SOURCE PARAMETERS FOR EARTHQUAKES IN GREECE AS INFERRED FROM INVERSION AND SPECTRAL ANALYSIS OF TELESEISMIC BODY WAVES

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A B S T R A C T

Source parameters, such as source dimension, stress drop, effective tectonic stress, dynamic energy release, have been estimated using the source-time functions and seismic moments obtained by inversion of body waves of some large and moderate earthquakes in Greece. The computations were mainly based on the characteristic time and total duration of the source time function.

Earthquakes that occurred in Ionian Sea (western Greece) and in the southern part of the Hellenic arc are characterized by low stress drop, whereas earthquakes in the Gulf of Corinth and in Aegean Sea show relatively higher values. However, in all cases the obtained stress drop values are in deviation of the "constant stress drop model". An attempt is made to explain this fact in terms of the "partial stress model" proposed by Brune et al. (1985).

ΠΑΡΑΜΕΤΡΟΙ ΤΗΣ ΠΗΓΗΣ ΓΙΑ ΣΕΙΣΜΟΥΣ ΤΗΣ ΕΛΛΑΔΑΣ ΟΠΩΣ ΠΡΟΚΥΠΤΟΥΝ ΑΠΟ ΑΝΑΣΤΡΟΦΗ ΚΑΙ ΦΑΣΜΑΤΙΚΗ ΑΝΑΛΥΣΗ ΚΥΜΑΤΩΝ ΧΩΡΟΥ ΜΕΓΑΛΩΝ ΑΠΟΣΤΑΣΕΩΝ.

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Π Ε Ρ Ι Λ Η Ψ Η

Παράμετροι της σεισμικής πηγής, όπως διαστάσεις της πηγής, πτώση τάσης, ενεργός τάση, δυναμική ενέργεια έχουν υπολογιστεί χρησιμοποιώντας τις χρονικές συναρτήσεις της πηγής και τις σεισμικές ροπές που υπολογίστηκαν από αναστροφή κυμάτων χώρου μερικών ισχυρών και ενδιαμέσου μεγέθους σεισμών σε διαφορετικές περιοχές στην Ελλάδα. Οι υπολογισμοί είναι κύρια βασισμένοι στη διάρκεια της χρονικής πηγής. Σε όλες τις περιπτώσεις οι τιμές της πτώσης τάσης αποκλείνουν από το "μοντέλο σταθερής πτώσης τάσης". Στην εργασία αυτή γίνεατι προσπάθεια να εξηγηθεί το φαινόμενο αυτό με βάση το μοντέλο "μερικής τάσης" που προτάθηκε από τον Brune και τους συνεργάτες του (1985).

INTRODUCTION

It is generally assumed that ground motion results from unstable slip accompanying a sudden drop in shear stress on a geological fault. Seismic radiation is the main source of information on the faulting processes of earthquake phenomena. The radiation can be computed by a space-time convolution of a slip function with a Green's function. The slip function describes the fault displacement during an earthquake as a function of time and position on the fault plane, while Green's function represents the Earth response to the slip.

One of the most important development in the understanding of the earthquake source in the last years is the documentation of the complexity of earthquake rupture. This complexity can be due to geometrical complexities of the fault plane or heterogeneities in the fault strength or tectonic stress. Several studies suggest that small areas of high stress drop are embedded in larger, low stress drop areas.

The high-frequency radiation comes from these small areas, although they contribute only a fraction of the total seismic moment. The most important source parameters for the dynamic process of fault motion are the effective tectonic stress and the fracture length. The former one is defined as the static frictional stress minus the sliding frictional stress. The later one is defined as the static frictional stress minus the initial stress (YAMASHITA, 1977). Although the effective tectonic stress is the simplest gross parameter, its determination has a dual importance in seismology. First, the effective tectonic stress with the stress drop determines how much of the effective stress is released, as a stress drop during an earthquake. On the other hand, the stress drop in the main earthquake determines the principal characteristics of the aftershock sequences. Gibowicz (1973) has shown that a low stress drop mainshock leads to high magnitude of the largest aftershock and conversely. From sophisticated waveform modeling using far-field source time functions it is possible to compute the source parameters as well as the effective tectonic stress.

In the present study, we use the source-time functions obtained from inversion of long-period teleseismic body waves of some large and moderate earthquakes in Greece to compute the source parameters. A comparison is also made between these parameters and those revealed using spectral theories.

