47	27.9	–	3.55	3.66	70	196	5.5
							PEL84-1	27	146.3	154.1			44	188	
16	Mar 22, 1985	20:37:39	38.98	21.11	2	4.0	AMF85-3	14	55.9	62.9		2.98	43	68	4.2
17	Mar 22, 1985	20:38:54	38.91	21.06	7	4.1	AMF85-4	12	32.7	37.4		2.85	34	116	4.1
18	Nov 9, 1985	23:30:43	41.26	24.02	12	5.0	DRA85-1	17	39.6	98.7	3.32	3.45	49	122	5.1
							KAV85-1	49	43.4	30.1			49	29	
19	Sep 13, 1986	17:24:34	37.11	22.14	8	5.5	KAL86-1	9	269.3	306.7		3.98	94	37	5.6
20	Sep 15, 1986	11:41:30	37.04	22.13	8	4.9	KAL86-7	1	266.6	156.0	3.17	3.46	76	30	4.9
							KAL86-2	1	180.5	295.3			67	34	
							MES86-1	10	35.0	52.8			91	40	
21	Dec 21, 1990	06:57:22	40.93	22.37	10	5.1	EDE90-1	32	114.1	110.0	3.45	3.74	126	76	5.1
							KIL90-1	42	43.2	36.1			70	178	
							ABS90-2	70	14.2	14.7			98	206	

ninakis, 1975; Papazachos, 1990) with mainly normal faulting (McKenzie, 1978). The attenuation structure has not been extensively studied in this area. Hashida et al. (1988) found Q values ranging from 50 to over 1000 using macroseismic data. Using the same kind, but a larger amount of data (Papazachos, 1992), found an average Q value of 350 for the upper crust of this area and gave evidence about the high attenuation of seismic waves. Coda-Q values near 200 at frequencies around 3 Hz were found for the Peloponnese (southern Greece; Martin, 1988) and values around 150 for the Mygdonia basin (northern Greece) and 200 for southern Aegean were also estimated for the same frequency domain (Hatzidimitriou, pers. commun., 1991). This high attenuation is especially prominent in the inner (concave) part of the southern Aegean as shown by O.B.S. (Kovatchev et al., 1991) and other geophysical data (Papazachos and Comninakis, 1971; Tassos, 1984).

In the present study, horizontal peak ground acceleration data were used for the estimation of local magnitude. Local site effects were not considered and peak ground acceleration was attributed to direct S-waves. Anelastic attenuation was also estimated using the peak ground acceleration and the log spectrum of the accelerograms. Quality factor values were calculated and their spatial behaviour was roughly explained by

