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EFFECTIVE TECTONIC STRESS AND DYNAMIC ENERGY FROM SOURCE-TIME FUNCTIONS OF THE 24 FEBRUARY CORINTH EARTHQUAKE SEQUENCE

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ABSTRACT

Recent developments in earthquake mechanism studies made it possible to relate observed seismograms to various fault parameters such as the fault dimension, dislocation, stress drop and effective tectonic stress. The effective tectonic stress which accelerates the fault motion relates to dynamics of faulting, while the rest of the parameters relates to the statics..

Making use the results of the synthetic seismograms we estimate the source dimension, the stress drop, the apparent stress, the effective tectonic stress as well as the dynamic energy release during the earthquake faulting of the Corinth sequence of the 24 February, 1981.

INTRODUCTION

In recent years many studies have been carried out in order to estimate the dynamic source parameters. It is clarified that the most important focal parameters for the dynamic process of fault motion are the effective tectonic stress, defined as statistic frictional stress minus the sliding frictional stress,and the fracture length, defined as the static frictional stress minus the initial stress (YAMASHITA, 1977).

The former accelerates the fault motion and is directly related to the dynamical process of the fault motion, and the latter decelerates it. BRUNE (1970) has introduced the source time function connecting the effective stress acting on the dislocation surface without considering the role of the friction. WYSS and MOLNAR (1972) based on the lower limit for the initial shear stress, estimated effective stress and frictional stress of Denver, Colorado, earthquakes. HANKS (1977) has estimated the stress drops of small earthquakes and that of major plate margin earthquakes as well as the stresses that drive plate motions. RICHARDSON and SOLOMON (1977) have also estimated the apperent stress drop for intraplate earthquakes and tectonic stress in the plates.

Although the effective tectqnic stress is the simplest gross parameter, its determination has a dual importance in Seismology. First, the ef-

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fective tectonic stress with stress drop determines how much of the effective stress is released, as a stress drop, during an earthquake. The fault plane irregularity (asperities or barries) (KANAMORI and STEWART, 1978; DAS and AKI, 1977; AKI, 1979) may be a dominator factor in preventing the complete stress release. Second, the determination of the effective tectonic stress has an important bearing on the engineering seismology. If it is determined for a certain area, the possible maximum acceleration expected of a future earthquake in the area may be estimated.

From sophisticated wareform modelling using far-field source-time functions it is possible today to estimate the effective tectonic stress in the high seismically active regions.

Using the results of the synthetic seismograms (STAVRAKAKIS et al., 1986) we estimated the above mention parameters for the 24 February, 1981, Corinth earthquake sequence.

Finally, in an attempt to verify the method proposed by VASSILIOU and KANAMORI (1982) for dynamic estimates of the energy release in shallow earthquakes we compute the energies of the main shock of the 24 February 1981 and its principal aftershocks of the February 25 and March 4 using the revealed seismic moments and far-field desplacement source-time functions.

1. SOURCE DIMENSION - STRESS DROP

The source dimensions for the main shock can be estimated if some assumptions are made about the mode of rupture and source geometry. If we assume a horizontal rupture propagation with constant velocity U_r, then the rupture time t_c is given by

$$t_c = L\{1/U_r - (1/\beta)\cos\theta\} \qquad (1)$$

where θ is the angle between the rupture propagation and the ray taking-off at the source and β the shear wave velocity (KANAMORI & STEWART, 1978). In our case,

$$\cos\theta = \sin i_h \cos(\phi_s - \phi_f) \qquad (2)$$

where i_h is the take-off angle, ϕ_s the station azimuth and ϕ_f the fault strike. At teleseismic distance, $i_h \leq 30^\circ$ and $|\cos\theta| \leq 0.5\cos(\phi_s - \phi_f)$, t_c does not change very much with azimuth. Thus, the fault length can be approximated by $L \approx t_c U_r$ and $L \approx 2t_c U_r$ for unilateral and bilateral rupture propagation, respectively.

Figure 1(a) shows the far-field source time function for the main shock of the 24 February 1981 (STAVRAKAKIS et al., 1986), with rise time 3sec, process (rupture) time t_c=6sec and seismic moment 7. $\times 10^{25}$ dyne.cm. Assuming U_r=2.76km/s (0.8 times the S-wave velocity at the source depth) a fault length

$$L = 16.56\text{km (for unilateral propagation)}$$
$$\text{and} \quad L = 33.12\text{km (for bilateral propagation)}$$

is obtained. Based on teleseismic wave abalysis, JACKSON et al., 1984 estimated a fault length of 15km which is in a good agreement with our estimation. KIM et al., 1984 reported that most of the main shock aftershocks are concentrated in a zone within a rectangular-shaped region of about 28 by 17 km.