SYNTHETIC SEISMOGRAMS

Several methods have been proposed to compute synthetic seismograms of body-and surface waves. The most common one is based on a trial-and-error fit of synthetic seismograms calculated for a point shear dislocation source to observe P and S waveforms (e.g. Helmberger, 1974; Langston and Helmberger, 1976). To compute synthetic seismograms, the direct P and two free- surface reflected rays, pP and sP, are considered for the P wave, and the direct S and free- surface reflected sS are also considered for the SH wave. Then, the synthetic seismograms for P waves are calculated by a series of convolution (*) operations according to the formula: $W=R_{pz}\ (S*\ T*\ Q*\ I)$, where W is the vertical displacement at the receiver, R_{pz} is the receiver function, S is the response of the source structure to a point double couple of specified orientation, T is the source-time function, Q is the Earth's attenuation, and I is the instrument response. The procedure used to compute S is the modification of the Thomson- Haskell layer matrix method by Harkrider and Anderson (1966). The details of the method are summarized in Langston (1976). Synthetic seismograms for SH waves are generated in the same way using the appropriate shear-wave receiver function, source structure response, and attenuation operator.

Another usefull method for computing synthetic seismograms is that proposed by Kikuchi and Kanamori (1982). In this method, body waves are deconvolved into a multiple shock sequence. The fault plane is divided into N x M discrete points in the strike and dip directions in order to compute Green's functions al all possible locations. Then, for each subevent, a set of parameters (m_i, X_i, Y_i, t_i), i=1,2,...,N, is computed, where m_i is the seismic moment, (X_i, Y_i) is the relative location of the i^{th} subevent on the fault plane and t_i is the onset time. The revealed source time function reflects to the complexity of the rupture propagation. Details of the method are given by Kikuchi and Sudo (1985), Stavrakakis et al. (1987). Here, some results of this method obtained for the Chephalonia earthquake of Jan.17, 1983 will be used to compute source parameters of this event. Nowadays, the most popular methods for computing synthetic seismograms are, that proposed by Nabelek (1984) and that which is based on moment tensor solutions (Dziewonski et al., 1981; Dziewonski and Woodhouse, 1983, Giardini et al., 1984).

SEISMIC SPECTRA

The spectral shape, either of far-field or near-field body waves is often utilized to obtain source parameters of earthquakes by using the spectral model proposed by Brune (1970, 1971) or that proposed by Madariaga (1976). The main difference between the two models is that in Madariaga's model the slip increases linearly after rupture, so that the final slip is greater than that expected for a static solution. Consequently, the stress drop for a dynamic solution is larger than the effective stress considered by Brune.

However, the source parameters obtained by using both models are based on 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. Brune's model considers that for a circular fault the source radius is $r=2.34\alpha/2\pi f_0$, where α is the P wave velocity, and f_0 is the corner frequency. On the other hand, Madariaga (1976) proposed for the source dimension the relation $r = 0.32\beta/f_0$, where β is the shear wave velocity. Assuming the ratio $f_0(P)/f_0(S)=1.5$, Brune's model computes a radius of about two times that calculated using Madariaga's model, but gives a stress drop that is about six times lower.

ESTIMATION OF SOURCE PARAMETERS

Once the synthetic seismograms or the spectra of body waves have been computed, several source parameters can be obtained, such as seismic moment, source dimension, stress drop, effective stress and the radiated seismic energy.

In the present study, we use the seismic moments and the source time functions obtained in inversion of body waves to compute the source radius in two different ways. In the first model, we consider a circular fault in which the rupture initiates at a point and then propagates out in every direction on the fault plane with a constant rupture velocity. Assuming that the radius of the circular fault is proportional to the product of the rise time and the shear-wave velocity at the source, then the fault radius a is given (Geller, 1976) by $a=28\pi\beta T_o/(35\pi+16.17\pi\sin\delta+64)$ where T_o is the total duration of the far-field source time function and δ is the angle between the normal to the fault plane and P-ray direction.

Once the source radius has been computed, the stress drop can also be obtained by the relationship $\Delta\sigma=0.44M_o/a^3$, where M_o is the seismic moment. Then, the source area is $S=\pi a^2$, the rupture velocity $U_r=\langle u\rangle\tau$ ($\langle u\rangle$ is the average displacement on the fault plane equal to $M_o/\mu S$) and τ is the rise time. Finally, the effective stress is $\Delta\sigma_{eff}=\mu\langle u\rangle/\beta$.

