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Spectral characteristics of the 1986 September 13 Kalamata (southern Greece) earthquake

G. N. Stavrakakis, J. Drakopoulos, J. Latoussakis, D. Papanastassiou and G. Drakatos
National Observatory of Athens, Seismological Institute, 118 10 *Athens, Greece*

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SUMMARY

The seismic source parameters seismic moment M_0, source dimension r, fault length l, average displacement $\langle u \rangle$, shear–stress drop $\Delta\sigma$, radiated energy E_S, and apparent stress $n\langle\sigma\rangle$ are calculated for the Kalamata (southern Greece) earthquake of 1986 September 13. These source parameters are obtained by using teleseismic P and SH-wave spectra and making use of three independent spectral parameters, the long period spectral level Ω_0, the corner frequency f_0, and the parameter ε which controls the high-frequency decay of the displacement spectrum. The calculated far-field spectra are based on the direct phases of P- and S-waves and they have been corrected for the instrumental response and for an average radiation pattern assuming normal faulting mechanism. The striking feature of the obtained results is the low stress drop value of about 6–7 bar which is in deviation from the average value (30 bar) for interplate earthquakes. A seismic moment of 1.9×10^{25} and 2.2×10^{24} dyne · cm, a fault length of about 22 and 11 km, and an average displacement of 15 and 9 cm are obtained on the basis of P- and S-waves, respectively.

Key words: seismic source parameters, earthquake, Greece

INTRODUCTION

On 1986 September 13, a moderate earthquake of magnitude $M_L = 5.5$ ($M_S = 6.0$, Athens; $M_S = 5.7$, Neis) occurred in the SW part of the Hellenic arc, about 10 km north of the city of Kalamata (Fig. 1). The shock caused considerable damage in the city, 20 people lost their lives and several hundred were injured.

The main shock was followed by only one large aftershock of magnitude $M_L = 4.9$ on September 15, which caused additional severe damage in the city. A large number of smaller aftershocks have been recorded by a portable network (MEQ-800) of 15 stations, installed in the epicentral area after the occurrence of the main shock by the National Observatory of Athens, in collaboration with the Seismological Laboratory of Athens University and the Institute de Physique du Globe de Paris. Fig. 1 shows the epicentral distribution of 500 well-located aftershocks with $M_L > 1.8$ during the first month after the main shock. Only aftershocks characterized by A and B solutions from HYPO 71 (Lee & Lahr 1976) computer program are adopted and plotted in Fig. 1.

Lyon-Caen *et al.* (1987) interpreted the main shock by normal faulting along a plane orientated about 30°N and dipping westward at about 45°, on the basis of teleseismic P- and SH-wave modelling. The same authors suggest that the Kalamata earthquake occurred on the eastern flank of the graben forming the Messinian Gulf (Fig. 1), in the southern part of the Peloponnesus which is actively extending in the east–west direction. Delibasis, Drakopoulos & Stavrakakis (1987b) based on P-wave first motions as reported by WWSSN stations, as well as on short period recordings from the National Network of Greece suggest also normal faulting along a plane striking N 173°E and dipping 50°SW.

However, direct *in situ* measurements of the horizontal stress (Paquin *et al.* 1982) yielded EW compression in the southern part of Peloponnesus. The main objective of the present paper is to study the seismic source parameters seismic moment M_0, source dimension r, shear–stress drop $\Delta\sigma$, radiated energy E_S, and apparent stress $n\langle\sigma\rangle$, by using the far-field displacement seismic source model proposed by Brune (1970, 1971). For this purpose, P- and SH-waveforms recorded at teleseismic distances by WWSSN long-period instruments have been processed.

REPRESENTATION OF SEISMIC SOURCE PARAMETERS

According to Brune's model (1970, 1971) the general features of the far-field displacement spectra are the long-period spectral level Ω_0, which is proportional to the seismic moment M_0, the spectral corner frequency f_0, which is proportional to the reciprocal of the radius r of a circular fault area, and a parameter ε which controls the high-frequency decay of the displacement spectrum.