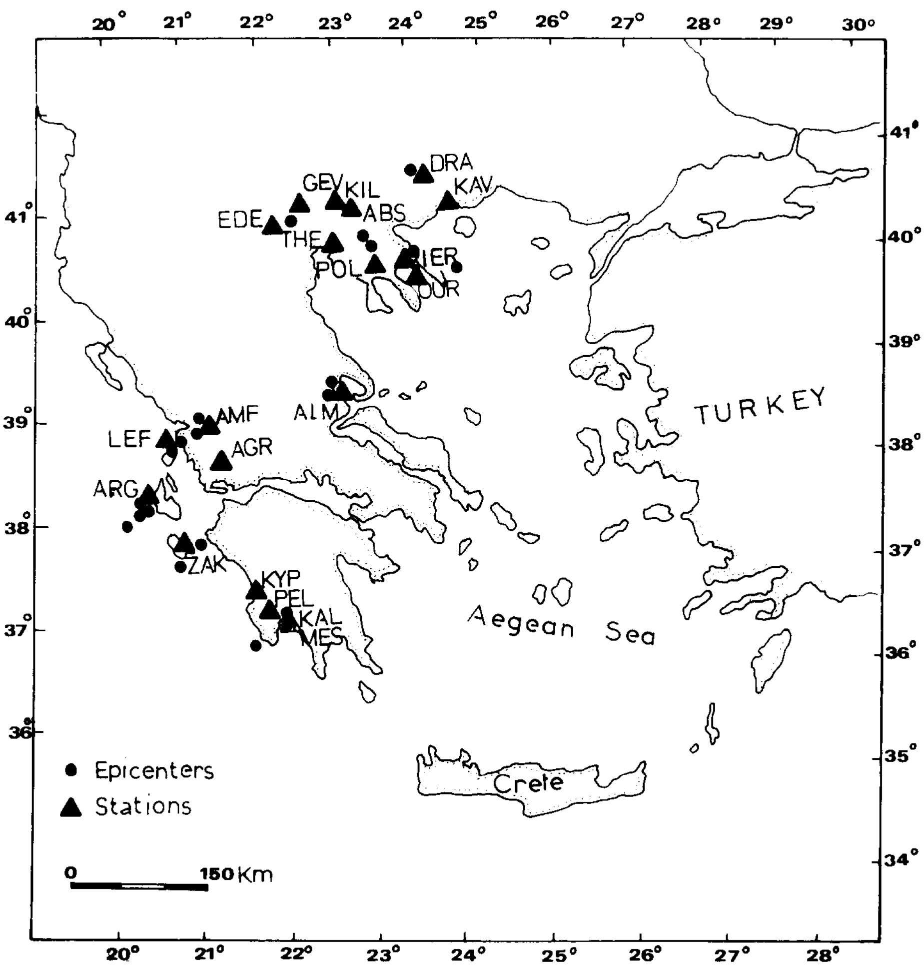

Fig. 1. The epicenters of the earthquakes used in this analysis (circles) and the distribution of the recording stations (triangles).

the existence of a surface, highly attenuative, "sedimentary" layer, which strongly controls the total wave attenuation.

The data

The data set used in this study consists of 67 horizontal components seismograms from 21 shallow earthquakes in Greece with $4.0 \leq M_L \leq 6.1$, recorded at epicentral distances from 1 to 128 km. Information on the parameters of these events is given in Table 1. The first six columns of this table list the date, the origin time, the epicentral coordinates, the depth and the local magnitudes of the causative earthquakes. The next three columns give information on the recording codes, on the distance from station to receiver and on the values of the recorded horizontal peak ground acceleration, a_g^*, reduced to intermediate local geological conditions.

Earthquake parameters for the events numbered 1 to 15 (Table 1) have been taken from Karacostas (1988), for the events numbered 16 to 18 from Papazachos and Drakopoulos (1985), for events 19 and 20 from Papazachos et al. (1988) and for event 21 from Panagiotopoulos (pers. commun., 1991). All the parameters of the accelerograms are taken from Theodulidis (1991). Figure 1 shows the location of the epicenters and the distribution of the recording stations.

All records have been automatically digitized and corrected in ENEA, Italy, by B. Margaris according to the methodology developed in this center (Basili, 1987). The main innovation of this methodology lies in the choice of the high- and low-pass signal filtering, by using the Fourier spectra of the fixed trace and of the uncorrected accelerogram component (Rinaldis, 1985; Margaris et al. 1989).

Reduction of the horizontal peak ground acceleration to "intermediate" local geological conditions, $s = 0.5$, has been performed using the formula proposed for the same data set by Theodulidis (1991):

$$\ln a_g^* = \ln a_g + 0.27s$$

where a_g is the uncorrected peak ground acceleration, a_g^*, is the corrected peak ground acceleration and s characterizes the local site geological conditions, that is $s = 1$ for "rock" and $s = 0$ for "alluvium".

Methods of data analysis and results

In the present study, peak ground acceleration is attributed to direct S-waves. This is a plausible assumption, usually made in similar studies. Dahle et al. (1990) showed that the effect of Lg-waves can be neglected and that only S-waves should be taken into account if we are confined in epicentral distances up to near 100 km. This result was confirmed for Greece from macroseismic data (Papazachos, 1992). In the present study, all the data used have epicentral distances up to approximately 100 km (see Table 1) except for two cases (AGR83-1 and POL83-1) which reach up to 128 km. Moreover, we assume that earthquakes occur in a half-space, that is, in the granitic layer. Therefore, the ray paths are straight lines.