Another useful model of an earthquake source is 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, β, at the source (GELLER, 1976) the fault radius a is then given by

$$a = \frac{28\pi\beta T_0}{35\pi + 16.17\pi\ \sin\delta + 64} \qquad (3)$$

where T_0 is the total duration of the far-field displacement source time function and δ is the angle between the normal to the fault plane and P-ray direction. In our case (normal faulting) δ is within the range of about 60° to 90°.

Using the value of T_0=12sec (STAVRAKAKIS et al., 1986) a fault radius for the main shock of 8.04km is obtained. The fault diameter is then equal to 16.08km which is consistent with that for the case of unilateral rupture propagation. The average dislocation can be obtained by the relation $\bar{D} = M_0/\mu . A$, where μ is the rigidity and A is the fault area. Assuming $\mu = 3\times10^{11}$ dyne/cm² and $A = \pi r^2 = 203.07\text{km}^2$, we obtain

$$\bar{D} = 118\text{cm}$$

JACKSON et al., 1982 based on teleseismic observations obtained a value of 115cm for the mean displacement, which is in a good agreement with our result. The same authors reported that observed displacement along the 12- to 15 km long fault break on the southeastern end of the Gulf near the Pision reached about 50 to 70 cm, after the main and 25 February shocks. KIM et al., 1984 obtained a value of 37cm for the main shock and 22cm for the 25 February aftershock, based on the inversion of teleseismic body waves.

Their interpretation is that the observed displacements are produced by both shocks. However, it is possible that the surface layers are partially decoupled from the layers at depth, so that the surface displacement represents a fraction of the fault displacement at depth (KANAMORI and STEWART, 1978).

For a circular fault, the stress drop is given by (KEILIS-BOROK,1959; KANAMORI and ANDERSON, 1975)

$$\Delta\sigma = \frac{7\pi}{16}\frac{\mu\bar{D}_0}{r} = 60.51\times10^6\,\mathrm{dyne/cm^2} \approx 60\ \mathrm{bars} \qquad (4)$$

where $\bar{D}_0$ is the average displacement and r the radius.
JACKSON et al., 1982 and KIM et al., 1984 obtained 33.6 and 10.2 bars, respectively. We believe, that their values are relatively low for shallow earthquakes of this size. One reason might be the large value for the fault area (700km^2) deduced from aftershock distribution which is usually overestimated. Moreover, if we regard the Corinth earthquake as an intraplate event, then we expect higher stress drop (KANAMORI and ANDERSON, 1975).

Same calculations are made for the two principal aftershocks of the 25 February and 4 March 1981 using the source time functions shown in Fig.1b and 1c (STAVRAKAKIS et al., 1986). The obtained values are summarized in Table 1.

2. EFFECTIVE TECTONIC STRESS

The effective tectonic stress can be calculated assuming that the fault plane is accelerated by this stress to a finite particle velocity (KANAMORI, 1972). For an infinite fault with infinite rupture velocity, BRUNE (1970) proposed the relation

$$\dot{D}_0 = \frac{\beta}{\mu}\sigma_{eff} \qquad (5)$$

where
$\dot{D}_0$ is the particle velocity, β the shear wave velocity, σ_{eff} the tectonic stress and μ the rigidity. For a finite fault of length 2r with infinity rupture velocity, the particle velocity varies with the time and can be approximated by (BRUNE, 1970).

$$\dot{D}_0 = \frac{\beta}{\mu}\sigma_{eff}e^{-\beta t/r} \qquad (6)$$

By averaging over the process time r/β, the average particle velocity (dislocation velocity) is then given by

$$\bar{\dot{D}} = (1-e^{-1})\beta\sigma_{eff}/\mu \cong 0.63\beta\sigma_{eff}/\mu \qquad (7)$$

For a bilateral fault of length 2r with rupture velocity U_r (BURRIDGE, 1969)