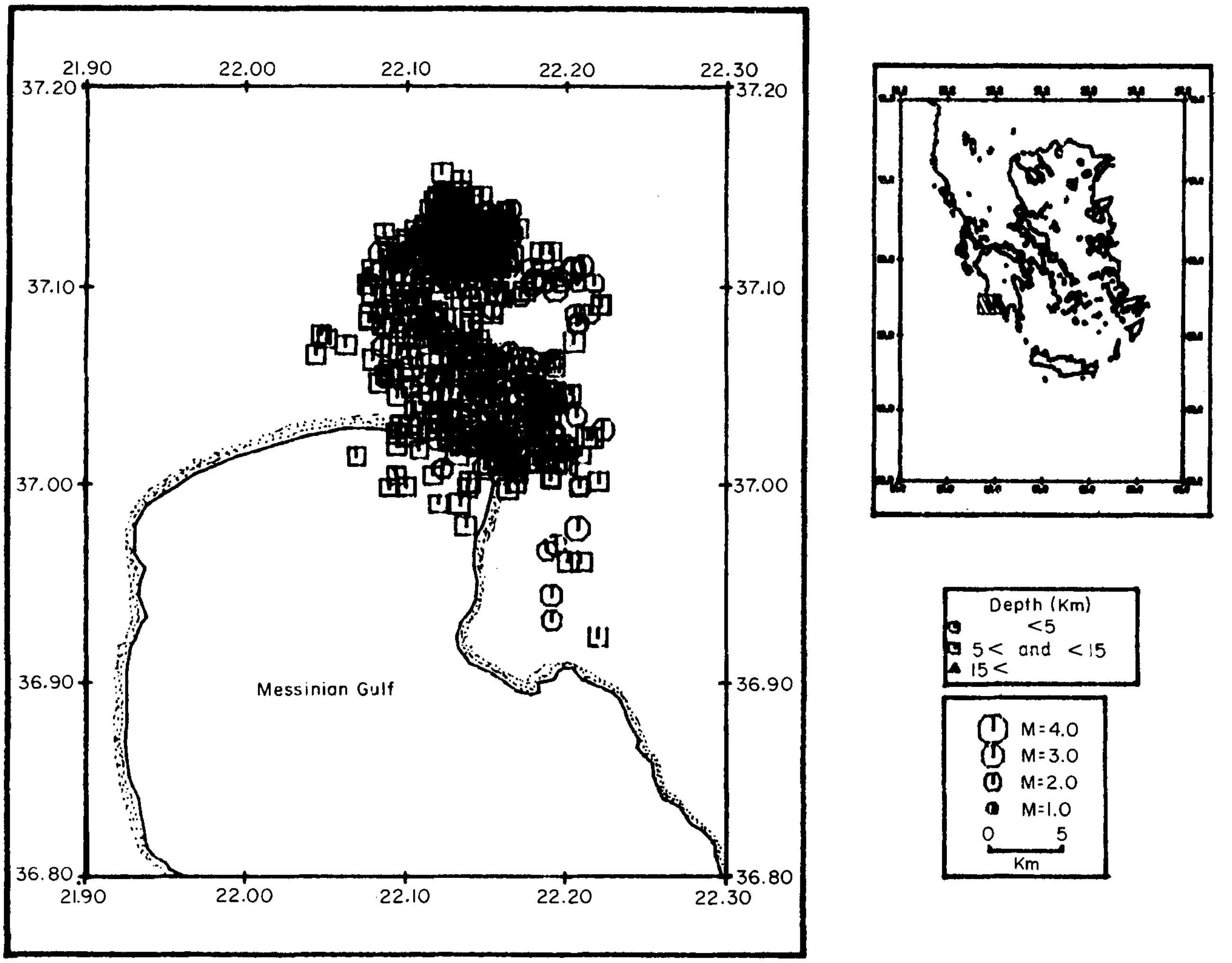

Figure 1. Distribution of aftershocks with $M_L > 1.8$ during the first month after the occurrence of the Kalamata earthquake of 1986 September 13.

The parameter ε is given by (Wyss & Molnar 1972; Hanks and Thatcher 1972)

$$\varepsilon = (\sigma_1 - \sigma_2)/(\sigma_1 - \sigma_f), \tag{1}$$

where σ_1 is the average shear–stress in the fault prior to the occurrence of earthquake, σ_2 is the average shear–stress after the earthquake, and σ_f is the average frictional stress opposing motion on the fault surface.

The shear–stress drop $\Delta\sigma$ is

$$\Delta\sigma = \sigma_1 - \sigma_2 \tag{2}$$

and the effective shear–stress is (Brune 1970)

$$\sigma_{\text{eff}} = \sigma_1 - \sigma_f. \tag{3}$$

In the present analysis we assume the case of complete stress drop, i.e. when $\varepsilon = 1$. In Brune's model enters only the difference $\sigma_1 - \sigma_f$, although σ_1 and σ_f are both independent and only this difference can be determined seismically (Wyss & Molnar 1972).

The seismic moment $M_0(S, P)$ is determined from the P- and S-wave spectrum according to Keilis-Borok's model (Keilis-Borok 1960)

$$M_0(S, P) = \Omega_0(S, P)4\pi\rho G(\Delta)(\alpha, \beta)^3 k/R_{if}(S, P), \tag{4}$$

where $\Omega_0(S, P)$ is the long-period spectral level of the S- or P-waves, ρ is the denisty, $G(\Delta)$ is the geometrical spreading factor, α and β are the compressional and shear wave velocities, respectively, k is a factor to account for the free-surface reflection at the recording station, and R is the radiation pattern for S- or P-waves.

The geometrical spreading factor $G(\Delta)$ is given by

$$G(\Delta) = R(\cos i_0 \sin \Delta/\sin i_h)^{1/2}, \tag{5}$$

where $R = 6317$ km the radius of the earth, i_0, i_h are the incidence and take-off angles, respectively, and Δ is the epicentral distance in degrees.

The radius r of a circular fault area is related to the spectral corner frequency f_0 through the relationship

$$r = 2.346/2\pi f_0 \tag{6}$$

(corrected from Brune (1970) by Brune (1971)) and the average displacement $\langle u \rangle$ is then

$$\langle u \rangle = M_0(P)/\mu\pi r^2(P), \tag{7}$$

where $\mu = 3 \times 10^{11}$ dyne cm^{-2} is the shear modulus. When $\varepsilon = 1$, then the shear effective stress σ_{eff} is (Hanks &

Thatcher 1972)

$$\sigma_{\text{eff}} = 106\rho G(\Delta)(\Omega_0 f_0^3)/\varepsilon \tag{8}$$

and for the case of complete stress drop ($\varepsilon = 1$) the shear–stress drop is given by

$$\Delta\sigma = 7M_0/16r^3. \tag{9}$$

The energy E_S radiated in the form of the S-wave is expressed in terms of its spectral amplitude $\Omega(f)$

$$E_S = 1/(2\pi)I_S\rho\beta G(\Delta)^2\,|\Omega(\omega)\omega|^2\,d\omega, \tag{10}$$

where ω is the circular frequency and I_S is a term resulting from the integration of the S-wave radiation pattern about the source. Wu (1966) assumed $I_S = 24\pi/15$ and obtained the following expression to E_S

$$E_S = (128\pi/15)\rho\beta G^2(\Delta)\Omega_0^2 f_0^3 \tag{11}$$

for the case of complete stress drop ($\varepsilon = 1$).