Since peak ground acceleration is attributed to direct S-waves in a half-space, the following formula is supposed to hold:

$$A = A_f R^n \exp(DR) \tag{1}$$

where n is the geometrical spreading factor, D is the anelastic attenuation coefficient, A_f is the peak acceleration near the source (at $R = 1$ km) and R is the ray path length, equal to $(\Delta^2 + h^2)^{1/2}$ for a half-space (Δ = epicentre–site distance, h = focal depth). The anelastic attenuation coefficient is known to be equal to:

$$D = -\frac{\pi f}{Q_s V} \tag{2}$$

In the present paper we consider A_f to be representative of the "size" of the earthquake (Papazachos, 1992). It is obvious that if n and D are known, then A_f can be calculated for each accelerogram. Since we have direct body waves, we accept the theoretical value of -1 for n. The calculation of D is then performed by two different procedures which are explained in the text to follow.

Estimation of D using peak ground acceleration

If we take the logarithm of eqn. (1) we find:

$$\log A + \log R - \log A_f = D \log e R \tag{3}$$

We initially assumed an average value for $D(=-0.003)$ and calculated A_f values for each accelerogram and, therefore, for each earthquake. We then calculated the quantities $X = \log e R$ and $Y = \log A + \log R - \log A_f$ which are linearly dependent (eqn. 3). The best-fit line, $Y = a + DX$, was estimated using a maximum-likelihood method (least-absolute deviations). The resulting value of D was used again in eqn. (1) and the whole procedure was repeated. In this case we used earthquakes with two or more accelerograms because single-recording earthquakes cannot be used to determine their A_f and contribute to the determination of D at the same time.

We reached convergence after fifteen iterations and found the value of -0.006 for D and 0.000 (as expected) for a. The values of X and Y and the corresponding best-fit line are shown in Figure 2. The log A_f values estimated for each earthquake are given in Table 1 under the column log A_{f1}. These values were correlated with M_L as this was calculated from the Wood-Anderson seismometers of the National Observatory of Athens (Kiratzi and Papazachos, 1984). Figure 3 shows this plot and the best-fit least-squares line:

$$\log A_f = 0.497 M_L + 0.81 \qquad (4)$$

The value of $\log A_f = 3.66$ for the October 25 1984 ($M_L = 4.7$) earthquake was not used for the determination of eqn. (4) since it is quite high for such a small earthquake, which may be attributed to errors in the epicentral coordinates due to its small M_L.

Substitution of log A_f from eqn. (4) in eqn. (3) yields:

$$M_L = 2.01 \log(AR) + 0.0052R - 1.63 \qquad (5)$$

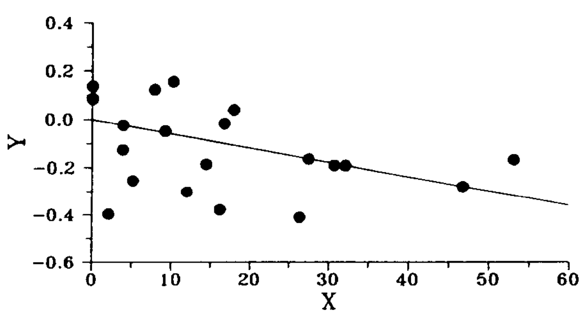

Fig. 2. Plot of $Y = \log(A * R / A_f)$ versus $X = \log e * R$.

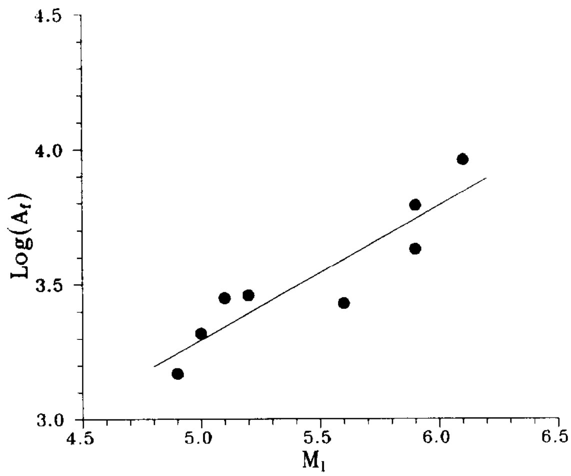

Fig. 3. Plot of the logarithm of source peak acceleration A_f, versus local magnitude M_L, when the anelastic attenuation is calculated from the peak acceleration data.