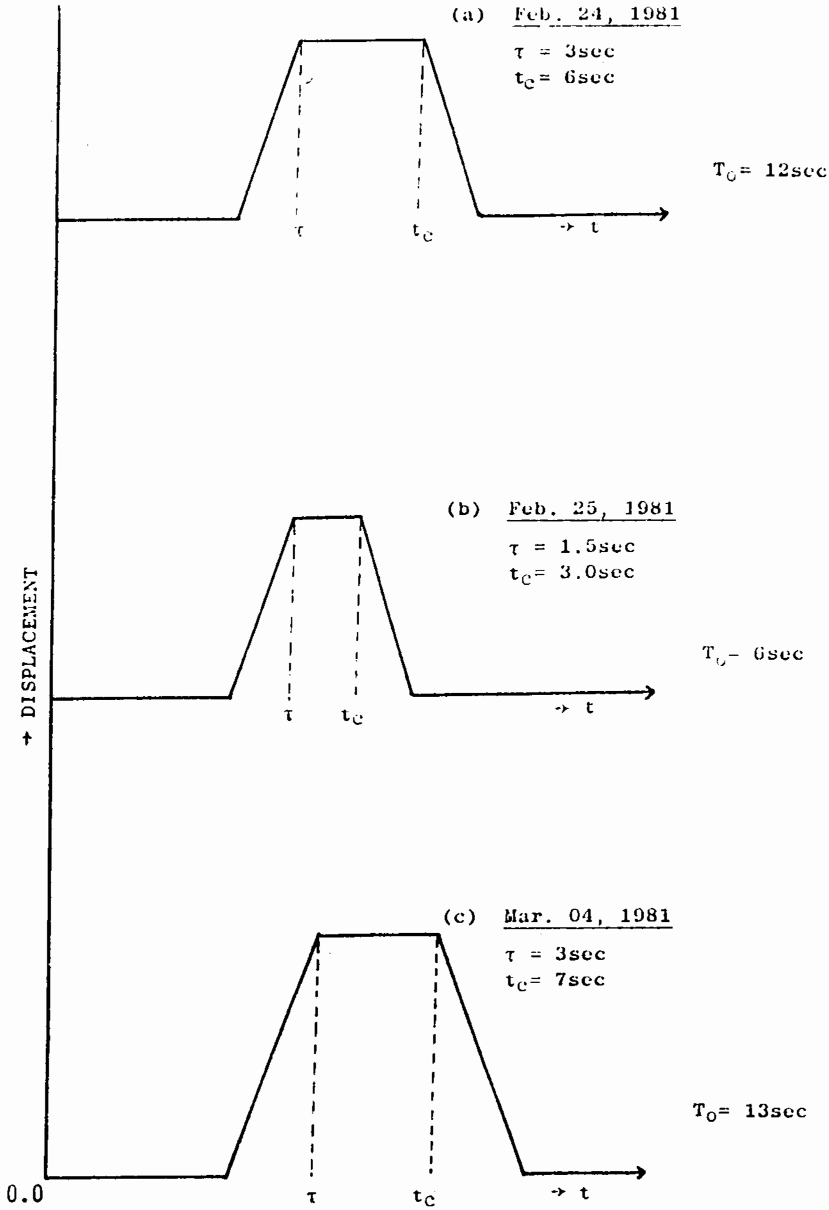

Fig. 1: Source time functions for the main shock of the 24 February, 1981 Corinth earthquake and its principal aftershocks of 25 February and 4 March.

TABLE 1: Source parameters of the Corinth earthquake sequence of 1981

EVENT	L (1) (km)	a (2) (km)	S (3) $(km)^2$	$\bar{D}_o$ (4) (cm)	$\Delta\sigma$ (5) (bar)	$\dot{\bar{D}}_o$ (6) $cm.sec^{-1}$	σ_{eff} (7) (bar)	σ_{app} (8) (bar)
24 FEB.1981	16.56* (33.12)**	8.04	203.07	118	60.5	39.4	33.7 53.6 85.1	29.53
25 FEB.1981	8.28 (16.56)	6.16	119.20	49.7	33.2	33.3	28.5 45.3 71.9	42.33
4 MAR.1981	11.04 (22.08)	7.07	157.03	22.7	13.2	7.57	6.4 10.2 16.3	42.33