The apparent stress drop $n\langle\sigma\rangle$ is given by (Wyss 1970)

$$n\langle\sigma\rangle = \mu(E_S/M_0), \tag{12}$$

where n is the seismic efficiency, $\langle\sigma\rangle = (\sigma_1 + \sigma_2)/2$. In the case of complete stress drop ($\varepsilon = 1$) the apparent stress is (Hanks & Thatcher 1972)

$$n\langle\sigma\rangle = 21.1\rho G(\Delta)\Omega_0 f_0^a. \tag{13}$$

INSTRUMENTAL DATA AND ANALYSIS

Long-period seismograms recorded at WWSSN stations in the epicentral distance range 35°–80° have been selected to obtain the far-field displacement spectra. Information on the stations used is summarized in Table 1, where Δ is the epicentral distance in degrees, AZE is the epicentral azimuth, AZS is the back-azimuth, i_0 is the emergent angle, i_h is the incidence angle, and $G(\Delta)$ is the geometrical spreading factor. The selected seismograms are manually digitized with a sampling frequency of 3 Hz. Since the spectra are affected by the signal truncation, several tests are performed to obtain reliable estimates of the flat part of the spectrum. The number of points used in FFT varies between 64 and 256. However, due to the shallow focus ($h = 10$ km), all portions include reflective phases and no attempt has been made to introduce corrections of pP-arrivals. Kulhanek & Meyer (1979) calculated also source spectra for the Thessaloniki (northern Greece) earthquake of 1978 June 20, without introducing any correction for the reflective phases. Wyss & Hanks (1972) analysed the 1971 February 9, San Fernando, California, earthquake and proposed that the P- and S-waves may be expected to overestimate the seismic moment because the surface reflections pP and sS could not be excluded in the analysis.

Table 1. Station information and ray parameters for the Kalamata earthquake of 1986 September 13.

STATION CODE	Δ (°)	AZE (°)	AZS (°)	i_0 (°)	i_h (°)	$G(\Delta)$ km
HKC	78.28	70.80	305.4	18.7	22.6	10697
KOD	56.32	103.6	307.9	24.9	30.2	8326
GRM	70.45	176.1	356.3	20.8	25.1	9876
SLR	63.03	173.8	354.5	22.9	27.7	9072
WES	68.30	307.4	59.0	21.4	25.9	9642
WIN	59.80	185.5	4.6	23.9	28.9	8706
AKU	36.88	333.70	120.8	30.1	36.8	6236
QUE	37.55	87.03	292.74	29.9	36.5	6315

All spectra presented here have been corrected for instrumental response. Using the fault plane solution as obtained by Lyon-Caen *et al.* (1987) (strike = 201°, dip = 47°, rake = −80°), the radiation pattern $R_{\text{fi}}(P)$ and $R_{\text{fi}}(S)$ is calculated according to the formula of Ben-Menahem, Smith & Teng (1965). Some researchers (Kulhanek & Meyer 1979; Thatcher 1972; Hanks & Wyss 1972; Wyss & Hanks 1972; Kulhanek *et al.* 1983) have calculated spectra corrected for anelastic attenuation along the propagation path by using curves from Julian & Anderson (1986) or by taking average Q-values according to Anderson, Ben-Menahem & Archambeau (1965) and Ibrahim (1971). We preferred, as done by Modiano & Hatzfeld (1982), to measure spectral parameters on spectra uncorrected for attenutation, because Q varies with depth and possibly with frequency (Aki & Chouet, 1975; Tsujira 1978; Aki 1980). We believe that taking account of propagation path attenuation correctly is complicated, and the spectral corner frequency f_0 as well as the long-period level Ω_0 does not change very much. To verify this fact, we computed source spectra for some stations by using the following average $Q = 500$ for $\Delta < 40°$, $Q = 1000$ for $40° < \Delta < 60°$ and $Q = 1500$ for $\Delta > 60°$ according to Anderson *et al.* (1965) and Ibrahim (1971). Figs 2 and 3 show the obtained P- and SH-source spectra, respectively, corrected for attenuation and Figs 4 and 5 illustrate the spectra without correcting for attenuation. Comparing these spectra it is evident that the spectral level and the corner frequency do not change very much. However, the change in the shape is significant for frequencies higher than 0.3 Hz.

Taking a density $\rho = 3.0\,\text{gr cm}^{-3}$, a P-wave velocity $\alpha = 6.0\,\text{km s}^{-1}$ (Delibasis, Makris & Drakopoulos 1987a), a $V_p/V_S = 1.73$, a free-surface reflection factor $k = 0.5$, and by using the previous equations, the seismic source parameters for the Kalamata earthquake of 1986 September 13 are obtained. Tables 2 and 3 summarize the obtained results on the basis of P- and SH-wave spectra, respectively.

The average values of seismic moment and source dimension as well as the error factors are obtained following Archuleta *et al.* (1982). The seismic moment is taken as a geometric mean of the values at each station. A simple arithmetic mean produces averages biased to higher values, because the errors associated with the interpretation of Ω_0 and f_0 on a log-log plot are lognormally distributed. The geometric mean is taken as

$$\log\langle M_0\rangle = 1/N \sum \log M_{0i}$$

where N is the number of stations used in the analysis. The standard deviation of $\log M_0$ is given by

$$\text{s.d.}\log\langle M_0\rangle = [1/(N-1)\sum(\log M_{0i} - \log\langle M_0\rangle)^2]^{1/2}.$$

To be consistent with seismic moment, we computed a multiplicative error factor EM_0