In the second model, we use the characteristic time t_c of the source-time function (Pacheco and Nebelek, 1988), that is defined as the time at which half of the total moment is released. Assuming an approximately circular rupture that propagates at a velocity $U_r=0.75\beta$, and the time of the rupture growth approximately equal to the time at which 50 per cent of the seismic moment is released, then the source radius is $r=U_r t_c$. Then the stress drop is $0.44M_o/r^3$. The dynamic energy release is computed following Vassiliou and Kanamori (1982). According to them, for a given source time function, the dynamic energy release, E_d, can be estimated through the relationship

$$E_d=2KM_{2o}/\{x(1-x)^2T_o^3\}, \text{ where}$$

$K = \{(1/15\pi\rho\alpha^5) + (1/10\pi\rho\beta^5)\}$, M_o is the seismic moment, T_o is the total duration of the source-time function, x is the ratio of rise time to total duration ($x\approx0.2$), α, β are the velocities of P- and S-waves, respectively and ρ is the density.

Finally, the radiated seismic energy can also be determined from the energy-magnitude relation (Both, 1977), $\log E_M=1.44M_s+12.24$.

SOURCE PARAMETERS FOR SOME LARGE AND MODERATE EARTHQUAKES IN GREECE

As we mentioned above, the source radius of a circular fault can be estimated using two different models. In the following, we compute the source parameters for those earthquakes in Greece for which the source time function is available.

(i) Ionian Sea (Western Greece)

In fig. 1 the fault plane solutions and the source time functions of some large earthquakes occurred in Ionian Sea are shown (after Ioannidou, 1989). It should be mentioned that several fault plane solutions for earthquakes occurring in this region are available (McKenzie, 1978; Drakopoulos and Delibasis, 1982; Papazachos et al., 1986; Anderson and Jackson 1987; Papazachos et al., 1991). Much attention has been paid to the source process of the mainshock of January 17, 1983 (M_s=7.0) (Scordilis et al., 1985; Stavrakakis et al., 1989b; Ioannidou, 1989; Kiratzi and Langston, 1991). It has been concluded that thrusting in the western Hellenic subduction zone terminates in this region against a strike-slip fault striking NE-SW. Using the source time functions illustrated in figure 1, and based on the formulae mentioned in the above sections, the source parameters are estimated (Table 1).

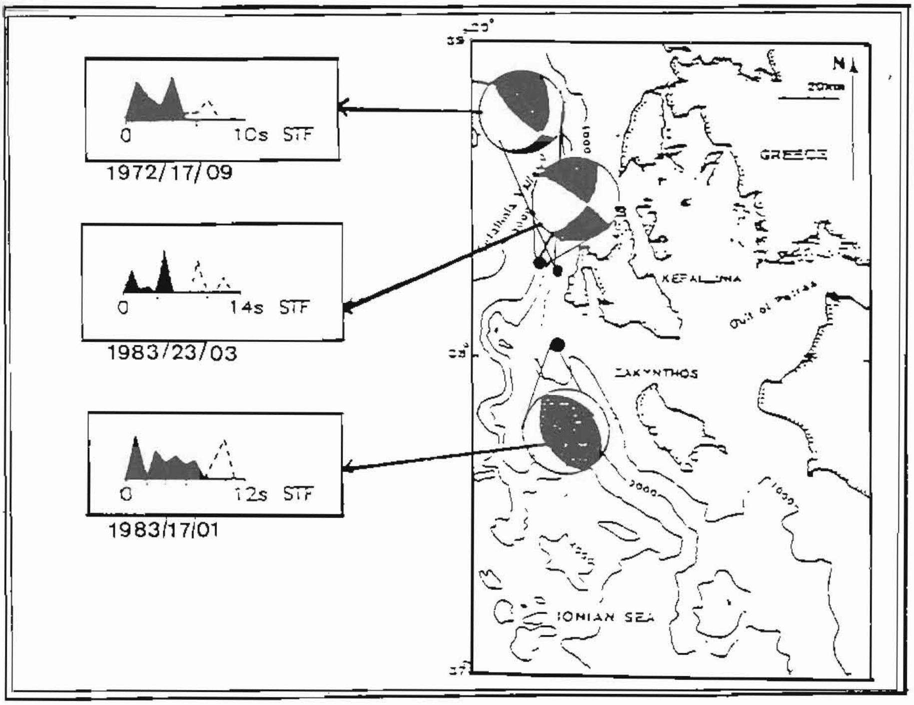

Fig.1. Fault plane solutions and source time functions of some earthquakes occurred in Ionion Sea. The shaded area and the dashed line represent the two point sources used for inversion (after Ioannidou, 1989).