This is the first relation that we use for the calculation of M_L from accelerograms.

Estimation of D using the spectral-decay method

The procedure used here is that described by Al-Shukri et al. (1988) and Hough et al. (1988). The instrumentally corrected acceleration spectrum observed at a station located at a distance R from the source is defined as:

$$A(R, f) = (2\pi f)^2 S(f) G(R) \, e^{-\pi f t^*} \qquad (6)$$

where $S(f)$ is the source displacement spectrum, $G(R)$ is the geometrical spreading and t^* is the attenuation time defined (Kanamori, 1967) as:

$$t^* = \int_{\text{path}} \frac{dR}{Q_s V_s} \qquad (7)$$

where V_s is the seismic wave velocity, Q_s is the quality factor and the integral is taken over the ray path. It is widely accepted that for frequencies greater than the corner frequency, f_c, the source displacement spectrum, $S(f)$, varies with f^{-2} (Brune, 1970; Madariaga, 1977). In the case that there is no anelastic attenuation ($t^* = 0$), $A(R, f)$ should be frequency independent for $f > f_c$, thus resulting in a flat spectrum. This flat spectrum continues up to f_{max} which is defined by local site conditions and is around 15 Hz for the Aegean area (Theodulidis, 1991). However, if

$t^* \neq 0$, then, if we take the logarithm of eqn. (6), we get:

$$\log A(R, f) = \log A_0(R) - \pi \log e \cdot t^* f \quad (8)$$

where A_0 is frequency independent for $f_c < f < f_{max}$. It is, therefore, expected that for this frequency window, the log spectrum will be linear, with a slope B, equal to:

$$B = -\pi t^* \log e \quad (9)$$

In the present paper we are dealing with a half-space and therefore with straight ray paths. If we assume a uniform value for Q_s and V_s, then t^*, as calculated from eqn. (7), reduces to:

$$t^* = \frac{R}{Q_s V_s} \quad (10)$$

From eqs. (2), (9) and (10) we find:

$$D = \frac{Bf}{R \log e} = \frac{\pi t^* f}{R} \quad (11)$$

Using the B-value we can estimate: (a) The D value for each accelerogram (eqn. 11) and then the corresponding value of log A_f (equation 3), and (b) the attenuation time t^* (eqn. 9) and then the Q_s value corresponding to the ray path (eqn. 10). If sufficient data are available (which is not our case) a 3D Q_s-inversion tomography can be performed as was shown by Al-Shukri and Mitchell (1990). In the present study the dominant frequency of the peak ground acceleration was taken to be equal to 3.5 Hz (Theodulidis, 1991) and the velocity of S-waves equal to that of the granitic layer, that is, 3.5 km/s (Papazachos et al., 1966).

For each horizontal component, the Fourier amplitude spectrum (FAS) of the shear-wave group of arrivals was computed. The initial time of the S-waves window was selected visually and the duration of the window was chosen very carefully in order to include all arrivals which might be direct S-waves since we are dealing with the attenuation of those. In one case (PEL84-1), where only one component was available, we used the FAS of this single component. In Figure 4 we show two typical horizontal-component accelerograms recorded at station LEF (earthquake No. 4) and the corresponding Fourier amplitude spectrum, which was calculated as the root mean square of the individual spectrum of each component. We observe that f_c is around 2 Hz and f_{max}