$$EM_0 = \text{antilog}\,[\text{s.d.}(\log\langle M_0\rangle)].$$

In a manner exactly analogous to the computation of

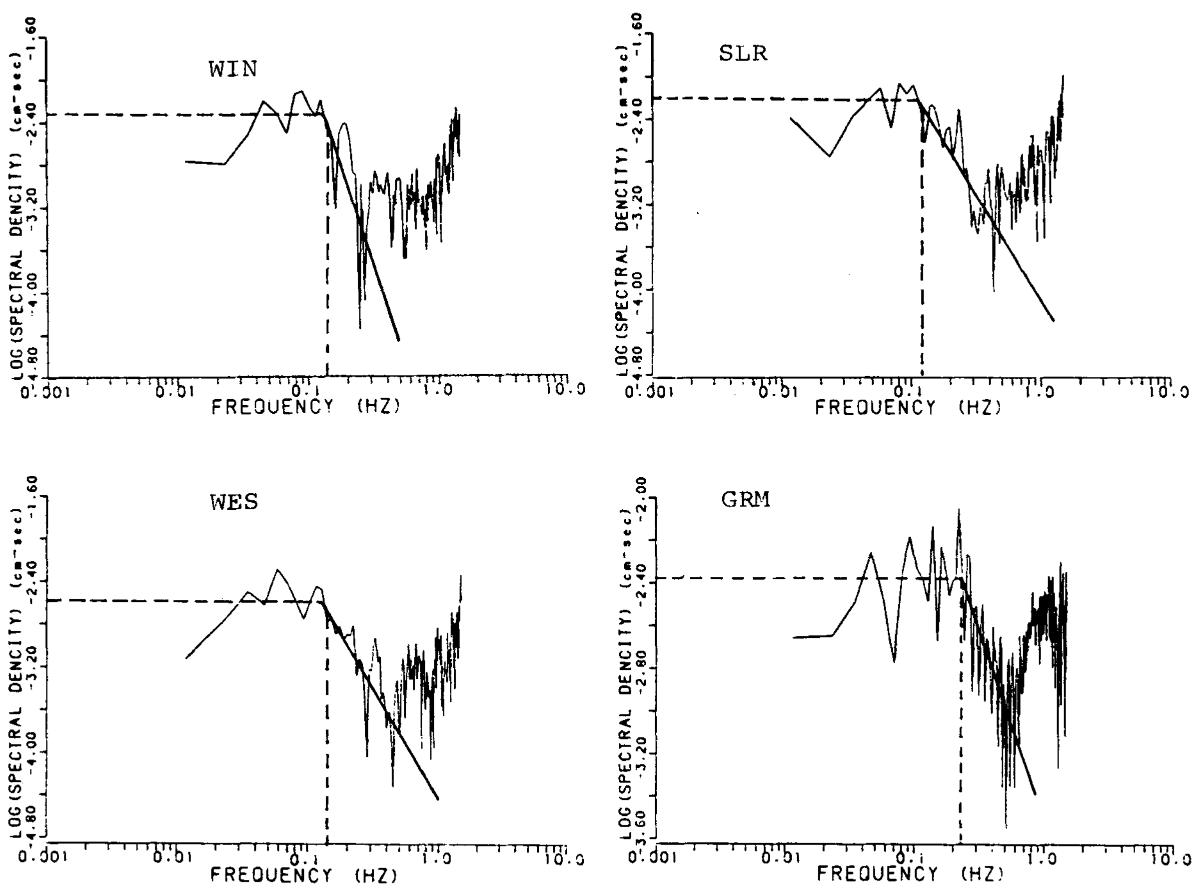

Figure 2. FFT far-field P-wave source spectra for the Kalamata earthquake of 1986 September 13, corrected for instrumental response, radiation pattern, and for attenuation along the propagation path.

average moment $\langle M_0 \rangle$ and EM_0 we computed the average source radius and its error factor Er. The obtained values are shown in Tables 2 and 3 for P- and S-waves, respectively.

DISCUSSION AND CONCLUSIONS

In the present study, far-field displacement spectra of body waves are employed to estimate the seismic source parameters seismic moment M_0, source dimension r, average displacement $\langle u \rangle$, shear stress drop $\Delta\sigma$, seismic energy E_s, and apparent stress $n\langle \sigma \rangle$ for the Kalamata earthquake of 1986 September 13 by using Brune's model. For this model the fault length is assumed to be equal to $2r$ (circular fault). The following average values are obtained from P- and SH-wave spectra which are computed by a Fast Fourier Transform routine.

	From P-waves	From SH-waves
Seismic moment	$\langle M_0 \rangle = 1.99 \times 10^{25}$	2.2×10^{24} dyne · cm
Stress drop	$\langle \Delta\sigma \rangle = 6$	7 bar
Fault length	$\langle L \rangle = 22$	11 km
Displacement	$\langle u \rangle = 15$	8.7 cm
Seismic energy	$\langle E_s \rangle = 1.7 \times 10^{19}$	24.0×10^{19} erg
Apparent stress	$n\langle \sigma \rangle = 1.4$	1.8 bar

The basic assumptions of this study are that, first the far-field displacement spectra as given by Brune (1970, 1971) are sufficient to describe the far-field radiation, and secondly that the physical interpretation of the spectral parameters as given by Brune is correct. The calculated spectra present a constant level at low frequencies, a linear decrease at high frequencies and a corner frequency which can be explained by all dislocation models (Aki 1967; Archambeau 1968).

The most critical point in the application of Brune's model is the determination of the shear–stress drop. For the Kalamata earthquake a low value of about 6 and 7 bar is obtained when P- or SH-spectra are used, respectively. The low stress drop even can be explained by the partial stress drop model. According to Brune *et al.* (1985), partial stress events might occur when the stress release is not uniform over the fault plane, but rather is more like a series of multiple events with parts of the fault remaining unbroken. This heterogeneity could occur as a result of asperities or barriers on the fault. Thus, the average displacement and the stress drop over the fault plane area will be less than the average displacement and stress drop over the individual subevents.

If Madariaga's model (1976) had been used to calculate the radius of the circular fault, according to the relationship $r(P) = 0.32\beta/f_0(P)$ with $\beta = 3.33\ \mathrm{km\ s^{-1}}$, an average corner frequency and seismic moment 0.195 Hz and 2.6×10^{25} dyne · cm, respectively, then a high stress drop of about 75 bar would have been obtained.