The first values of the source radius R and stress drop $\Delta\sigma$, are based on the characteristic time t_c of the source time function, whereas the second ones (in parenthesis) are based on Geller's model. The obtained stress drop values are low with respect to those expected from the constant stress drop model proposed by Kanamori and Anderson (1975) for interplate earthquakes. It is also interesting to note that the effective tectonic stress is higher than the stress drop indicating that the remaining tectonic stress in this region is also high. This fact might play an important role on the aftershock activity following the mainshock.

Table 1. Source Parameters of Earthquakes Occurred in Ionian Sea.

EQ Y/M/D	M_b	M_s	M_o $x10^{25}$dyn.cm	t_c (sec)	t_o (sec)	R^* (a^+),km	$\Delta\sigma 1$ $(\Delta\sigma 2)$ (bars)	S (km^2)	<u> (cm)	σ_{eff} (bars)	E_d, E_M $x10^{20}$ erg
1972/09/17	5.6	6.3	3.45	4.0	8	10.5 (10.9)	13 (11)	376	30.5	26	0.8, 20.5
1983/01/17	6.1	7.0	20.80	6.0	11	15.7 (15.0)	23 (27)	712	97.4	83	11.9, 208
1983/03/23	5.8	6.2	2.50	5.5	11	14.4 (15.1)	4 (3)	712	11.7	10	0.17, 14.7

* The source radius, R, is based on the source time function
\+ The source radius, a, and the stress drop (values in parenthesis) are based on Geller's model

Far-field displacement spectra are available only for the Cephalonia earthquake of January 17, 1983 (Stavrakakis and Blionas, 1990). These authors obtained a stress drop of 14 bars and 167 bars using Brune's and Madariaga's models, respectively.

(ii) Southern Hellenic Arc

Figure 2 shows the fault plane solutions and the source time functions obtained from inversion of body waves of some large and moderate earthquakes occurring in the southern part of the Hellenic arc (from Taymaz et al., 1990), and Table 2 summarizes their source parameters.

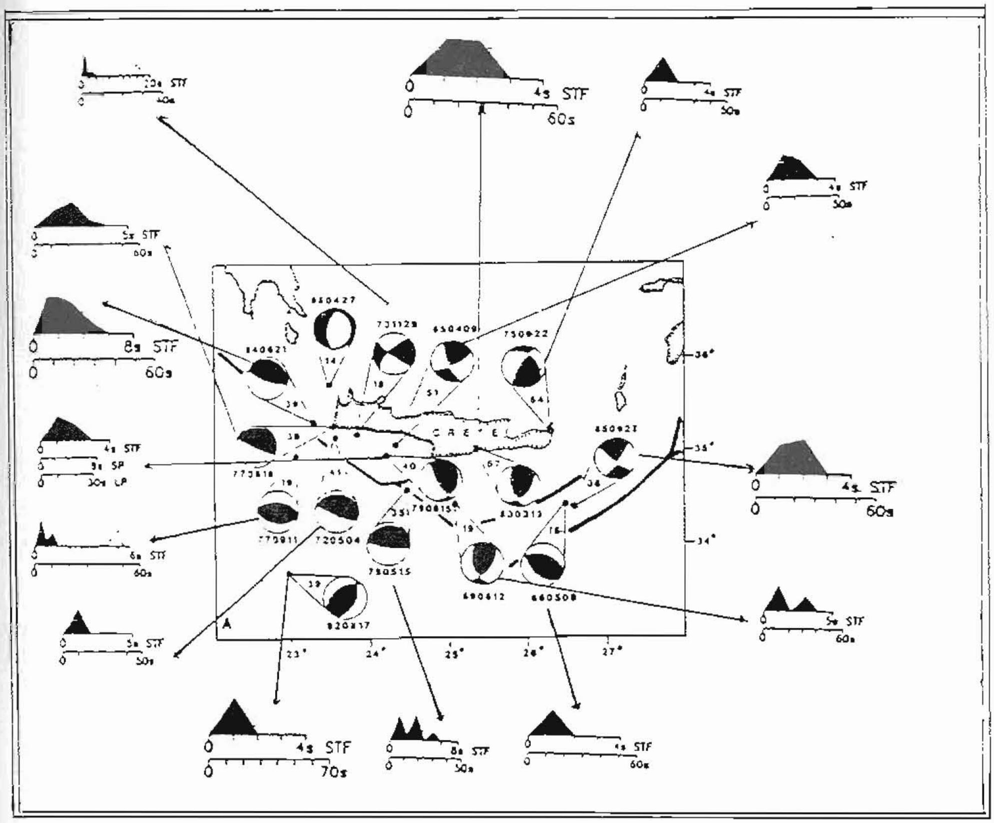

Fig.2. Fault plane solutions and source time functions of some earthquakes in the southern part of Hellenic arc (after Taymaz et al.,1990)

(iii) Gulf of Corinth (Central Greece)

Figure 3 shows the fault plane solutions and the source time functions of some large and moderate earthquakes occurring in the

Gulf of Corinth (after Ioannidou, 1989). Table 3 summarizes their source parameters.

