Pure appl. geophys. 150 (1997) 91–111
0033–4553/97/010091–21 $ 1.50 + 0.20/0

Pure and Applied Geophysics

Seismic Source Parameters from a New Dial-up Seismological Network in Greece

G. CHOULIARAS[1] and G. N. STAVRAKAKIS[1]

Abstract—A regional dial-up PT telemetric network of eight digital short-period seismic stations has been in full operation since the beginning of 1995 in Greece. During the first year of full operation, three destructive earthquake sequences struck different regions of central and northern Greece. The dial-up network managed to successfully record more than 85% of these seismic events including foreshocks, main shocks and aftershocks, and 45 of these events were selected in order to determine their seismic source parameters by spectral analysis. The results show that seismic moment correlates very well with earthquake magnitude, fault radius and fault displacement. The determined scaling relations are in general agreement with those obtained by other studies for Greek earthquakes that used different methodologies. Since digital seismic instrumentation is now expanding in Greece, these first results from spectral analysis of digital short-period data can be considered useful for future seismic hazard studies.

Key words: Dial-up telemetry, spectral analysis, seismic source parameters, greek earthquakes.

Introduction

A network of eight Lennartz MARS88MC, digital seismic stations has been installed by the Institute of Geodynamics of the National Observatory of Athens (NOA), as part of its national seismic network in the locations shown on Figure 1. Four stations are equipped with Teledyne S13 vertical component seismometers (ATH, SMG, NEO, VLS) and four stations are equipped with Lennartz LE3D three-component seismometers (PLG, JAN, EVR, ITM). All the remote stations are operating in triggered mode, using the classical STA/LTA criterion combined with high and low pass filters and a coincidence criterion in the case of the three-component sensors. The sampling frequency has been set to 62.5 Hz and timing accuracy around 0.1 pps is achieved by synchronization of the internal clock with a DCF antenna. Each MARS88MC station has a 1 Mb data storage capacity of a 120 db dynamic range.

Each of the eight remote seismic stations is connected via an asyncronous RS-232 port to a US Robotics Sportster 14400 modem which operates on a

[1] Institute of Geodynamics, National Observatory of Athens, P.O. Box 20048, 118 10 Athens, Greece.

conventional dial-up telephone line. Communication with the remote stations from Athens is achieved by an Ethernet Gateway PC which performs a cyclic interrogation of the stations with a host modem using MNP protocol and error correction. The Gateway is interfaced with a HP 750, Unix based workstation which is used to store the seismic events for further analysis. This data base is accessible on request to the Internet users and also to the MEDNET workstations in Italy and Spain by satellite via a Digital DEC 3000 system that is linked to a personal earth station. In this manner, it is possible for the three working groups (Greece, Italy and Spain) to quickly exchange seismic data in order to monitor the seismicity of the Mediterranean region.

The NOA dial-up network has been testing different data acquisition and retrieval techniques in the last three years and for this reason there were many nonoperative periods. 1995 was the first year of continuous data retrieval from the remote stations and during this year three destructive earthquake sequences occurred in three different regions, Arnea northern Greece, Kozani northern Greece and Aigio central Greece, with intense aftershock activity.

It is the purpose of this study to use the recorded data from 45 well recorded events from the NOA dial-up seismological network (Fig. 2) in order to determine earthquake source parameters by spectral methods and to compare their scaling relations with the ones previously determined by different investigations.

Data and Analysis

It is well known that Greece has the highest seismicity in Europe and that it accounts for 2% of the global seismicity, even though it covers only 0.09% of the global area (BÅTH, 1983).

In 1995, a total of 680 seismic events of magnitude $M_L > 3.5$ were registered in Greece according to the bulletins of NOA. Of these events, 32 were larger than $M_L > 45$ and most of these seismic events were due to three consecutive and damaging earthquake sequences that devastated parts of Greece within weeks of each other. The main shocks of the three sequences occurred at:

i) Arnea (40.56°N, 23.69°E, Depth = 9 km) on May 4, 1995 at 00:34 GMT with a local magnitude $M_L = 5.0$.
ii) Kozani (40.24°N, 22.74°E, Depth = 5 km) on May 13, 1995 at 08:47 GMT with a local magnitude $M_L = 6.1$.
iii) Aigio (38.36°N, 22.15°E, Depth = 26 km) on June 15, 1995 at 00:15 GMT with a local magnitude $M_L = 5.6$.

The Arnea sequence in northwestern Greece and the Aigio sequence in central Greece occurred on known seismic zones that have caused strong earthquakes. However, the Kozani earthquake was a total surprise to the seismological community because the area was never activated according to the available instrumental and historical data. More than 85% of the seismic events of magnitude $M_L > 3.5$ were recorded by the NOA dial-up network during the period of intense activity in May–June 1995. The main reasons for missing otherwise detectable events are due to instrumental limitations.

In particular, the MARS88MC digital seismograph that is employed in the remote stations does not allow for simultaneously acquiring and sending data or information by modem to the host computer in Athens. Therefore, if a seismic event occurs during these communication intervals it will not be recorded by that station. This interval lasts some tens of seconds approximately every 10 minutes at each station and this minimizes detection failure. Another limitation is that the remote stations are powered by 220V AC supply and data acquisition or transmission cannot occur in times of power failure.

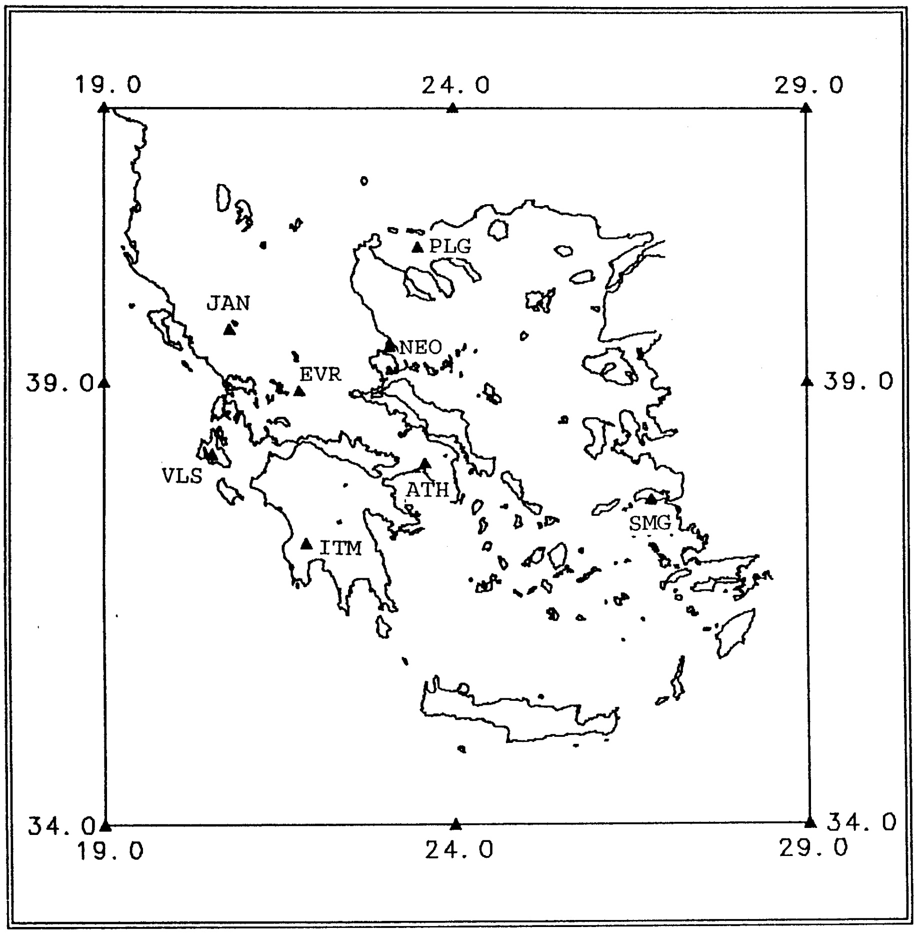


Figure 1
Map of Greece showing the locations of the MARS88MC digital stations.