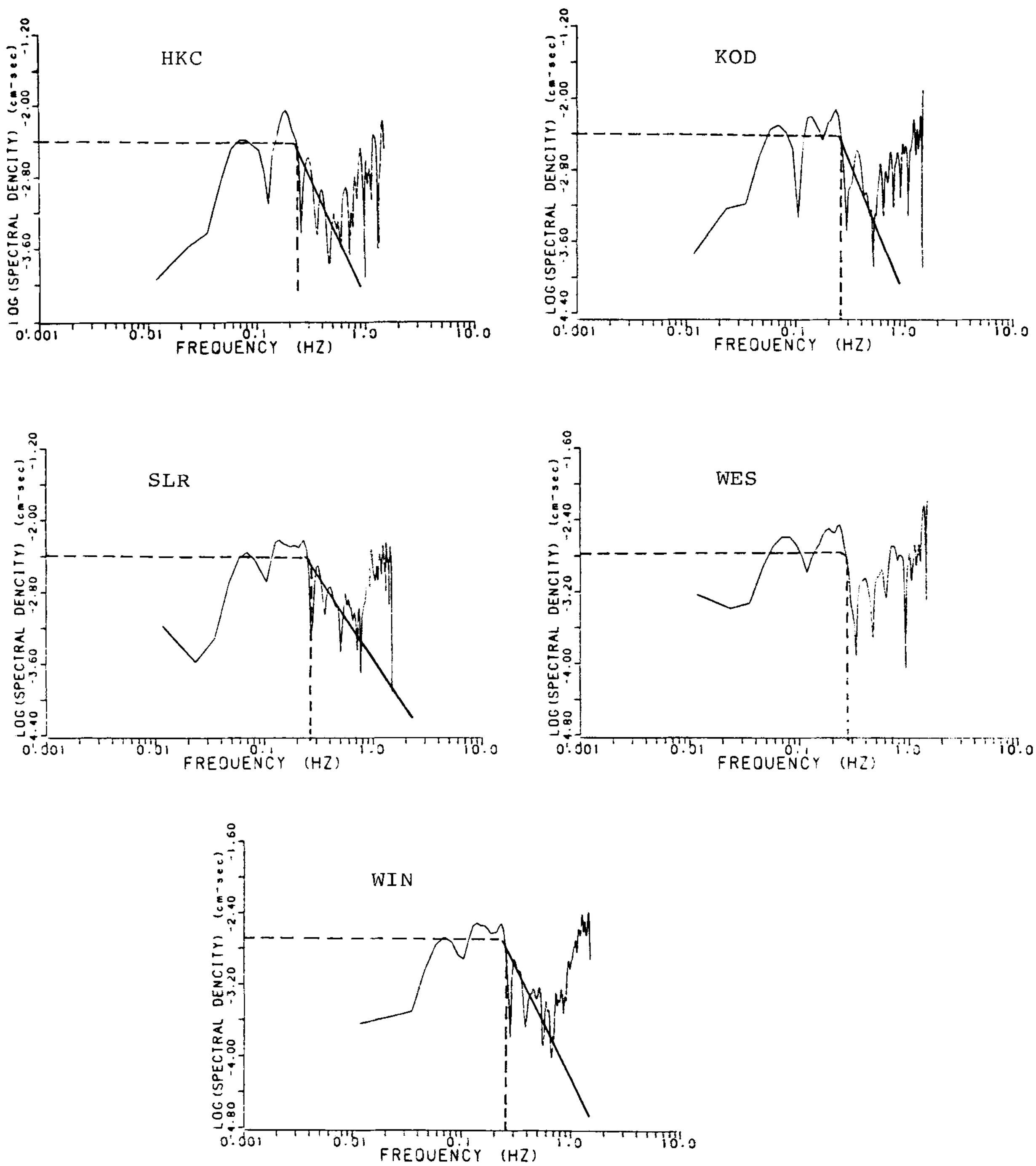

Figure 3. FFT far-field SH-wave source spectra for the Kalamata earthquake of 1986 September 13, corrected for instrumental response, radiation pattern, and for attenuation along the propagation path.

It should be emphasized at this point, that low stress drop has also been revealed for other earthquakes which occurred in different seismotectonic areas in Greece by using Brune's model. Kulhanek & Meyer (1979) obtained a stress drop of 6.6 bar for the Thessaloniki (northern Greece) earthquake ($M_s = 6.4$) of 1978 June 20. Kulhanek & Meyer (1984) obtained a stress drop of 10.2 bar for the main shock ($M_s = 6.7$) of the Corinth (central Greece) earthquake of 1981 February 24 and a value of 8.0 and 6.0 bar for the largest aftershocks of February 25 ($M_s = 6.4$) and March 4 ($M_s = 6.4$), respectively. Higher values of about 60, 30 and 15 bar are obtained (Stavrakakis, Drakopoulos & Makropoulos 1986) for the main shock and its principal aftershocks of the Corinth earthquake sequence on the basis of the total duration of the far-field source time function according to the Geller's model (1976). However, it should be emphasized that for the Thessaloniki earthquake as well as for the Corinth earthquake sequence the multiple source mechanism has been resolved (Stavrakakis, Tselentis & Drakopoulos 1987), and the partial low stress drop model