Several authors (Jackson et al., 1982; Papazachos et al., 1984b; Kim et al., 1984; Stavrakakis et al., 1986) have studied the focal process of the earthquake sequence of Feb. 24, 1981. Table 4 summarizes the obtained source parameters.

On the other hand, teleseismic P-wave spectra for this earthquake sequence have also been computed (Stavrakakis et al., 1991) using FFT and iterative maximum entropy techniques, and a stress drop of 58, 17, 16 bars for the main shock of Feb. 24, 1981 and its largest aftershocks of Feb. 25 and March 4, 1981 has been obtained. These values are smaller than those obtained by Stavrakakis et al., (1988) using Geller's model.

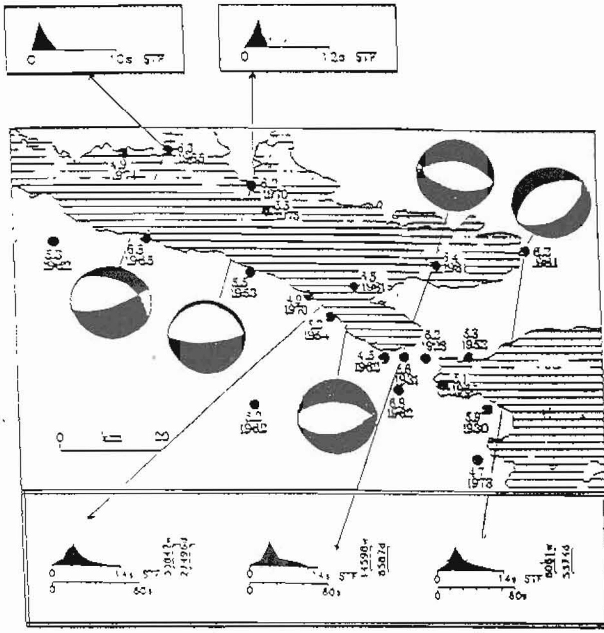

Fig.3. Fault plane solutions and source time functions of some earthquakes in the Gulf of Corinth. The dashed line represents the second point source used for inversion. (after Ioannidou, 1989).

(iv) North Aegean Sea

Figure 4 shows the fault plane solutions and the source time functions of some large and moderate earthquakes occurring in the broad region of North Aegean Sea, as obtaided from inversion of teleseismic body waves (Taymaz et al., 1991). Table 4 summarizes the obtained source parameters. Focal mechanisms and source process of some large earthquakes occurred in Northern Aegean Sea have been studied by Papazachos et al., 1984a, Ionnidou et al., 1986; Kiratzi et al., 1991.

Teleseismic long-period displacement spectra are available only for the earthquake of August 6, 1983 (Drakopoulos and Stavrakakis, 1991). For this event a stress drop of 8.5 bars is

Table 2. Source Parameters of Earthquakes Occurred in Southern Hellenic Arc.

EQ Y/M/D	M_b	M_s	M_q $x10^{25}$dyn.cm	t_c (sec)	t_o (sec)	R^* (a^+),km	$\Delta\sigma_1$ $(\Delta\sigma_2)$ (bars)	S (km^2)	<u> (cm)	σ_{eff} (bars)	E_q, E_M $x10^{20}$ erg
1965/04/09	6.0	6.1	1.5	1.5	3	3.9 (4.1)	108 (95)	53	94.4	80	3.0, 10.
1966/05/09	5.5	5.8	0.2	1.0	2	2.6 (2.7)	48 (42)	23	28.3	24	0.2, 3.9
1969/06/12	5.8	6.1	1.0	1.5	4	3.9 (5.5)	72 (27)	94	35.4	30	0.6, 10.
1975/09/22	5.4	5.5	0.3	1.0	2	2.6 (2.7)	73 (64)	23	42.5	36	0.4, 1.4
1977/08/18	5.5	5.6	0.2	1.8	6	4.7 (8.2)	8 (1.5)	212	3.1	3	0.07, 2.0
1977/09/11	5.8	6.3	1.3	1.9	15	4.9 (20.5)	46 0.6	1324	3.3	3	0.02, 20.
1979/05/15	5.5	5.4	0.4	2.2	6	5.8 (8.6)	9 3	211	6.3	5	0.03, 1.0
1979/06/15	5.5	4.9	0.3	1.2	3	3.1 (4.1)	42 (19)	53	18.8	16	0.12, 0.2
1982/08/17	6.0	6.0	3.0	1.0	2	2.6 2.7	729 (643)	24	424	361	41., 7.5
1983/03/19	5.7	5.0	0.3	1.5	3	3.9 4.1	22 19	53	19	16	0.12, 0.3
1984/06/21	5.8	5.8	1.2	2.4	6	6.3 8.2	21 10	211	19	16	0.24, 3.9
1985/09/27	5.6	5.2	0.5	1.5	3	3.9 4.1	36 31	52	31	26	0.33, 0.5