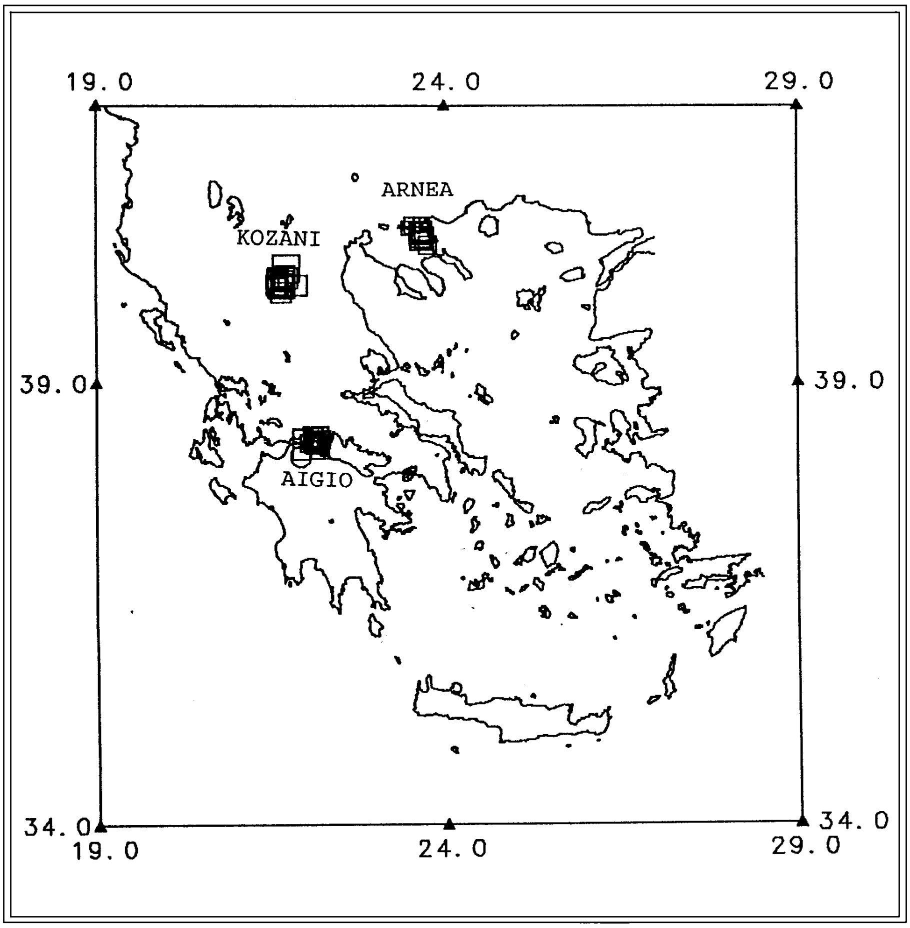


Figure 2
Epicentres of the Arnea, Kozani and Aigio earthquakes used in this study.

In order to contribute to the seismotectonic and seismic hazard studies of the Greek area, the digital short-period data from the regional dial-up network were used to determine the earthquake source parameters and scaling relations of 45 well recorded events of local magnitude $2.8 < M_L < 6.1$ by spectral analysis methods.

To the authors knowledge it is the first time that short-period digital data in this magnitude range from a regional network in Greece were used to determine earthquake source parameter scaling relations by spectral methods.

Previous studies concerned with scaling relations of earthquake source parameters in Greece have used different methods such as spectral modelling of surface waves, teleseismic inversion of body waves and double-couple solutions (NORTH 1977; KIRATZI *et al.*, 1985; TSELENTIS *et al.*, 1988; MAIN and BURTON, 1990).

More recently, spectral analysis methods on microearthquake data have also provided determinations of scaling relations for the Gulf of Patras in central Greece (MELIS *et al.*, 1995).

Source parameter determination from spectral analysis of P- or S-wave data has been the focus of many studies, and good examples include the pioneering works of KEILIS-BOROK (1959), HANKS and WYSS (1972), THATCHER and HANKS (1973) and for the Greek area the work of BURTON *et al.* (1995).

The analysis procedure used in this study employs the PITSA signal processing toolbox (SCHERBAUM and JOHNSTON, 1993) and it involves spectral analysis of selected time windows of the displacement trace which is always corrected for the appropriate instrument response. The discrete Fourier spectrum (DTF), containing the P or S wave is modelled by a Brune source spectrum (BRUNE, 1970, 1971) and an attenuation operator using a Marquardt-Levenberg nonlinear inversion technique (PRESS *et al.*, 1988). As to the noise estimate required for the chi-square minimization, this is obtained from a single noise window prior to the P-wave arrival, using the same window length and tapering parameters as for the signal window. The starting model parameters for the inversion are obtained interactively using the spectral fitting tool of the PITSA program by SCHERBAUM and JOHNSTON (1993). Details of the fitting procedure can be found in SCHERBAUM (1990).

The characteristics of the source for a seismic event are determined from the two parameters of the P- or S-wave log-displacement spectra. These parameters are the long-period spectral level Ω_0, and the spectral corner frequency f_c (AKI and RICHARDS, 1980).

A general feature of all dislocation models is that the long-period level Ω_0 is proportional to the seismic moment and that the corner frequency f_c is inversely proportional to the source dimension, r.

Following KEILIS-BOROK, (1959):

$$M_{0_{(p,s)}} = \frac{(4\pi \cdot \rho \cdot v^3_{(p,s)} \cdot \Omega_{0_{(p,s)}} \cdot R)}{(k \cdot R\theta\phi_{(p,s)})}, \tag{1}$$

where $\rho = 2.64\ \mathrm{g/cm^3}$ is the density of the medium, R is the hypocentral distance between the source and the receiver, $V_{p,s}$ are the P- or S-wave velocities near the source with $V_p = 6$ km/sec and $V_s = 3.3$ km/sec, k is the free surface operator $= 2$ and $R_{(p,s)} = 0.5$ is the average radiation pattern coefficient. Logarithmic weighted averages over all records have been computed (ARCHULETTA *et al.*, 1982) to obtain mean long-period spectral levels and seismic moments.

Source radii have been computed using the corner frequencies $f_{c_{(p,s)}}$ also by weighted logarithmic averages based on the models of BRUNE (1970, 1971) and MADARIAGA (1976):

$$\text{Brune model} \quad r = \frac{0.37\, V_{(p,s)}}{f_{c_{(p,s)}}} \tag{2}$$

$$\text{Madariaga model} \quad r = \frac{0.32\,V_s}{f_{c_{(p)}}} = \frac{0.21\,V_s}{f_{c_{(s)}}}. \tag{3}$$

Stress drops ($\Delta\sigma$) and average displacement on the fault plane ($\langle u \rangle$) have been computed from the mean values of moments and source radii of P and S waves according to: (KEILIS-BOROK, 1959; BRUNE, 1968).

$$\Delta\sigma = \frac{0.44 M_0}{r_3} \tag{4}$$

$$\langle u \rangle = \frac{M_0}{\pi \mu r^2}, \tag{5}$$

where $\mu = 3 \times 10^{10}\ \mathrm{Nm}^{-2}$ is shear modulus.

Results and Discussion

The earthquakes used in this study are listed in Table 1 together with their hypocentral coordinates and local magnitudes obtained from the monthly bulletins of NOA. Events 1 to 10 correspond to the Arnea seismic sequence, 11 to 26 to the Kozani seismic sequence and 27 to 45 to the Aigio seismic sequence.

Figure 3 shows the velocity traces of an $M_L = 4.0$ aftershock event from the Aigio seismic sequence, as recorded by the dial-up network, and in Figure 4 we see an example of the P- and S-wave displacement spectra and the noise spectra from the EVR station. In our study, a whole path Q model with a constant Q value of 200 was chosen, and Ω and f_c were adjusted accordingly using the interactive spectral fitting option in the PITSA signal processing toolkit (SCHERBAUM and JOHNSTON, 1993). It is well known that spectra in general are equally well fitted by a low Q/high f_c or a high Q/low f_c spectral model as has been discussed by SCHERBAUM (1990). An example of this tradeoff between Q and f_c is that a $Q = 85$ will give $f_c = 16.6$ Hz while for the same event a $Q = 184$ will give $f_c = 6.3$ Hz. Since the corner frequency is used in calculating the seismic source dimension, stress drop and average displacement (see equations (2) to (5)), the effect of Q is considerable in obtaining reliable source parameters. On the other hand, the effect of Q on the seismic moment is less pronounced.

For the Greek area, HASHIDA *et al.* (1988) give Q values in the range of 100 to 400 and we chose a $Q = 200$ in this study because this value was found to produce results that were in general agreement with the results from macroseismic surveys in the epicentral region, as we will see in this section. The seismic moments, corner frequencies, together with the fault radii, stress drops and average displacements determined according to Brune's and Madariaga's models, are listed in Table 2.

In general observation of the seismic moments obtained from the SH spectra are lower than those determined from the P spectra for our entire data set. The fact

Table 1

List of earthquakes used in this study together with their epicentral coordinates (°N, °E), depth (H) and local magnitudes (M_L) from the monthly bulletins of NOA.