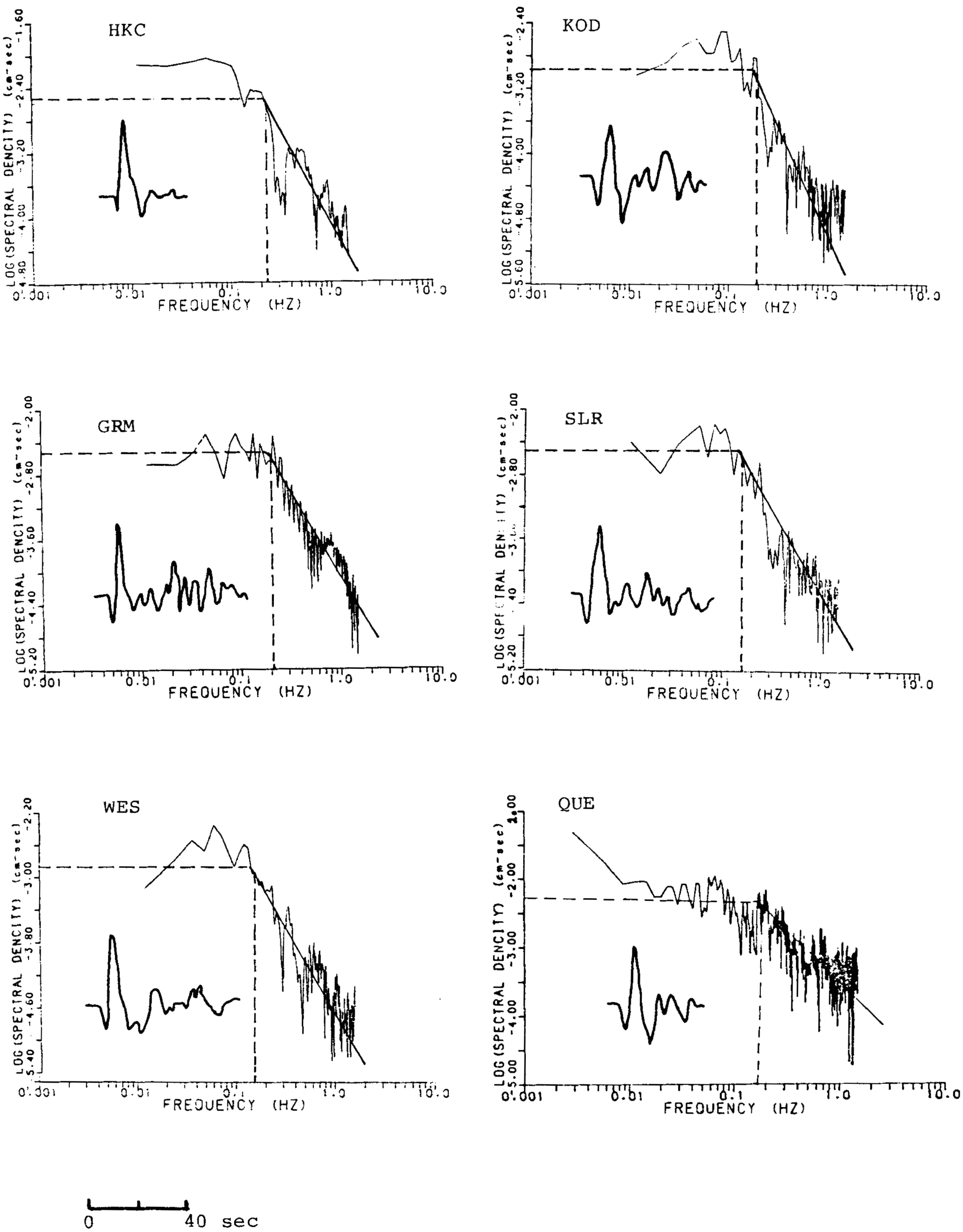

Figure 4. FFT far-field P-wave source spectra for the Kalamata earthquake of 1986 September 13, corrected only for instrumental response and for radiation pattern.

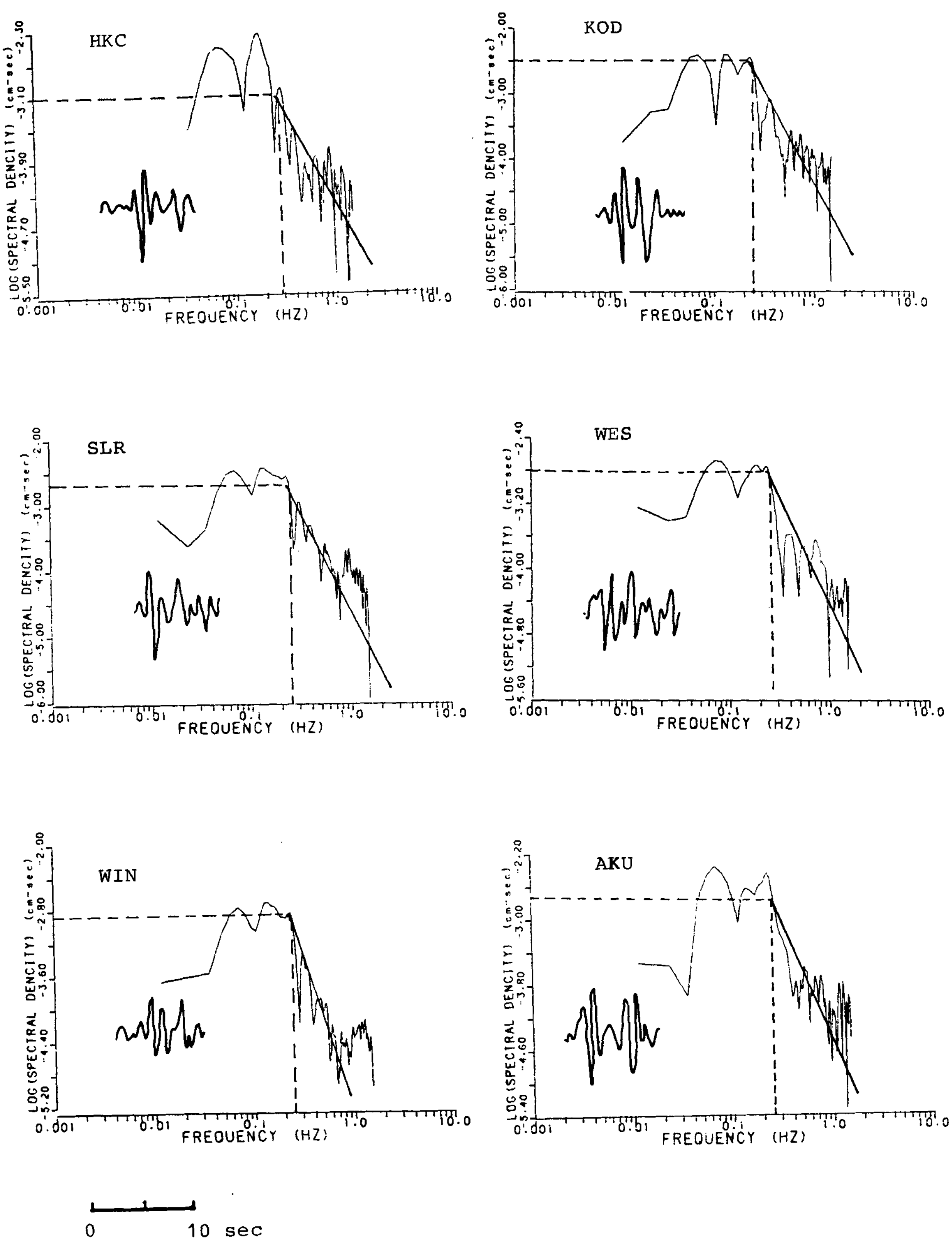

Figure 5. FFT far-field SH-wave source spectra for the Kalamata earthquake of 1986 September 13, corrected only for the instrumental response and radiation pattern.