(*,+ see Table 1)

Table 3. Source Parameters of Earthquakes Occurred in the Gulf of Corinth.

EQ Y/M/D	M_b	M_s	M_q $\times 10^{25}$dyn.cm	t_c (sec)	t_o (sec)	R^* (a^+),km	$\Delta\sigma_1$ ($\Delta\sigma_2$) (bars)	S (km^2)	<u> (cm)	σ_{eff} (bars)	E_q, E_M $\times 10^{20}$erg
1965/07/06		6.4	1.4	2.0	3.5	5.3 (4.8)	42 (56)	72	64.7	55	1.6, 28.0
1970/04/08		6.2	4.0	3.0	7.0	7.9 (9.8)	36 (20)	288	46.0	39	1.7, 14.0
1972/09/13		6.3	1.4	2.0	5.0	5.3 (6.8)	42 (19)	147	31.0	26	0.6, 20.0
1981/02/24	5.9	6.7	87.5	5.0	13.0	13.8 (17.7	170 (68)	994	293.	249	12.7, 7.7
1981/02/25	5.5	6.4	39.5	4.8	13.0	12.6 (17.7)	87 (31)	994	32.0	112	2.7, 2.80
1981/03/04	5.9	6.4	27.0	4.5	13.0	11.8 (17.7)	72 (21)	994	90	77	1.2, 2.80

(*,+ see Table 1)

Table 4. Source Parameters of Earthquakes Occurred in North Aegean Sea.

EQ Y/M/D	M_b	M_s	M_o $x10^{25}$dyn.cm	t_c (sec)	t_o (sec)	R^* (a^+),km	$\Delta\sigma_1$ $(\Delta\sigma_2)$ (bars)	S (km^2)	<u> (cm)	σ_{eff} (bars)	E_d, E_M $x10^{20}$,erg
1965/03/09	5.7	6.3	1.5	1.8	4	4.7 (5.5)	63 (40)	94	53.1	45	1.3, 39.0
1967/03/04	6.0	6.5	2.4	2.0	4	5.2 (5.5)	73 (64)	94	84.9	72	3.3, 20.0
1968/02/19	6.0	7.0	34.5	4.9	9	12.6 (12.3)	76 (81)	477	241.	205	59., 200.
1969/03/28		6.6	14.9	4.2	9	11.0 (12.3)	49 (35)	476	104.	88	1.1, 5.5
1975/03/27	5.5	6.7	2.0	3.0	8	7.9 (10.9)	18 (7.0)	377	17.7	15	0.3, 0.8
1979/06/14	5.9	5.7	0.7	3.0	7	7.9 (9.6)	6 (4)	288	7.7	7	0.05, 0.3
1981/12/19	6.0	7.2	26.0	3.8	7	9.9 (9.6)	115 (129)	288	300.	255	72., 400.
1981/12/27	5.4	6.5	3.8	3.0	7	7.9 (9.6)	34 (19)	288	43.9	37	0.14, 3.9
1982/01/18	5.8	6.9	7.3	3.3	5	8.7 (6.8)	49 (100)	147	165.	140	15., 150.
1983/08/06	6.0	6.9	19.4	4.5	9	11.8 (12.3)	52 (46)	477	135.	115	18.0,150.
1985/04/30	5.4	5.5	0.3	1.8	5	4.7 (6.8)	13 (4)	147	6.8	6	0.03,1.4

(*,+ see Table 1)

obtained using Brune's model.

DISCUSSION - CONCLUSIONS

In the present study, the source time functions, and seismic moments obtained from inversion of teleseismic body waves of some large and moderate earthquakes in Greece have been used to estimate their source parameters: source dimension, stress drop and effective tectonic stress.