No.	Date	Time (GMT)	°N	°E	H (km)	M_L	
1	1995 APR 4	17 10 9.7	40.63	23.68	6	4.1	Arnea
2	1995 MAY 3	14 16 40.9	40.49	23.77	12	3.8	Arnea
3	1995 MAY 3	15 39 55.5	40.64	23.69	8	4.0	Arnea
4	1995 MAY 3	21 36 54.3	40.58	23.67	14	4.3	Arnea
5	1995 MAY 3	21 43 27.5	40.61	23.66	24	4.5	Arnea
6	1995 MAY 4	00 34 10.7	40.56	23.69	9	5.0*	Arnea
7	1995 MAY 4	00 43 41.6	40.55	23.72	7	3.9	Arnea
8	1995 JUN 4	03 18 8.9	40.70	23.55	5	3.2	Arnea
9	1995 JUN 12	10 14 51.8	40.67	23.50	10	2.8	Arnea
10	1995 JUN 27	02 52 2.0	40.70	23.69	2	3.0	Arnea
11	1995 MAY 13	08 47 13.6	40.24	21.74	5	6.1*	Kozani
12	1995 MAY 13	18 06 1.5	40.16	21.72	17	4.5	Kozani
13	1995 MAY 14	02 47 .6	40.12	21.61	4	4.5	Kozani
14	1995 MAY 14	05 59 16.8	40.06	21.59	5	4.3	Kozani
15	1995 MAY 14	09 45 41.9	40.14	21.77	2	4.3	Kozani
16	1995 MAY 14	21 31 12.9	40.05	21.71	2	4.3	Kozani
17	1995 MAY 15	04 13 57.3	40.06	21.68	5	5.0	Kozani
18	1995 MAY 16	23 00 41.9	40.03	21.63	5	4.3	Kozani
19	1995 MAY 16	23 57 28.5	40.07	21.71	2	4.6	Kozani
20	1995 MAY 17	04 14 26.0	40.05	21.70	5	5.1	Kozani
21	1995 MAY 19	06 48 50.4	40.09	21.62	6	4.8	Kozani
22	1995 JUN 11	18 51 48.5	39.99	21.67	8	4.3	Kozani
23	1995 JUN 19	03 54 .2	40.06	21.90	7	4.3	Kozani
24	1995 JUL 17	23 18 16.1	40.11	21.62	5	4.6	Kozani
25	1995 JUL 18	07 42 55.1	40.14	21.63	5	4.4	Kozani
26	1995 JUL 19	18 23 15.8	40.11	21.71	7	4.5	Kozani
27	1995 JAN 2	12 36 9.8	38.17	21.94	35	3.8	Aigio
28	1995 JAN 23	17 34 59.6	38.23	21.95	7	3.5	Aigio
29	1995 MAY 8	05 11 7.9	38.31	22.18	7	4.0	Aigio
30	1995 MAY 28	19 56 40.7	38.37	21.96	9	4.1	Aigio
31	1995 JUN 15	00 15 50.9	38.36	22.15	26	5.6*	Aigio
32	1995 JUN 15	00 31 3.7	38.33	22.18	26	5.1	Aigio
33	1995 JUN 15	01 16 22.4	38.36	22.18	8	3.5	Aigio
34	1995 JUN 15	04 51 19.9	38.25	22.21	3	4.0	Aigio
35	1995 JUN 15	07 01 .6	38.33	22.11	6	3.9	Aigio
36	1995 JUN 15	10 41 50.9	38.29	22.14	3	3.6	Aigio
37	1995 JUN 17	14 20 29.7	38.36	22.24	5	3.8	Aigio
38	1995 JUN 18	01 14 6.9	38.32	22.06	7	3.7	Aigio
39	1995 JUN 18	04 28 24.1	38.34	22.11	6	3.7	Aigio
40	1995 JUN 20	14 38 34.5	38.33	22.10	15	3.5	Aigio
41	1995 JUL 1	21 58 5.3	38.33	22.11	4	3.6	Aigio
42	1995 JUL 3	22 29 59.9	38.33	22.12	3	3.8	Aigio
43	1995 JUL 5	18 24 37.9	38.38	22.11	7	4.2	Aigio
44	1995 JUL 10	18 12 38.2	38.39	22.12	3	4.0	Aigio
45	1995 JUL 15	09 36 58.4	38.37	21.95	5	3.6	Aigio

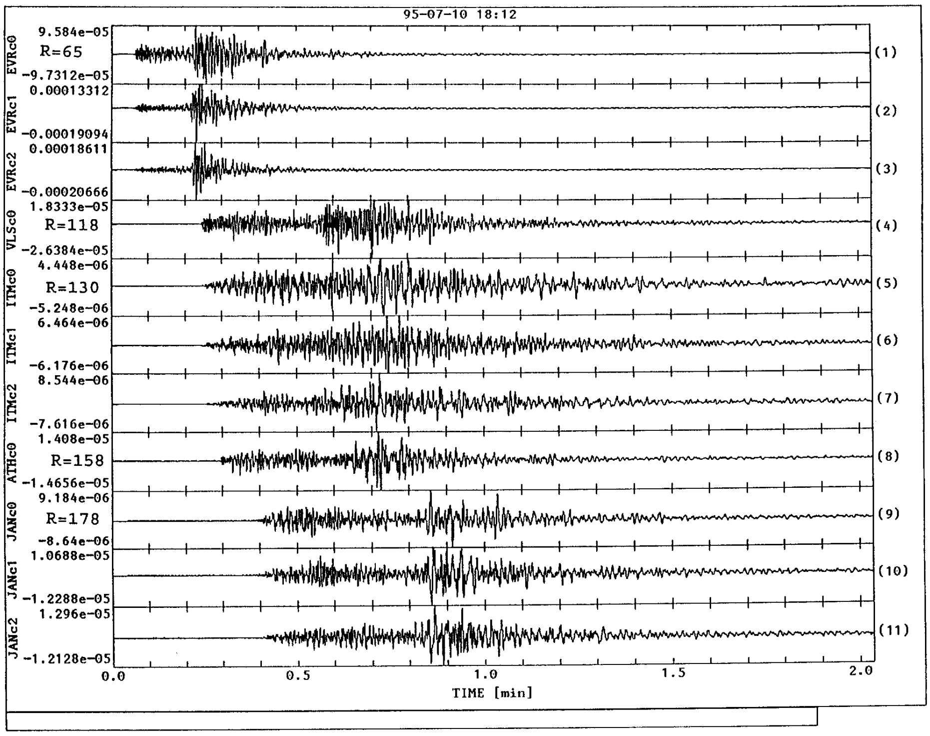


Figure 3

Recordings of $M = 4.0$ Aigio aftershock from the dial-up network. c0, c1, c2 indicate Z, N-S and E-W components of the recorded ground velocity in m/sec. R is the stations epicentral distance.

that we used the same average value of 0.5 for the radiation pattern for both sets of data (P and S waves) can explain this difference since the low frequency spectral amplitudes are corrected for radiation pattern when computing the seismic moment.

In order to examine the validity of these results, we first compare the moments determined in this study (LENET) by spectral analysis to the moments obtained from centroid moment tensor solutions by USGS, Harvard, Tokyo ERI and GFZ Potsdam, for the May 13 and June 15 main shocks in Kozani and Aigio, respectively:

Kozani Earthq.: M_0 (USGS) $= 7.2 * 10^{25}$ (dyne-cm)
M_0 (Harvard) $= 7.6 * 10^{25}$ (dyne-cm)
M_0 (ERI) $= 4.79 * 10^{25}$ (dyne-cm)
M_0 (LENET) $= 2.53 * 10^{26}$ (dyne-cm)

Aigio Earthq.: M_0 (Harvard) $= 5.7 * 10^{25}$ (dyne-cm)
M_0 (GFZ) $= 5.5 * 10^{25}$ (dyne-cm).

$$M_0 \text{ (LENET)} = 9.53 * 10^{24} \text{ (dyne-cm)}.$$

In this comparison, we observe that the LENET results for the Kozani main event overestimate the seismic moments of the other sources, while for the Aigio main event, the spectral analysis results underestimate the seismic moment from the moment tensor inversion.

In our analysis we used the same radiation pattern coefficient in determining the seismic moments for all events, $R = 0.5$ in equation (1). Thus differences in the station distribution with respect to the different epicentral areas and differences in the focal mechanisms between the analyzed earthquakes accommodate for the observed differences in seismic moments.

In both cases however, the difference is within the error limits in determining seismic moments from short-period data which in many cases finds M_0 to vary by a factor of 2 or even 3 for the same event (NORTH, 1973).