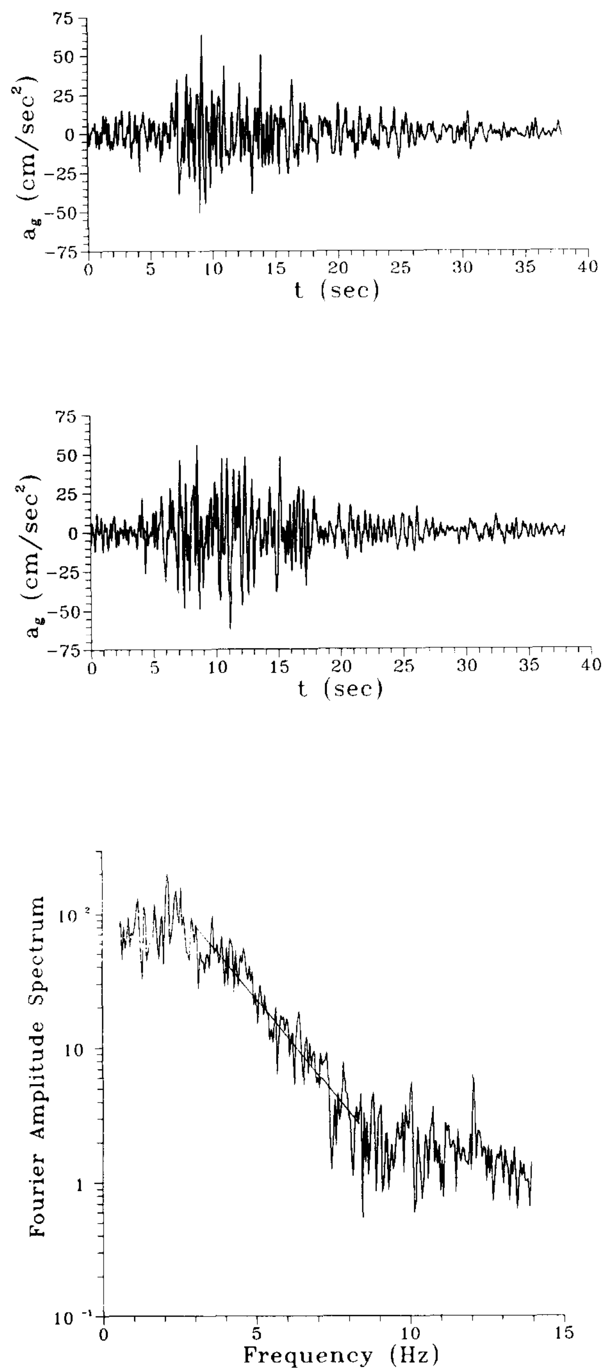

Fig. 4. The two horizontal components of the January 17, 1983 ($M_L = 6.1$) earthquake as recorded in station LEF and the corresponding Fourier Amplitude Spectrum. The least-squares best-fit line determining the slope of spectral decay has been drawn.

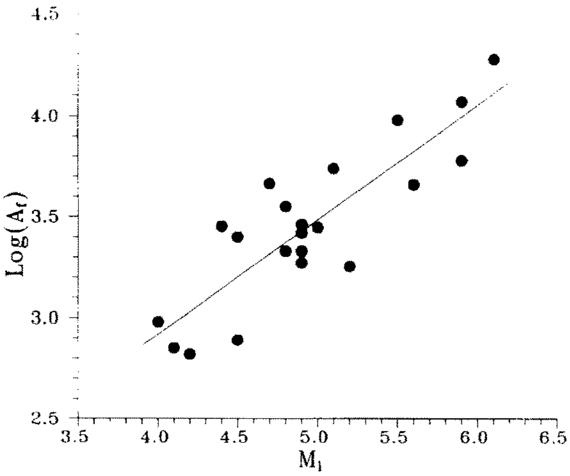

Fig. 5. Plot of the logarithm of source peak acceleration A_f, versus local magnitude M_L, when anelastic attenuation is calculated by the spectral decay of the log spectrum.

around 9–10 Hz. The B value was found by a least-squares fit in the frequency window 3.0 Hz $< f <$ 8.35 Hz.

The calculated t^* values, and the estimated values of log A_f for each earthquake are listed in Table 1 under the columns t^* and log A_{f2}. In Figure 5 the correlation of these values with M_L are presented together with the least-squares best-fit line:

$$\log A_f = 0.567 M_L + 0.652 \tag{12}$$

Figure 6 shows a plot of t^* versus the epicentral distance, Δ. A least-squares line fit to the data resulted in:

$$t^*(msec) = 60 + 0.48\Delta(\text{km}) \tag{13}$$

The slope as well as the intercept values are quite high as compared to values found for other regions (Al-Shukri and Mitchell, 1990). This is indicative of the high attenuation of seismic waves in the broader Aegean area. As Anderson and Hough (1984) pointed out, the intercept of the line corresponds to attenuation due to vertical propagation of seismic waves from the hypocentre to the surface while the slope corresponds to attenuation mainly caused by horizontal propagation of seismic waves.