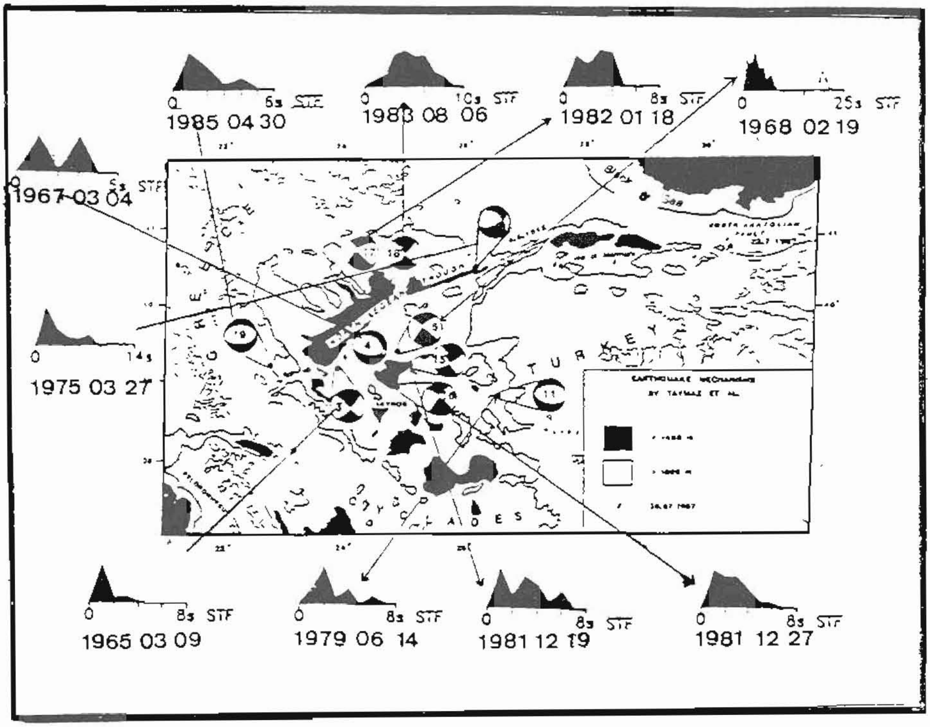

Fig.4. Fault plane solutions and source time functions of some earthquakes in the northern Aegean Sea (after Taymaz et al., 1991)

These parameters are also compared with those obtained by using spectral models, for events for which far-field displacement spectra were available.

However, the difference in the present model is that the source radius r is obtained from the characteristic time t_c of the source time function and not from the relationship $r=2.34/2f_0$ (f_0 is the corner frequency of the far-field displacement spectra). For the stress drop the same relationship, that is $\Delta\sigma=0.44M_0/r^3$, has been used.

In most cases, the stress drop obtained using Brune's model is very low and is in deviation from the constant stress drop model proposed by Kanamori and Anderson (1975) for interplate earthquakes. Low stress drop values have also been obtained for some large earthquakes in Greece (Kulhanek and Meyer, 1979; Soufleris and Stewart, 1981; Kim et al., 1984; Kiratzi et al., 1985; Stavrakakis et al., 1989a; Karakaisis, 1990). Thus, a question that has to be answered is whether the earthquakes in Greece are actually low-stress events or the revealed values are model dependent.

One possible reason for the relatively low stress drop values may be the presence of softer (low strength) materials near the source region. To accept this, however, additional geophysical data are needed. For instance, low electrical resistivity would indicate high temperatures at certain depths, or gravity data might suggest the occurrence of shallow low-density material near the focal region. The low stress drop values can also be explained in terms of the frequency content of earthquakes. For instance, anomalous m_b versus M_s relation has been interpreted as the relative enrichment of short period waves, which may result from short rupture duration or small source dimension (Kim and Nuttli, 1977). For earthquakes in Greece, it has been found (Kiratzi and Papazachos, 1984) that for a given M_s the m_b magnitudes were considerably smaller than those of other regions. Furthermore, considering that the ratio M_s/m_b may indicate the stress level at the source (Archambeau, 1978), events with a smaller m_b, for a given M_s, would indicate a low stress drop.

For the present, we explain the low stress drop values obtained, in terms of the "partial stress model" proposed by Brune et al. (1985). According to these authors, partial stress events might occur when the stress release is not uniform over the fault plane. Thus, if we consider an asperity model whereby a fault is composed of blocks of high and low stress, then the fracture of high stress asperity will produce a high stress drop at that point on the fault. However, the radius of the dislocation may include blocks which have moved with very low stress drop or slips. The parameters derived therefore using the conventional Brune's model are general of the actual fault movement and should be considered as an average estimate of the stress drop. In the following, we give an example which indicates that the partial stress model seems to be valid for some large earthquakes in Greece.