Source radii, stress drops and average displacements have been computed using two models. Brune's model (BRUNE, 1970, 1971) is used to compare our results

(a)

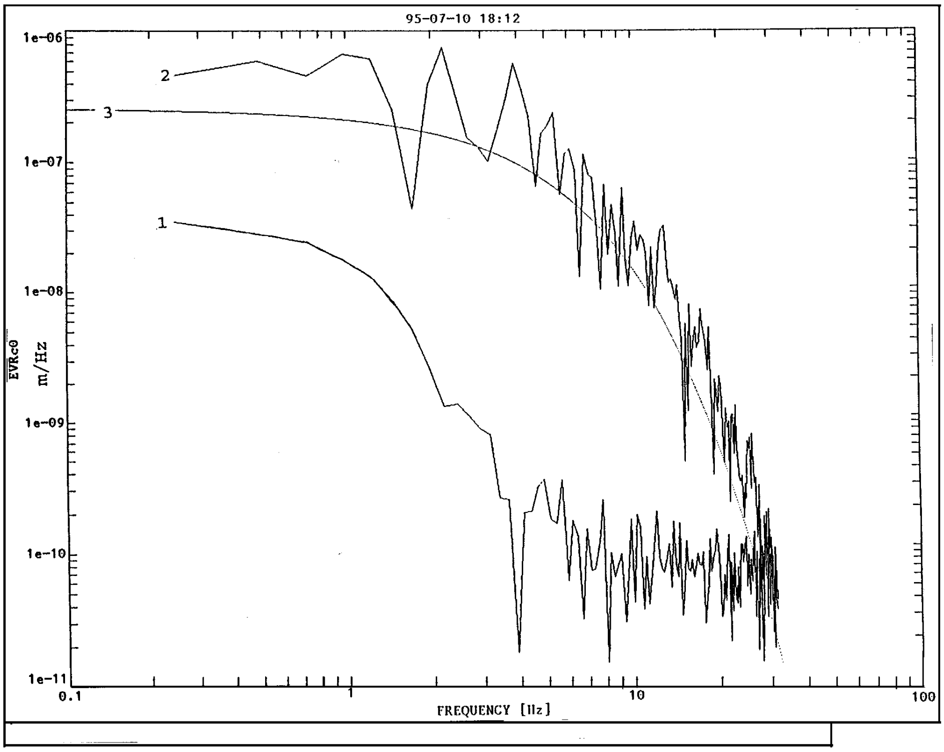


Figure 4a.

(b)

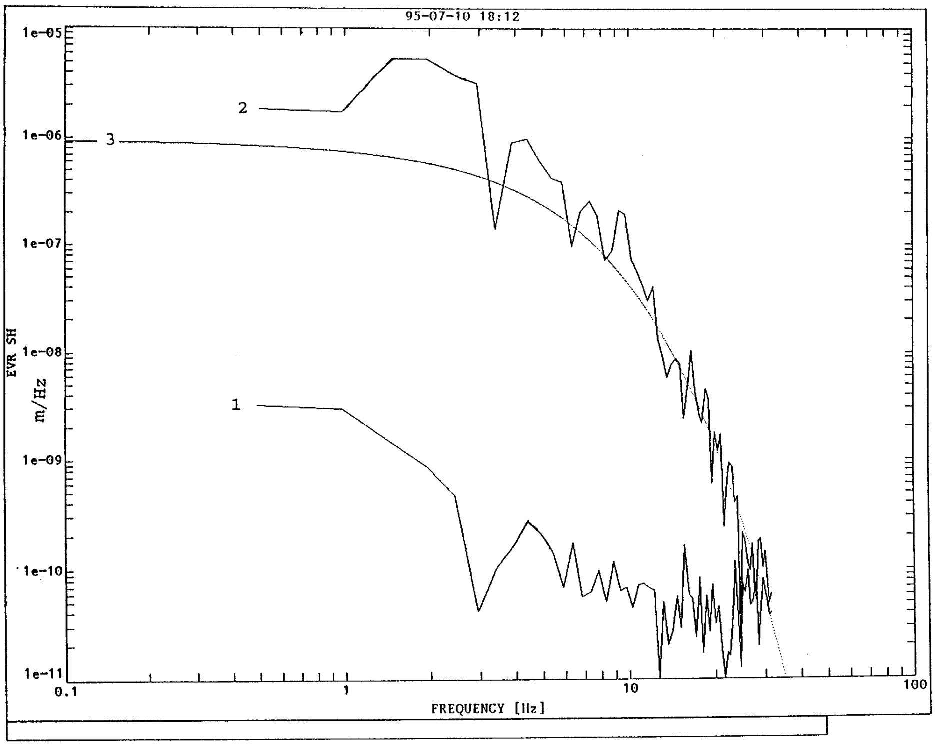


Figure 4
a. Displacement spectra of a noise window (trace 1), of a *P*-wave window (trace 2) and the spectral fitting result (trace 3) from the event shown on Figure 3 at the EVR station. b. Displacement spectra of a noise window (trace 1), of a *SH*-wave window (trace 2) and the spectral fitting result (trace 3) from the event shown on Figure 3 at the EVR station.

with previous studies that adopted this model, and MADARIAGA's (1976) model is used because it is more realistic in the sense that it accounts properly for propagation and stoppage of the rupture process for circular faults.

Macroseismic surveys of the Aigio epicentral region have reported 6–7 cm fracture displacements (I.G.M.E., 1995) while in Kozani PAVLIDES *et al.* (1995) report fracture displacement of 20 cm near the epicentral area. When comparing these results to the average displacements determined in this study from Table 2, (events 31 and 11, respectively for Aigio and Kozani), we observe that MADARIAGA's (1976) fault model provides more realistic results while BRUNE's model (1970, 1971) overestimates the observed displacements.

Figure 5a shows the moment magnitude behavior of our entire data set. The correspondence seems fair with very little scatter, considering that the data set is from 3 different seismic regions. Least squares fit gives the relationship:

Table 2

List of the spectral analysis results concerning seismic moments (M_0), corner frequencies (f_c), fault radii (R_B, R_M), stress drops ($\Delta\sigma_B$, $\Delta\sigma_M$), average displacements (U_B, U_M). Subscripts B and M refer to the Brune and to the Madariaga fault models, respectively. Event numbers correspond to those listed on Table I

No.	M_L	M_0 (dyne cm)	f_c (Hz)	R_B (km)	R_M (km)	$\Delta\sigma_B$ (bar)	$\Delta\sigma_M$ (bar)	U_B (cm)	U_M (cm)	
1	4.1	6.14e22	0.75	2.27	1.16	2.31	17.31	1.26	4.84	A
2	3.8	2.54e22	0.99	1.70	0.88	2.27	16.40	0.93	3.48	
3	4.0	5.95e22	0.74	2.31	1.16	2.12	15.93	1.18	4.53	R
4	4.3	1.30e23	0.71	2.43	1.24	3.98	30.00	2.33	8.97	
5	4.5	1.13e23	0.48	3.55	1.81	1.11	8.38	0.95	3.66	N
6*	5.0	5.34e23	0.37	4.55	2.34	2.49	18.33	2.74	10.46	
7	3.9	4.46e22	0.90	1.87	0.97	3.00	21.50	1.35	5.03	E
8	3.2	7.99e20	2.30	0.74	0.38	0.87	6.41	0.15	5.53	
9	2.8	8.78e19	6.00	0.29	0.15	1.58	11.45	0.11	0.41	A
10	3.0	9.99e20	2.50	0.66	0.35	1.52	10.25	0.24	0.86	
11*	6.1	2.53e26	0.06	26.10	13.53	6.26	44.94	39.41	146.64	
12	4.5	1.78e23	0.60	3.70	1.76	1.55	14.37	1.38	6.10	
13	4.5	1.74e23	0.54	3.15	1.62	2.45	18.01	1.86	7.03	K
14	4.3	1.86e23	0.72	3.08	1.47	2.80	25.76	2.08	9.13	
15	4.3	1.39e22	0.59	2.88	1.48	0.26	1.89	0.18	0.67	O
16	4.3	1.38e22	0.79	2.12	1.09	0.64	4.69	0.33	1.23	
17	5.0	6.08e23	0.32	5.55	3.02	1.56	9.71	2.09	7.07	Z
18	4.3	1.27e23	0.71	2.38	1.22	4.15	30.77	2.38	8.55	
19	4.6	5.89e23	0.49	3.56	1.81	5.74	43.71	4.93	19.08	A
21	5.1	8.65e23	0.24	7.28	3.71	0.99	7.45	1.73	6.66	
21	4.8	3.63e23	0.38	4.49	2.31	1.76	12.96	1.91	7.22	N
22	4.3	4.55e22	0.64	2.62	1.34	1.11	8.32	0.70	2.69	
23	4.3	1.43e23	0.64	2.66	1.36	3.34	25.01	2.14	8.20	I
24	4.6	3.79e23	0.45	3.74	1.92	3.19	23.56	2.87	10.91	
25	4.4	7.18e22	0.49	3.43	1.75	0.78	5.89	0.65	2.49	
26	4.5	1.83e23	0.43	3.94	1.94	1.32	11.03	1.25	5.16	
27	3.8	6.21e22	1.06	2.09	1.00	2.99	27.32	1.51	6.59	
28	3.5	2.48e22	1.53	1.45	0.69	3.58	33.22	1.25	5.53	
29	4.0	3.51e22	0.84	2.03	1.04	1.85	13.73	0.90	3.44	A
30	4.1	5.40e22	0.85	2.01	1.02	2.93	22.39	1.42	5.51	
31*	5.6	9.53e24	0.15	11.30	5.78	2.90	21.71	7.92	30.27	
32	5.1	8.40e23	0.37	4.64	2.38	3.70	27.42	4.14	15.73	I
33	3.5	1.85e21	1.55	1.09	0.56	0.63	4.64	0.17	0.63	
34	4.0	6.86e22	0.76	2.22	1.14	2.76	20.37	1.48	5.60	
35	3.9	2.72e22	0.75	2.26	1.15	1.04	7.87	0.57	2.18	G
36	3.6	6.67e21	1.60	1.39	0.66	1.09	10.21	0.37	1.62	
37	3.8	1.05e22	0.84	2.02	1.04	0.56	4.11	0.39	1.03	
38	3.7	8.08e21	0.97	1.73	0.89	0.69	5.04	0.31	1.08	I
39	3.7	3.35e22	0.88	1.93	0.99	2.05	15.19	0.95	3.63	
40	3.5	1.67e21	1.60	1.39	0.66	0.27	2.56	0.09	0.41	
41	3.6	2.85e22	1.29	1.31	0.67	5.58	41.69	1.76	6.74	O
42	3.8	8.94e22	0.86	1.96	1.00	5.22	39.34	2.47	9.49	
43	4.2	9.87e22	0.61	2.77	1.42	2.04	15.17	1.36	5.19	
44	4.0	1.94e22	0.84	2.00	1.03	1.07	7.81	0.51	1.94	
45	3.6	6.12e21	1.77	0.97	0.50	2.95	21.54	0.69	2.60	