The four accelerograms recorded at the station LEF resulted in very large values of t^* (near 175 m s) regardless of the epicentral distance. This is propbably due to local high attenuation of seismic waves or local site effects. These values were not incorporated in the estimation of eqn. (13) but the corresponding peak accelerations were used for the estimation of A_f for two earthquakes (May 16 and 23, 1983) where no other recordings existed.

If we substitute t^* from eqns. (13) and (11) and log A_f from eqn. (12) in eqn. (3) and solve for M_L we find:

$$M_L = 1.76 \log(AR) + 0.004\Delta - 0.64 \tag{14}$$

This relation is an alternative one to eqn. (5), but must be considered as more reliable since it was calculated for the whole data set, using a more reliable method for the estimation of D.

Estimation of local magnitude

The relations (5) and (14) were used to calculate the local magnitude M_{Lacc} from the strong-

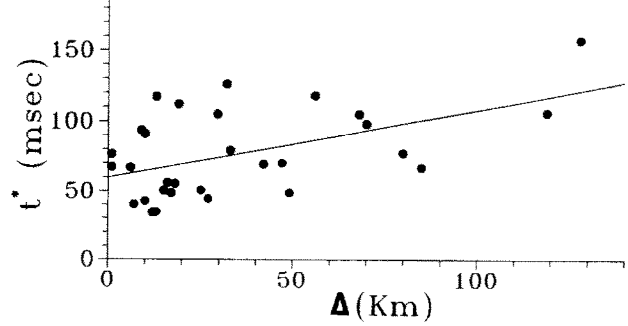

Fig. 6. Attenuation time t^*, as a function of distance Δ.

motion records of the whole data set. These values are presented in Table 1 under the column M_{La}. These magnitudes were compared to the local magnitude, M_{Lwa}, calculated from the records of the Wood-Anderson seismographs of the National Observatory of Athens. The average mean square difference between these local magnitudes was 0.35, a value which is just above the error limits of M_{Lwa} estimation.

In Figure 7 the peak ground acceleration versus distance for an earthquake with $M_L = 6.0$ and $h = 6$ km (equal to the average depth of the data) is presented. The three curves were calculated by eqn. (14), proposed in the present paper, by the relation proposed by Theodulidis and Papazachos (1990) using the same data set and by the relation proposed by Trifunac and Brady (1976) for West U.S.A. One can see that values predicted by eqn. (14) are quite similar to those proposed by Theodulidis and Papazachos (1990) although they followed a completely different method. It is worth noting, however, that both relations do not follow the one of Trifunac and Brady (1976) for the western USA. This may be attributed to differences between the two regions in both source parameters (focal depth, etc.) and crustal structure.

Estimation of Q_s values

As we already mentioned above, the slope B, of the log spectrum provides information for the quality factor, Q_s. From eqs. (9) and (10) we have:

$$Q_s = -\frac{\pi \log e R}{BV_s} \tag{15}$$

Using eqn. (15) the Q_s values were calculated for each ray path, resulting in values that range from 30 to 360, with an average value equal to 130. These values are also listed in Table 1 under the column Q_s. As expected from eqn. (13), we have a strong increase of the calculated Q_s with distance, in accordance with other researchers (Anderson and Hough, 1984). This behaviour cannot be predicted by the homogeneous half-space model that we used but a rough explanation can be deduced if we assume that we have a "sedimentary" layer over the granitic one. The value of V_s in this layer is approximately 2 km/s