(1) $L \cong t_c U_r$

(2) $a = 28\pi\beta T_o/(35\pi + 16.17\pi\sin\delta + 64)$

(3) $S = \pi a^2$

(4) $\bar{D}_o = M_o/\mu A$

(5) $\Delta\sigma = (7\pi/16)(\mu\bar{D}_o/r)$

(6) $\dot{\bar{D}}_o = \bar{D}_o/\tau$

$\sigma_{eff} = \mu\dot{\bar{D}}_o/\beta$

(7) $\sigma_{eff} = (\mu\dot{\bar{D}}_o)/(0.63\beta)$

$\sigma_{eff} = \mu\dot{\bar{D}}_o(1+\beta/U)/\beta$

(8) $\sigma_{app} = n < \sigma > = \mu E_s/M_o$

* unilateral ** bilateral

TABLE 2: Energy release in the Corinth earthquake sequence of 1981

EVENT	$T_o^{(1)}$ (sec)	$M_o^{(1)}$ $x10^{25}$ dyne.cm	$\log E_D$ (2) (erg)	$\log E_G$ (3) (erg)	$\log E_B$ (4) (erg)
24 FEB.1981	12.0	7.02	19.61	21.85	21.88
25 FEB.1981	6.0	1.78	19.91	21.40	21.45
4 MAR.1981	13.0	1.07	18.65	21.40	24.45

(1) from synthetic seismograms (STAVRAKAKIS et al., 1986)

(2) dynamic estimate (VASSILIOU and KANAMORI, 1982)

(3) $\log E_G = 1.5M_s + 11.8$ (GUTENBERG-RICHTER, 1956)

(4) $\log E_B = 1.44M_s + 12.24$ (BÅTH, 1966)

the average dislocation velocity is given by

$$\dot{\bar{D}} = (\beta\sigma_{eff}/\mu) \,/\, (1+\beta/U_r) \qquad (8)$$

Assuming that the fault plane started moving with the above particle velocity and after τ sec was stopped for some reason with final discolation $\bar{D}_o$, then

$$\dot{\bar{D}}_o = \bar{D}_o / \tau \qquad (9)$$

For $\bar{D}_o$ = 118cm and rise time τ = 3sec a value of 39.4cm/sec for the dislocation velocity is inferred. Making use this value, from the above equations we obtain

σ_{eff}= 33.77 bars

σ_{eff}= 53.60 bars

σ_{eff}= 85.16 bars

The mean value for the effective tectonic stress is 57 bars. The agreement of the stress drop (60 bars) with the tectonic stress suggests that this earthquake represents an almost complete release of the effective tectonic stress in the Gulf of Corinth. Some calculations are made for the two principal aftershocks of the 25 February and 4 March 1981. The results are summarized in Table 1.

3. DYNAMIC ENERGY FROM SOURCE - TIME FUNCTION

The energy released in earthquakes can be estimated in a number of ways (BÅTH, 1966). Generally, we may divide energy estimates into two broad classes: the static estimates based on the values of seismic moment and stress drop, and the dynamic estimates based on the direct integration of an observed seismogram or on the integration of an inferred far-field source-time function.

In the present work we compute the energies of the main shock and its principal aftershocks of the Corinth sequence of the 24 February 1981 following the method proposed by VASSILIOU and KANAMORI (1982). They applied the theory of Haskell (1964) to estimate the energies of several shallow events using seismic moments and source-time functions obtained from synthetic seismograms.

Refering to LANGSTON and HELMBERGER (1975) the far-field displacement is given by

$$u(r,\ t) = \{R(\theta,\phi) \,/\, 4\pi\rho U^3 r\}\, M_o T(t) \qquad (10)$$

where $R(\theta,\phi)$ is the radiation pattern of the seismic waves, ρ is the density, U the velocity, and r the distance to the seismic source. M_o is the scalar seismic moment and $T(t)$ is the far-field source-time function. Combining eq. (10) with the energy given by HASKELL (1964)

$$E = \rho U \int_{-\infty}^{+\infty} \int_{0}^{2\pi} \int_{0}^{\pi} U^2 . dt . r^2 . \sin\theta . d\theta . d\phi \qquad (11)$$

VASSILIOU and KANAMORI (1982) (see and fig. 2) obtained the relationship

$$E = 2KM_o / \{ x(1-x)^2 T_o^3 \} \qquad (12)$$

where $K = \{(1/15\pi\rho\alpha^5) + (1/10\pi\rho\beta^5)\}$, M_o the seismic moment, T_o the total duration of the source-time function, x the ratio of rise time to total duration ($x \approx 0.2$), α,β the V_P and V_S wave velocities and ρ the density. Making use the source time function parameters obtained by inversion of teleseismic P-waves for the Corinth earthquake sequence of 1981 (STAVRAKAKIS et al., 1986) we estimate the dynamic energies released during the main shock and its two largest aftershocks. In table 2, the computed values are listed and compared with the energies based on the Gutenberg's (1956) and Båth's (1966) relationships.