The source radius of the Cephalonia earthquake of Jan. 17, 1983 can be estimated from the source time function shown in Fig.5. (Stavrakakis et al., 1989b). Using the concept of the characteristic time t_c (see previous section) and assuming that a circular rupture propagates at a velocity $V_r=0.75\beta$ (β is the shear wave velocity, 3.5 km/sec), then a source radius equal to 13.2 km is obtained for $t_c=5$ sec. For a seismic moment

M_o=7.33x10^{25} dyn.cm (Kiratzi and Langston, 1991), the average stress drop is 14 bars. Notice that a value of 23 bars is obtained in the present study for the same event using the source time function of Fig.1.

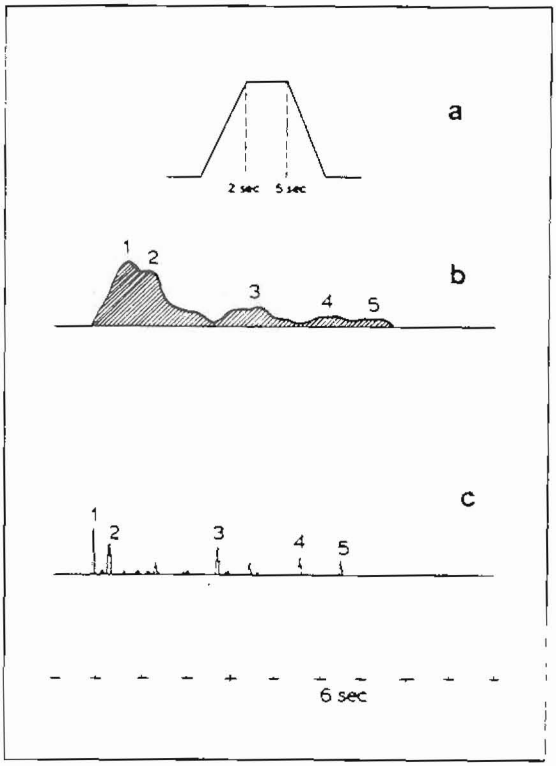

Fig.5. (a) Source time function used as input for inversion, (b) source time function revealed from inversion of the long period P-waves of the Cephalonia earthquake of Jan.17, 1983, and (c) moment rate function.

The difference between the two source time functions is the inversion procedure applied by Stavrakakis et al. (1989b) and by Ioannidou (1989). However, the source time functions of this event indicate that the total seismic moment was released in three main, distinct phases, corresponding probably to three asperities.

For the first asperity, during which approximately 20 per cent of the total moment is released, the rise time is 1.0 sec implying a radius of 2.6 km, and a stress drop of 355 bars. For the second asperity, a stress drop of 263 bars is obtained by assuming that about 50 per cent of the total moment is released.

For the last one a value of 67 bars is obtained corresponding to 30 per cent of the total moment and to a rise time of 2 sec.

The revealed values indicate that the stress release is not uniform over the fault plane but performed in a series of multiple events, with parts of the fault remaining unbroken. Each subevent on the fault plane may be associated with large displacement and large stress drop, but between the areas of individual events, slip may be small. In other words, small areas of high stress drop appear to be embedded in larger, low stress drop areas. This has also been proposed by Mori and Shimazaki (1984) (among others) by inverting strong motion records.

Thus, the value of 23 or 27 bars obtained in the present study may be regarded as an average estimate. It should be mentioned that our analysis is limited to the WWSSN-data used to derive the source time function. Of cource, if additional broad-band data had been used, then a more detailed source time function, therefore more details of the rupture propagation would have been obtained.

One, also important parameter estimated in the present study is the effective tectonic stress. When this parameter is of the same order with the stress drop, then it indicates that the earthquake rupture represents an almost complete release of the tectonic stress accumulated in the focal region. In such cases, no large aftershock is expected. On the contrary, if the stress drop is small in comparison to the effective stress, then a large aftershock might occur. For instance, the stress drops of the main shock of the Cephalonia island earthquake of Jan. 17, 1983 and the Gulf of Corinth of Feb. 24, 1981 were estimated to be 23 bars and 170 bars, whereas the effective stress equal to 83 bars and 249 bars, respectively. In both cases, large aftershocks (M=6.2 and M=6.4, respectively) followed the main shock.

The agreement of the stress drop and effective tectonic stress of the earthquakes in the southern Hellenic arc suggests that these events represent a complete release of the tectonic stress. Actually, these events were not followed by intense aftershock activity. The same conclusion is made for most earthquakes occurred in Aegean Sea.

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