Table 2. Spectral characteristics and source parameters, based on far-field P-wave spectra of the Kalamata earthquake of 1986 September 13.

STATION CODE	f_0 (Hz)	$\Omega_0(P)$ cm-sec	$M_0(P)\times10^{25}$ dyne-cm	$r(P)$ km	$E_s\times10^{19}$ ergs	$n\langle\sigma\rangle$ bars
HKC	0.22	0.0030	2.56	10.15	3.04	2.2
KOD	0.20	0.0016	1.06	11.17	0.39	0.7
GRM	0.20	0.0032	2.52	11.17	2.22	1.6
SLR	0.20	0.0032	2.31	11.17	1.87	1.5
WES	0.17	0.0014	1.07	13.14	0.25	1.2
QUE	0.18	0.0059	2.97	12.41	2.25	1.3

$\log\langle M_0\rangle = 25.28$ $\quad$ $\log\langle r\rangle = 1.059$
$\langle M_0\rangle = 1.9\times10^{25}$ dyne cm $\quad$ $\langle r\rangle = 11.46$ km
s.d.$(\log\langle M_0\rangle) = 0.20$ $\quad$ s.d.$(\log\langle r\rangle) = 0.041$
$EM_0 = 1.58$ $\quad$ $Er = 1.09$
$\langle\Delta\sigma\rangle = 6$ bars
$\langle u\rangle = 15$ cm

Table 3. Spectral characteristics and source parameters, based on far-field S-wave spectra of the Kalamata earthquake of 1986 September 13.

STATION CODE	$f_0(S)$ Hz	$\Omega_0(S)$ cm-sec	$M_0(S)\times10^{24}$ dyne-cm	$r(S)$ km	$E_s\times10^{19}$ ergs	$n\langle\sigma\rangle$ bars
HKC	0.25	0.0010	1.39	5.22	7.4	1.0
KOD	0.25	0.0032	3.47	5.22	46.22	2.6
SLR	0.25	0.0031	3.66	5.22	51.49	2.8
WES	0.26	0.0016	2.01	5.02	17.43	1.7
WIN	0.25	0.0016	1.81	5.22	12.63	1.4
AKU	0.25	0.0019	1.54	5.22	9.14	1.2

$\log\langle M_0\rangle = 24.33$ $\quad$ $\log\langle r\rangle = 0.714$
$\langle M_0\rangle = 2.2\times10^{24}$ dyne cm $\quad$ $\langle r\rangle = 5.18$ km
s.d.$(\log\langle M_0\rangle) = 0.19$ $\quad$ s.d.$(\log\langle r\rangle) = 0.007$
$EM_0 = 1.53$ $\quad$ $Er = 1.01$
$\langle\Delta\sigma\rangle = 7$ bars
$\langle u\rangle = 8.7$ cm

seems to be realistic for some seismogenic regions in Greece.

Low stress drop values (between 1 to 30 bar) for earthquakes in Greece have also been obtained by Kiratzi *et al.* (1985). They explained the low stress drop on the basis of Archambeau's model (1978), which relates the low stress drop earthquakes with the ratio of the body- and surface-magnitude. In reality, Kiratzi & Papazachos (1984) found that for a given surface magnitude the body-wave magnitude assigned to the earthquakes of Greece were considerably smaller than those of other regions. Since the body-wave and surface magnitudes are proportional to the amplitude of approximately 1 and 20 s, respectively, the low stress drop might be explained on the basis of the frequency content of earthquakes.

The observed surface faulting is of about 10 km and the coseismic displacement varies between 6 and 18 cm. For the seismic moment we obtained a value of 2.6×10^{25} and 5.4×10^{24} dyne · cm on the basis of P- and SH-wave spectra, respectively.

A discrepancy exists between the average fault length obtained in this study which is almost twice the corresponding value estimated by Lyon-Caen *et al.* (1987). The observed coseismic displacement varies between 6 and 18 cm. However, the inferred fault length is in agreement with that obtained from the aftershock distribution (Fig. 1). The seismic energy is smaller than that obtained by using the Gutenberg–Richter relationship ($E_s = 1.25\times10^{21}$ erg). This seems to enhance the observation (Wyss & Molnar 1972) that the Gutenberg energy may be a poor estimate of the true energy and also that the Gutenberg energy overestimates the true energy by factors of between 10 and 100.

Lastly, the question as to whether the low stress drop earthquakes which occurred in the area of Greece are a source-model consequence or a characteristic feature of the area must await more data.

REFERENCES

Anderson, D. L., Ben-Menahem, A. & Archambeau, C. B., 1965. Attenuation of seismic energy in the upper mantle, *J. geophys. Res.*, **70**, 1441–1448.

Aki, K., 1967. Scaling law of seismic spectrum, *J. geophys. Res.*, **72**, 1217–1231.

Aki, K. & Chouet, B., 1975. Origin of coda waves: source, attenuation and scattering effects, *J. geophys. Res.*, **80**, 3322–3342.

Aki, K., 1980. Attenuation of shear-waves in the lithosphere for frequencies from 0.05 to 25 Hz, *Phys. Earth planet. Int.*, **21**, 50–60.

Archambeau, C. B., 1968. General theory of elastodynamic source fields, *Rev. geophys. Space Phys.*, **6**, 241–287.