It is clear that the Gutenberg energy (static estimate) overestimate the true energy by factors of between 10 and 100.

4. APPARENT STRESS

Both the apparent stress and stress drop represent measures of some fraction of the shear stress acting during an earthquake and can be estimated from seismic observations. The apparent stress is the product of the average shear stress $<\sigma>$ on the fault before and after faulting and the seismic efficiency factor n, and is given by (AKI, 1966)

$$n < \sigma > = \mu \frac{E_S}{M_O} \qquad (13)$$

where μ is the rigidity (3 x 10"dyne/cm²), E_S the radiated seismic energy in dyne-cm and M_O is the seismic moment in dyne-cm. It is clear that without additional independent measurements in the source region of parameters such as total elastic energy, absorbed energy, or in situ stress, the efficiency and the average stress cannot be determined separately. We estimate only their product based on the values of seismic energy and seismic moment for the Corinth earthquake sequence of 1981. The results are summarized in Table 1.

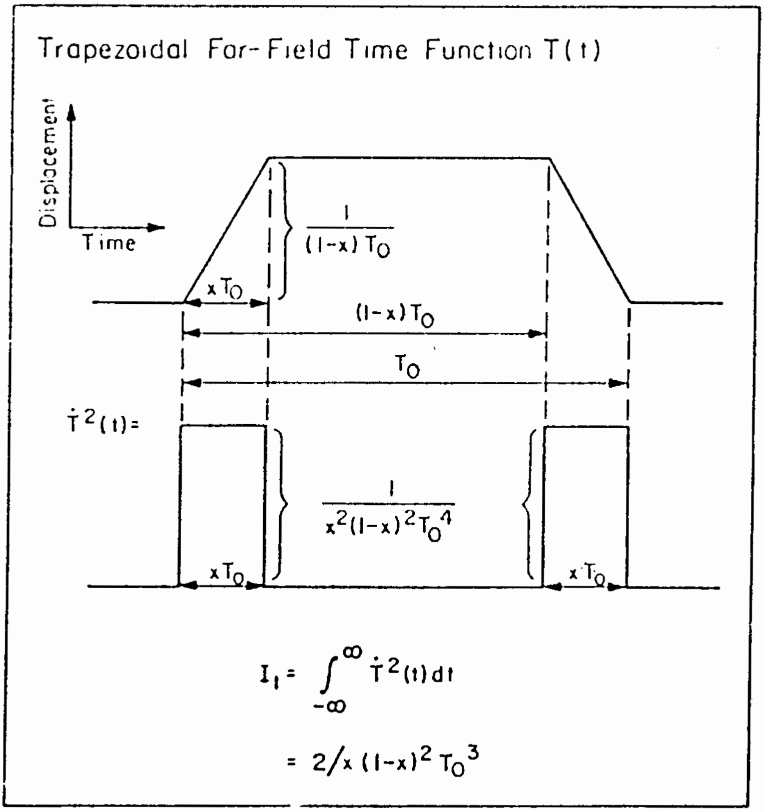

Fig. 2 Source time functions for the main shock of the 24 February, 1981 Corinth earthquake and its principal aftershocks of 25 February and 4 March.

5. CONCLUSIONS

The purpose of this paper was to estimate some important source parameters such as source dimension, stress drop, apparent stress and effective tectonic stress for the strong Corinth earthquake of the 24 February, 1981 and its principal aftershocks of the 25 February and 4 March. Using the seismic moment and the parameters of the far field displacement source-time function we estimated the dynamic energy released in the earthquake sequence.

The approximate agreement of the stress drop with the effective tectonic stress suggests that the Corinth earthquake sequence of 1981 represents an almost complete release of the effective tectonic stress in the region.

It has been also shown that static estimates of energy overestimate the true energy by factors of between 10 and 100. As pointed out by VASSILIOU and KANAMORI (1982) dynamic estimates contain more high-frequency information than the static ones. However, they are still made at teleseismic distances using long-period WWSSN instruments unable to resolve displacement components of frequency greater than 1 to 2 Hz. It is a critical point for further research work.

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