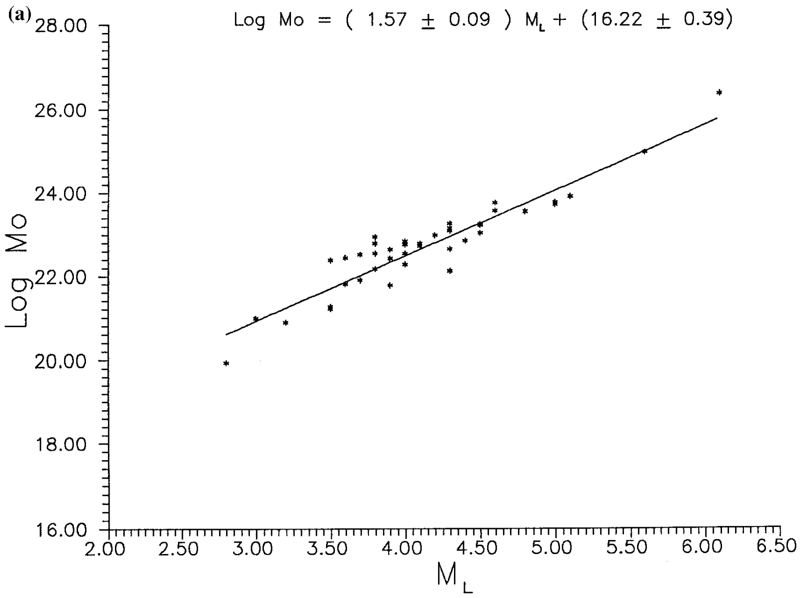


Figure 5a
Log M_0 (seismic moment) versus M_L (local magnitude) for the whole data set

$$\text{Log } M_0 = (1.57 \pm 0.09)M_L + (16.22 \pm 0.39) \tag{6}$$

for $6.1 > M_L > 2.8$ and for this magnitude range this result is in good agreement with the results of (NORTH, 1977; MAIN and BURTON, 1990) that determined seismic moments by surface and body wave modelling. Our result is also in agreement with the general result of PAPAZACHOS (1994) for the Greek area that was based on magnitude determinations and observations of fault displacement. Good agreement also exists between our result and those of MELIS *et al.* (1995) who also used spectral analysis of microearthquake data to determine scaling relations for central Greece. However, the results of KIRATZI *et al.* (1985) and TSELENTIS *et al.* (1988) differ from the above relationship, and the difference in the regression constants could be due to the fact that these studies focused on data sets from specific regions such as the western Hellenic arc and the Aegean Sea while the data set used in this study considers different regions of Greece. HANKS and BOORE (1984) argue that a difference in the regression coefficients between different regions "is a geographic appearance, not a geographic reality ... " and this is due to the differences in the data set of each region.

A summary of these results is given in Figure 5b for comparison and one must note that in all other studies the seismic moment is related to the surface wave magnitude whereas in this study the seismic moment is given as a function of the local magnitude.

From Table 2, we observe low stress drops for the Arnea, Aigio and Kozani main shocks as determined from Brune's model. Low stress drops for large Greek earthquakes have also been reported by many authors who employed Brune's model (KULHANEK and MEYER, 1979; SOUFLERIS and STEWART, 1981; KIM *et al.*, 1984; KIRATZE *et al.*, 1985; STAVRAKAKIS *et al.*, 1989; STAVRAKAKIS and BLIONAS, 1990). For the same events, Madariaga's model provides stress drops almost ten times larger than those determined by Brune's model and the question is whether these results are model dependent or whether low stress drop events are a reality for many seismogenic regions in Greece.

BRUNE *et al.* (1985) proposed that low stress drop events from Brune's model can be explained by partial stress drop events when the stress release is not uniform over the fault plane but occurs as a series of multiple events, with parts of the fault remaining unbroken. Thus the average displacement over the fault plane will be less than the displacement at the asperities and stress drop over the individual subevents. STAVRAKAKIS *et al.* (1986, 1987) have demonstrated that strong earthquakes in Greece were complex events and based on this result, the partial low stress drop model seems to be realistic.

Both Brune's and Madariaga's models use the corner frequency to obtain the radius of the fault plane, and in this sense we first examine the moment versus corner frequency behavior for our data set. In Figure 6 the relationship is clear and the linear least squares fit to our data gives:

$$\mathrm{Log}(M_0) = (-3.14 \pm 0.16)\ \mathrm{Log}\ f_c + (22.33 \pm 0.05). \tag{7}$$

PRESS (1967) and SCHICK (1970) found that a relationship exists between earthquake magnitude and the radius of a circular fault plane from the Brune model. Figure 7 illustrates that a similar relationship exists in our data set

$$M_L = (1.92 \pm 0.10)\ \mathrm{Log}\ r_B + (3.43 \pm 0.04) \tag{8}$$

$$M_L = (1.95 \pm 0.10)\ \mathrm{Log}\ r_M + (3.97 \pm 0.03) \tag{9}$$

and this is also in good agreement with the results of KIRATZI *et al.* (1984) that developed the relationship for a rectangular fault model from Greek earthquakes.

Very good correlation also exists between the seismic moment and the fault radius (Fig. 8) from the whole data set and least squares fit gives:

$$\mathrm{Log}\ M_0 = (3.23 \pm 0.15)\ \mathrm{Log}\ r_B + (21.52 \pm 0.08) \tag{10}$$

$$\mathrm{Log}\ M_0 = (3.18 \pm 0.19)\ \mathrm{Log}\ r_M + (22.44 \pm 0.06). \tag{11}$$

Equally good correlation between seismic moment and average coseismic slip is found from our data set (Fig. 9) and these relationships are:

$$\mathrm{Log}\ M_0 = (1.98 \pm 0.12)\ \mathrm{Log}\ U_B + (22.72 \pm 0.06) \tag{12}$$

$$\mathrm{Log}\ M_0 = (1.92 \pm 0.16)\ \mathrm{Log}\ U_M + (21.56 \pm 0.13). \tag{13}$$

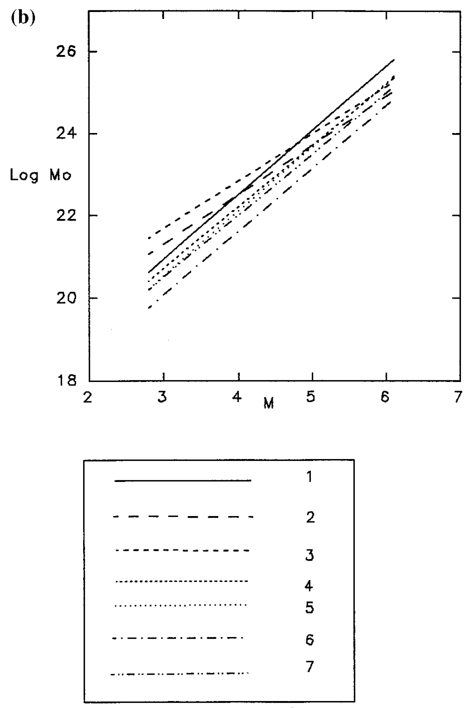


Figure 5b
Log M_0 (seismic moment) versus magnitude for Greek earthquakes by different studies. 1—this study, 2—KIRATZI *et al.* (1985), 3—TSELENTIS *et al.* (1988), 4—NORTH (1977), 5—MAIN and BURTON (1990), 6—MELIS *et al.* (1995), 7—PAPAZACHOS (1994).