Archambeau, C., 1978. Estimation of non-hydrostatic stress in the earth by seismic methods: Lithospheric stress levels along Pacific and Nazca Plate subduction zones, *US geol. Surv. Open-File Report 78-943*, 47–138.

Archuleta, R. J., Cranswick, E., Mueller, C. & Spudich, P., 1982. Source parameters of the 1980 Mammoth Lakes, California, earthquake sequence, *J. geophys. Res.*, **87**, 4595–4607.

Ben-Menahem, A., Smith, S. W. & Teng, T. L., 1965. A procedure for source studies from spectra of long-period seismic waves, *Bull. seism. Soc. Am.*, **45**, 203–235.

Brune, J. N., 1970. Tectonic stress and spectra of seismic shear waves, *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. & 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 Geophysics Monograph 37*, pp. 237–245, Washington, DC.

Delibasis, N., Makris, J. & Drakopoulos, J., 1987a. Crustal and upper mantle structure of western Greece, *Ann. Geol. Pays Hellen.*, **35.**

Delibasis, N., Drakopoulos, J. & Stavrakakis, G., 1987b. The Kalamata (s. Greece) earthquake of 13 September 1986, *Ann. Geophysicae*, **5B**, 731–734.

Geller, R. J., 1976. Scaling relations for earthquake source parameters and magnitudes, *Bull. seism. Soc. Am.*, **66**, 1501–1523.

Hanks, T. C. & Thatcher, W., 1972. A graphical representation of seismic source parameters, *J. geophys. Res.*, **77**, 4393–4405.

Hanks, T. C. & Wyss, M., 1972. The use of body-wave spectra in the determination of seismic-source parameters, *Bull. seism. Soc. Am.*, **62**, 561–589.

Ibrahim, A-B. K., 1971. The amplitude ratio PcP/P and the core–mantle boundary, *Pure appl. Geophys.*, **91**, 114–133.

Julian, B. R. & Anderson, D. L., 1968. Travel times, apparent velocities, and amplitudes of body waves, *Bull. seism. Soc. Am.*, **58**, 339–366.

Kanamori, H. & Anderson, D. L., 1975. Theoretical basis of some empirical relations in seismology, *Bull. seism. Soc. Am.*, **65**, 1073–1095.

Keilis-Borok, V. I., 1960. Investigation of the mechanism of earthquakes, *Soviet Res. Geophys.* (English transl.), **4**, 29.

Kim, W.-Y., Kulhanek, O. & Meyer, K., 1984. Source process of the 1981 Gulf of Corinth earthquake sequence from body-wave analysis, *Bull. seism. Soc. Am.*, **74**, 459–477.

Kiratzi, A. A. & Papazachos, B. C., 1984. Magnitude scales for earthquakes in Greece, *Bull. seism. Soc. Am.*, **74,** 969–985.

Kiratzi, A. A., Karakaisis, G. F., Papadimitriou, E. E. & Papazachos, B. C., 1985. Seismic source-parameter relations for earthquakes in Greece, *Pageoph*, **123,** 27–41.

Kulhanek, O. & Meyer, K., 1979. Source parameters of the Volvi-Langadhas earthquake of June 20, 1978 deduced from body-wave spectra at station Uppsala and Kituna, *Bull. seism. Soc. Am.*, **68,** 1289–1294.

Kulhanek, O., Eck, T. V., John, N., Meyer, K. & Wahlstrom, R., 1983. Spectra of the earthquake sequence February–March, 1981, in south-central Sweden, *Tectonophysics*, **93,** 337–350.

Lee, W. H. K. & Lahr, J. C., 1975. HYPO 71 (revised) computer program for determining hypocenter, magnitude and first motion pattern of local earthquakes, U.S.G.S. Open File Rep. 75-311.

Lyon-Caen, H., Drakopoulos, J., Armijo, R., Baskoutas, J., Delibasis, N., Gaulon, R., Kouskouna, V., Latoussakis, J., Makropoulos, K., Papastamatiou, P., Papanastasiou, D. & Pedotti, G., 1987. The September 13, 1986 Kalamata (Peloponnesus) earthquake, *EUG, Abstracts, XII General Assembly*, 9–14 April 1987, Strasbourg.

Madariaga, R., 1976. Dynamics of an expanding circular fault, *Bull. seism. Soc. Am.*, **66,** 639–666.

Modiano, T. & Hatzfeld, D., 1982. Experimental study of the spectral content for shallow earthquakes, *Bull. seism. Soc. Am.*, **72,** 1739–1758.

Paquin, C., Froidevaux, C., Bloyet, Y., Ricard, Y. & Angelidis, C., 1982. Tectonic stresses on the mainland of Greece: *in-situ* measurements by overcoring, *Tectonophysics*, **86,** 17–26.

Stavrakakis, G., Tselentis, G.-A. & Drakopoulos, J., 1987. Iterative deconvolution of teleseismic P-waves from the Thessaloniki (northern Greece) earthquake of June 20, 1978. *Pageoph*, **124,** 1039–1050.

Stavrakakis, G., Drakopoulos, J. & Makropoulos, K., 1986. A rupture model for the Corinth earthquake sequence of 1981. *Earthquake Prognostics Seminar*, 24–27 June 1986, Berlin, pp. 129–153.

Thatcher, W., 1972. Regional variations of seismic source parameters in the northern Baja California area, *J. geophys. Res.*, **77,** 1549–1565.

Tsujira, M., 1978. Spectral analysis of the coda waves of local earthquakes, *Bull. seism. Soc. Am.*, **53,** 1–48.

Wu, F. T., 1966. Lower limit of the total energy of earthquakes and partitioning of the energy among seismic waves, *Ph.D. thesis*, California Institute of Technology, Pasadena.

Wyss, M. & Molnar, P., 1972. Efficiency, stress drop, apparent stress, effective stress, and frictional stress of Denver, Colorado, earthquakes, *J. geophys. Res.*, **77,** 1433–1437.