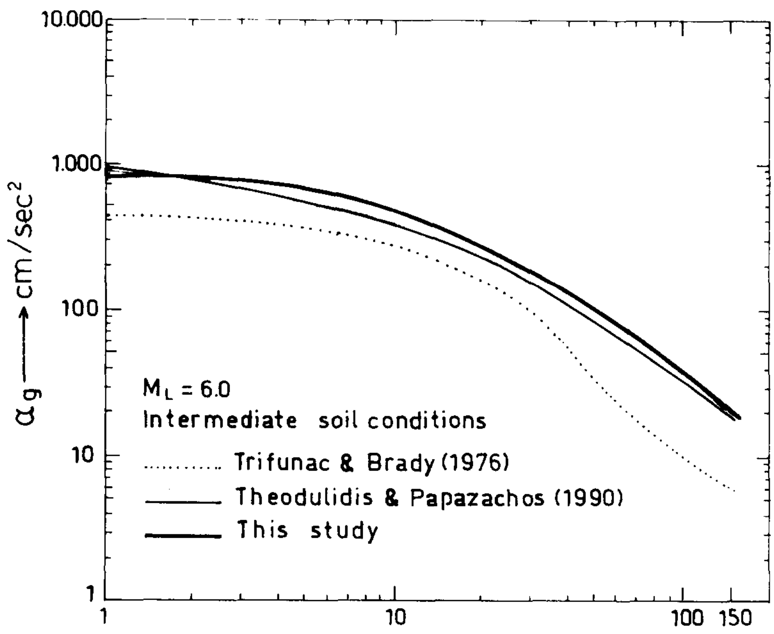

Fig. 7. Comparison between peak ground acceleration curves proposed in this study, in Theodulidis and Papazachos (1990) for Greece and by Trifunac and Brady (1976) for the western USA.

for Greece (Panagiotopoulos, 1984; Skordilis, 1985). The large velocity contrast with the granitic layer means that for large distances seismic waves travel almost horizontally, mainly through the granitic layer, for a distance approximately equal to Δ and are refracted at near the critical angle, i_c, near the recording stations. The total attenuation time is now equal to:

$$t^* = \frac{\Delta}{Q_{s_g} V_{s_g}} + \frac{H/\cos i_c}{Q_{s_s} V_{s_s}} \tag{16}$$

where V_{s_s}, V_{s_g}, Q_{s_s} and Q_{s_g} are the velocities and the quality factors for the two layers and H is the thickness of the "sedimentary" layer. The attenuation behaviour predicted by eqn. (16) is the same with that of eqn. (13). Using the values found in eqn. (13) and assuming an average value of $H = 3$ km for Greece (Panagiotopoulos, 1984) we found the values of 30 and 600 for the quality factors of the top and the bottom layers, respectively. Although the model used is a simplification of the true situation, the resulting Q-values are characteristic of the high attenuative properties of shallow depths in Greece. This feature, also observed in other areas (Al-Shukri and Mitchel, 1990), indicates that the total wave attenuation is mainly determined by attenuation at shallow depths. A typical case is that of the 9/15/86 earthquake in Kalamata (southern Greece) where the three near-epicenter recordings show the same low Q values (around 35). Such knowledge of some constraints on Q_s, in the near-surface settings, is quite useful in the estimation of seismic wave amplification in poorly consolidated basin or valley fill.

The whole procedure previously described is based on the hypothesis that Q is frequency independent. Usually, a frequency dependence of Q in the form of $Q = Q_0 \cdot f^n$ is assumed where n varies regionally (Mitchell, 1981; Singh and Hermann, 1983; Frankel, 1991). In this case a f^{1-n} dependence of the log spectrum and not a linear one should be expected. It is obvious that the method here applied cannot give any information on this dependence, but simply provides an average value for the quality factor in the frequency window studied, that is approximately 4 to 10 Hz in the majority of cases. However, in this study we are rather interested in the earthquake "size" assessment and in the vertical variations of attenuation in the area studied; hence the frequency-independent attenuation has little effect on the results and conclusions previously mentioned.

Conclusions

On the hypothesis that peak ground acceleration is due to direct S-waves propagation in a half-space, source "size" is estimated and correlated with M_L, thus providing calibration formulas for local magnitude estimation. Two formulas were derived, depending upon the way the anelastic attenuation is determined. Using the spectral decay of the log spectrum, attenuation times and corresponding quality factors were calculated for each accelerogram. The results show a clear increase of Q_s with distance and a strong attenuation of seismic waves for small epicentral distances. This behaviour is roughly explained by the existence of a highly attenuative "sedimentary" layer just above the lower attenuative granitic layer. A larger amount of data and a good upper crust velocity model are clearly needed in order: (a) to proceed to a full 3 D Q_s inversion, and (b) incorporate possible site effects which were neglected in the present study.

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