Finally, the seismic moment versus stress drop curve in Figure 10 shows great scatter in seismic moments for a given stress drop and the least squares fit gives:

$$\text{Log}\, M_0 = (1.66 \pm 0.43)\ \Delta\sigma_B + (22.34 \pm 0.18) \tag{14}$$

$$\text{Log}\, M_0 = (1.64 \pm 0.43)\ \Delta\sigma_B + (20.91 \pm 0.51). \tag{15}$$

To the authors knowledge, little work has been published in the literature concerning the determination of scaling relations 10–15 from spectral analysis of short-period data for Greek earthquakes in the magnitude range used in this study. Recently MELIS *et al.* (1995) also determined similar relationships such as the one of this study for the region around Aigio from a spectral analysis of micro-earthquakes. In our data set, 19 of 45 seismic events are from the Aigio region, with local magnitudes ranging from 3.5 to 5.6. Therefore these equations are quite representative of that region. When we compare these with those respectively

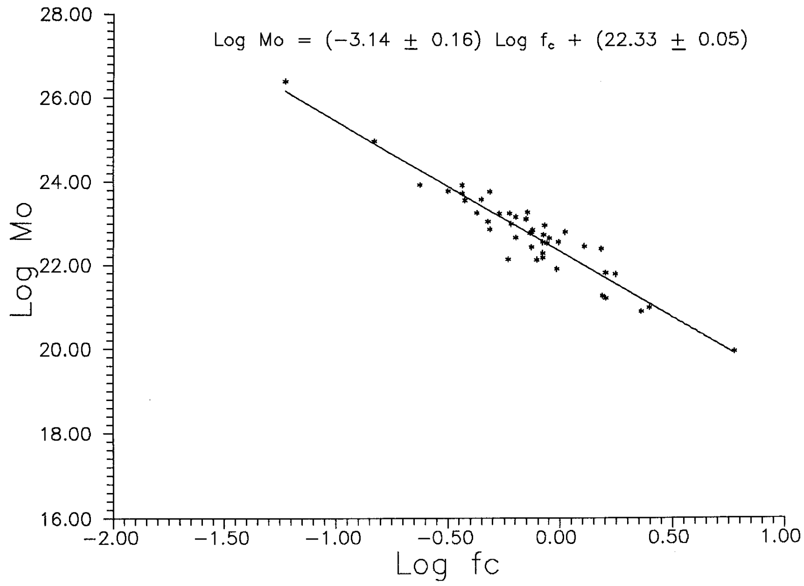


Figure 6
Log M_0 (seismic moment) versus Log f_c (corner frequency) for the whole data set.

determined by MELIS *et al.* (1995), we note many similarities of scaling relations. Small differences in the regression coefficients are most probably due to the fact that microearthquakes may have different scaling relations compared to larger earthquakes.

Conclusions

In this study, short-period seismic data from the first Greek national network of digital PT telemetry stations from 1995 were selected in order to determine their seismic source parameters by spectral analysis methods and to compare these and their scaling relations with those determined by other methods.

The source parameters and their scaling relations from this study are generally in good agreement with the results of different investigations for Greek earthquakes. Unfortunately, little work in spectral analysis of short-period data from Greece has been conducted primarily due to a lack of digital instrumentation. The Institute of Geodynamics of the National Observatory of Athens is in the process of expanding its digital network to 26 stations. In the meantime, it is believed that these first results and scaling relations from short-period data can be used by other researchers who are concerned with seismic source parameters and hazard analysis of the Greek region.

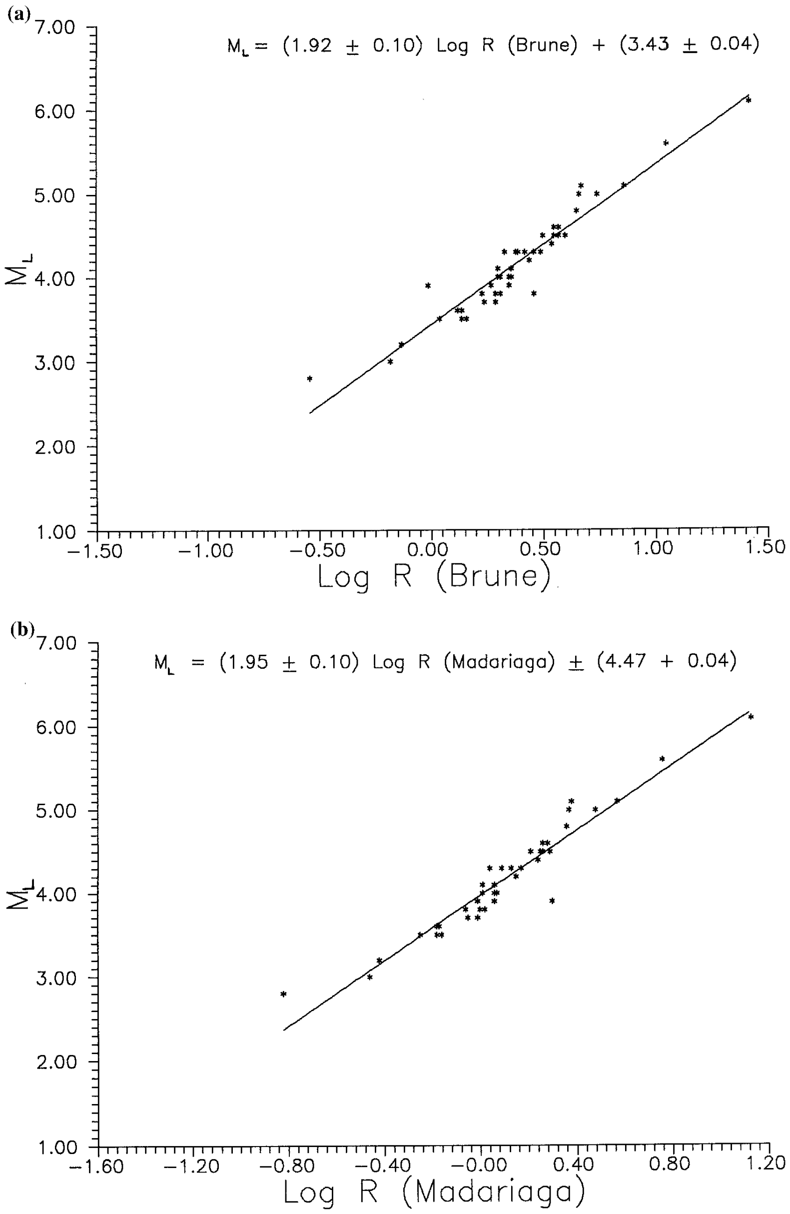


Figure 7

a. M_L (local magnitude) versus Log R_B for the whole data set. R_B is the fault radius for the Brune model. b. M_L (local magnitude) versus Log R_M for the whole data set. R_M is the fault radius for the Madariaga model.

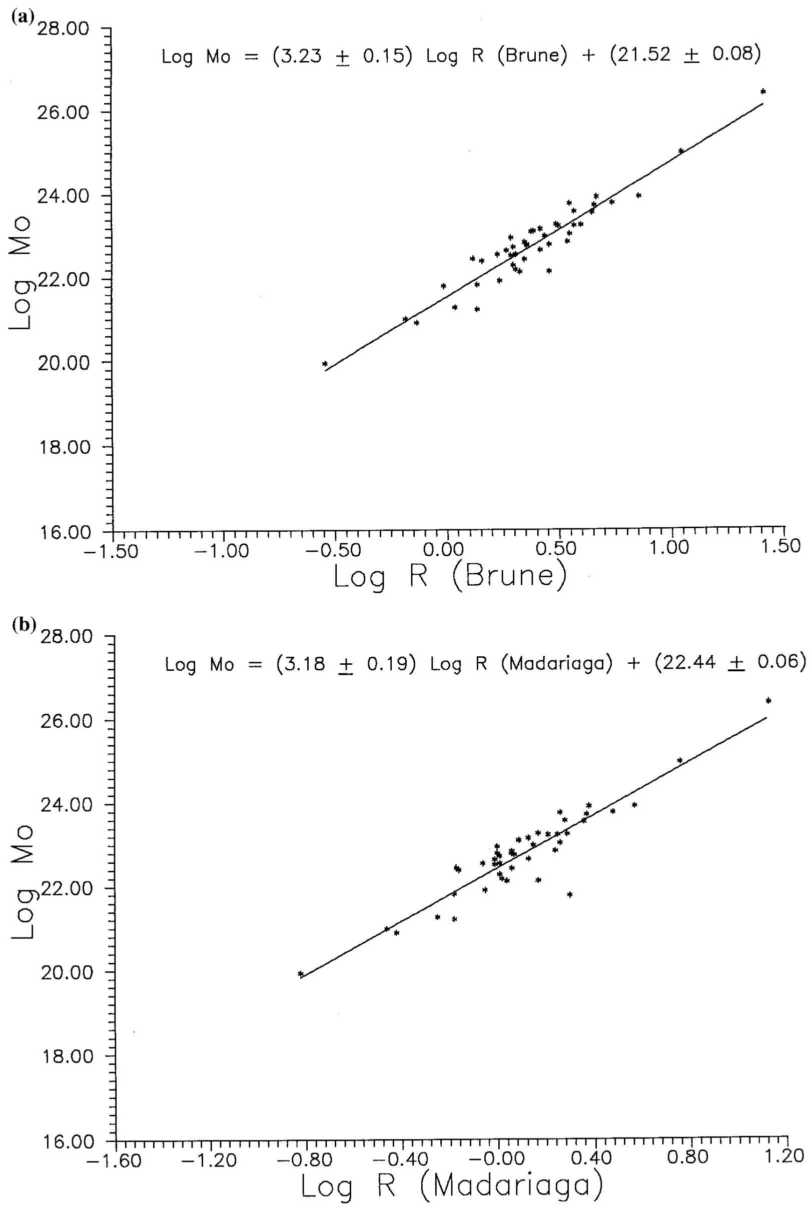


Figure 8

a. Log M_0 (seismic moment) versus Log R_B for the whole data set. R_B is the fault radius for the Brune model. b. Log M_0 (seismic moment) versus Log R_M for the whole data set. R_M is the fault radius for the Madariaga model.

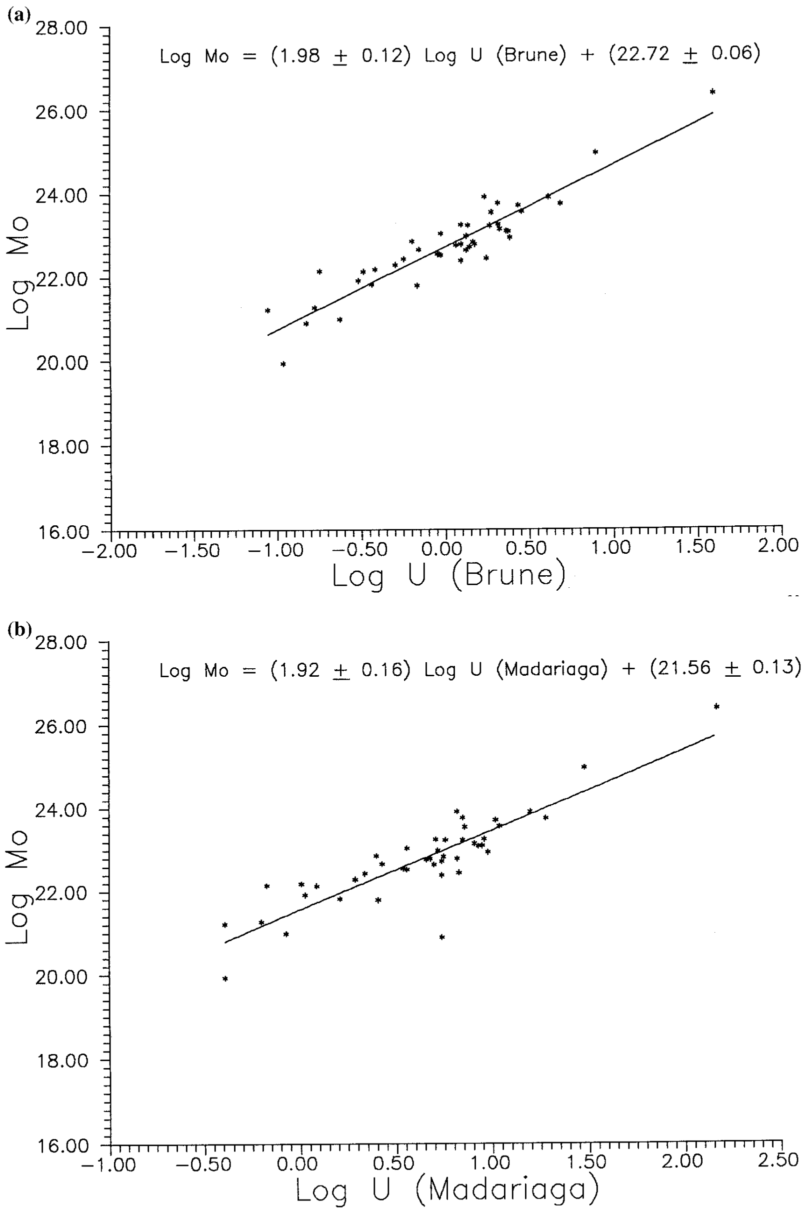
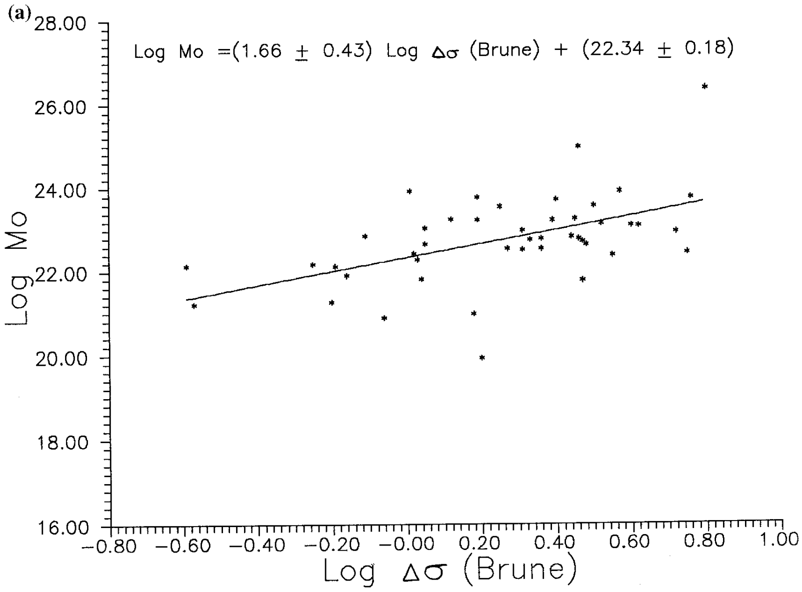


Figure 9

a. Log M_0 (seismic moment) versus Log U_B for the whole data set. U_B is the average displacement for the Brune model. b. Log M_0 (seismic moment) versus Log U_M for the whole data set. U_M is the average displacement for the Madariaga model.

(a)



(b)

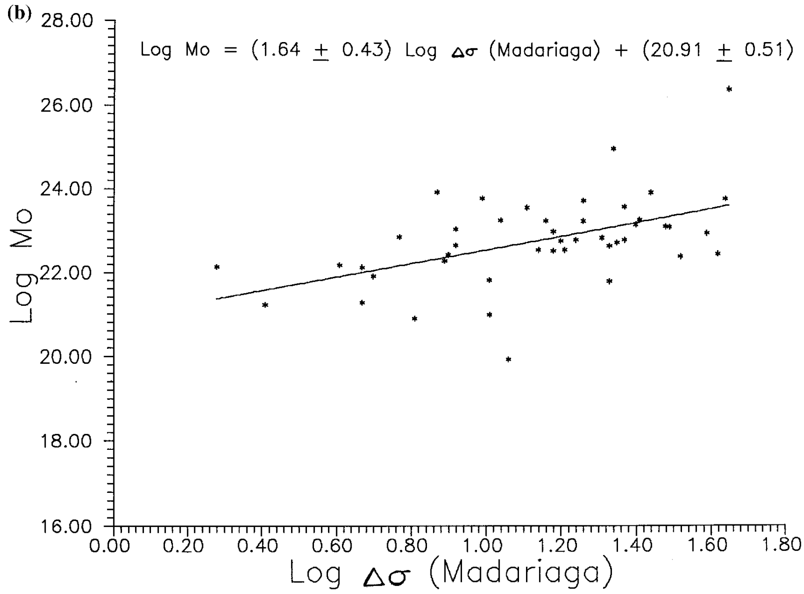


Figure 10

a. Log M_0 (seismic moment) versus Log $\Delta\sigma$ for the whole data set. $\Delta\sigma_B$ is the stress drop for the Brune model. b. Log M_0 (seismic moment) versus Log $\Delta\sigma$ for the whole data set. $\Delta\sigma_B$ is the stress drop for the Madariaga model.

References

AKI, K., and RICHARDS, P. G., *Quantitative Seismology: Theory and Methods* (Freeman, San Francisco, California 1980).

ARCHULETTA, R. J., GRANSWICK, E., MUELLER, C., and SPUDICH, P. (1982), *Source Parameters of the 1980 Mammoth Lakes, California, Earthquake Sequence*, J. Geophys. Res. *87*, 4595–4607.

BÅTH, M. (1983), *The Seismology of Greece*, Tectonophysics *98*, 165–208.

BERGER, J. (1985), *Low Stress-drop Earthquakes in the Light of New Data from the Anza, California, Telemetered Digital Array*, Maurice Ewing Series, Vol. 6, Am. Geophys. Union Geophys. Monograph *37*, 237–245.

BRUNE, J. N. (1968), *Seismic Moment, Seismicity and Rate of Slip along Major Fault Zones*, J. Geophys. Res. *73*, 777–794.

BRUNE, J. N. (1970), *Tectonic Stress and the Spectra of Seismic Shear Waves from Earthquakes*, J. Geophys. Res. *75*, 4997–5009.

BRUNE, J. N. (1971), *Correction*, J. Geophys. Res. *76*, 5002.

BRUNE, J. N., FLETCHER, J., VERNON, F., HAAR, L., HANKS, T., BURTON, P. W., MELIS, N. S., and BROOKS, M. (1995), *Coseismic Crustal Deformation on a Fault Zone Defined by Microseismicity in the Pavliani Area, Central Greece*, Geophys. J. Int. *123*, 16–40.

HANKS, T. C., and WYSS, M. (1972), *The Use of Body Wave Spectra in the Determination of Seismic Source Parameters*, Bull. Seismol. Soc. Am. *62*, 561–589.

HANKS, T. C., and BOORE, D. M. (1984), *Moment—Magnitude Relations in Theory and Practice*, J. Geophys. Res. *89*, 6229–6235.

HASHIDA, T., STAVRAKAKIS, G., and SHIMAZAKI, K. (1988), *Three-dimensional Seismic Attenuation Structure Beneath the Aegean Region and its Tectonic Implication*, Tectonophys. *145*, 43–54.

I.G.M.E (Institute of Geology and Minerology, Ministry of Energy) (1995), *Internal Report on the June 15, 1995, Aigio Earthquake*, pp. 1–56, Athens.

KEILIS-BOROK, V. I. (1959), *On the Estimation of the Displacement in an Earthquake Source and of Source Dimensions*, Ann. Geof. *12*, 205–214.

KIM, W. Y., KULHANEK, O., and MEYER, K. (1984), *Source Process of the 1981 Gulf of Corinth Earthquake Sequence from Body-wave Analysis*, Bull. Seismol. Soc. Am. *74*, 459–477.

KIRATZI, A. A., and PAPAZACHOS, B. C. (1984), *Magnitude Scales for Earthquakes in Greece*, Bull. Seismol. Soc. Am. *74*, 969–985.

KIRATZI, A. A., KARAKAISIS, G. F., PAPADIMITRIOU, E. E., and PAPAZACHOS, B. C. (1985), *Seismic Source-parameter Relations for Earthquakes in Greece*, Pure appl. geophys. *123*, 27–41.

KULHANEK, O., and MEYER, K. (1979), *Source Parameters of the Volvi-Langadhas Earthquake of June 20, 1978 Deduced from Body-wave Spectra at Station Uppsala and Kiruna*, Bull. Seismol. Soc. Am. *68*, 1289–1294.

MADARIAGA, R. (1976), *Dynamics of an Expanding Circular Fault*, Bull. Seismol. Soc. Am. *66*, 636–666.

MAIN, I. G., and BURTON, P. W. (1990), *Moment-magnitude Scaling in the Aegean Area*, Tectonophys. *179*, 273–285.

MARQUARDT, D. W. (1963), *An Algorithm for Least Squares Estimation of Nonlinear Parameters*, J. Soc. Ind. Appl. Math. *11*, 431–441.

MELIS, N. S., BURTON, P. W., and BROOKS, M. (1995), *Coseismic Crustal Deformation from Microseismicity in the Patras Area, Western Greece*, Geophys. J. Int. *122*, 815–836.

NORTH, R. G. (1973), *Seismic Source Parameters*, Ph. D. Thesis, Cambridge University.

NORTH, R. G. (1977), *Seismic Moment, Source dimensions, and Stresses Associated with Earthquakes in the Mediterranean and Middle East*, Geophys. J. R. Astron. Soc. *48*, 137–161.

PAPAZACHOS, B. C. (1994), *Seismology in Greece*, Europ. Seismol. Com. Proceedings of the XXIV Gen. Ass. *1*, 39–57, Athens, 19–24 September 1994.

PAVLIDES, S. B., ZOUROS, N. C., CHATZIPETROS, A. A., KOSTOPOULOS, D. S., and MOUNTRAKIS, D. M. (1995), *The May 13, 1995 Western Macedonia, Greece, Earthquake; Preliminary Results*, Terra Nova *7*, 544–549.

PRESS, F. (1967), *Dimensions of the Source Region for Small Shallow Earthquakes*, VESIAC *15*, (915), 155–163.

PRESS, W. H., FLANNERY, B. P., TEUKOLSKY, S. A., and VETTERLING, W. T., *Numerical Recipes in C* (Cambridge University Press, New York 1988).

SCHERBAUM, F. (1990), *Combined Inversion for the Three-dimensional Q Structure and Source Parameters using Microearthquake Spectra*, J. Geophys. Res. *95*, 12423–12438.

SCHERBAUM, F., and JOHNSTON, J. (1993), PITSA, IASPEI Software Library, Vol. 5, 1993.

SCHICK, R. (1970), *A Method for Determining Source Parameters for Small Magnitude Earthquakes*, Z. Geophys. *36*, 205–224.

SOUFLERIS, C., and STEWART, G. S. (1981), *A Source Study of the Thessaloniki (Northern Greece) 1978 Earthquake Sequence*, Geophys. J. R. Astr. Soc. *67*, 343–358.

STAVRAKAKIS, G. N., DRAKOPOULOS, J., and MAKROPOULOS, K. (1986), *A Rupture Model for the Corinth Earthquake Sequence*, Earthquake Prognostics, Proc. 2nd International Seminar, Berlin, June 24–27, 129–153.

STAVRAKAKIS, G. N., TSELENTIS, G. A., and DRAKOPOULOS, J. (1987), *Iterative Deconvolution of Teleseismic P waves from the Thessaloniki (Northern Greece) Earthquake of June 20, 1978*. Pure appl. geophys. *124*, 1039–1050.

STAVRAKAKIS, G. N., DRAKOPOULOS, J., LATOUSSAKIS, J, PAPANASTASIOU, D., and DRAKATOS, G. (1989), *Spectral Characteristics of the 1986 September 13 Kalamata (Southern Greece) Earthquake*, Geophys. J. Int. *98*, 149–157.

STAVRAKAKIS, G. N., and BLIONAS, S. V. (1990), *Source Parameters of Some Large Earthquakes in the Eastern Mediterranean Region Based on an Iterative Maximum Entropy Technique*, Pure appl. geophys., *132* (4), 679–698.

THATCHER, W., and HANKS, T. C. (1973), *Source Parameters of Southern California Earthquakes*, J. Geophys. Res. *78*, 8547–8576.

TSELENTIS, G. A., STAVRAKAKIS, G. N., MAKROPOULOS, K., LATOUSAKIS, J., and DRAKOPOULOS, J. (1988), *Seismic Moments of Earthquakes at the Western Hellenic Arc and their Application to the Seismic Hazard of the Area*, Tectonophys. *148*, 73–82.

(Received July 30, 1996, accepted, January 23, 1997)