SOURCE PARAMETERS OF THE ORIZABA EARTHQUAKE OF AUGUST 28, 1973

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RESUMEN

Se resume la teoría y la metodología con las cuales se pueden estimar parámetros focales de un temblor a partir de señales sísmicas. Se promedian 11 espectros de onda P (usando registros de corto y largo períodos por cada estación) y se combina con 11 espectros de onda G2 para definir el espectro promedio de la fuente entre períodos de 200 seg. y 0.5 seg. del temblor de Orizaba. El momento sísmico de este temblor de ($m_b \simeq 6.7$) se estima como $4.8 \cdot 10^{26}$ dina cm. La frecuencia de la esquina es 0.1 cps, a partir del cual se estima el radio de la fuente como 13 km con una caída de esfuerzo de 95 bar.

ABSTRACT

The theory and method by which the source parameters of an earthquake can be derived from seismic signals in reviewed. 11 P-wave spectra (short and long period records for each station) are averaged and combined with 11 G2 wave spectra to define the average source spectrum of the Orizaba earthquake between periods of 200 sec and 0.5 sec. The seismic moment of this ($m_b \simeq 6.7$) earthquake is determined as $4.8 \cdot 10^{26}$ dyne cm. The corner frequency is 0.1 cps, from which the source radius is estimated to have been 13 km, with a stress drop of 95 bars.

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INTRODUCTION

A detailed knowledge of the earthquake focal processes is very desirable since it helps (a) in understanding the present tectonics, (b) in seismic risk estimation, and (c) in earthquake prediction. An earthquake is commonly described by origin time, epicentral location, depth to the focus and magnitude. Determination of other source parameters requires special studies. Such studies have become possible in the last decade or so owing to developments in experimental and theoretical seismology. In this study we first summarize briefly the theory which relates the body-wave displacement spectra to some source parameters. We then discuss the corrections which have to be applied to the observed spectra since the theory is for an idealized earth. Finally we report some source parameters of Orizaba earthquake of August 29, 1973 ($m_b \simeq 6.7$; $h \simeq 80$ km; epicenter $\simeq 18.3°$ N, $96.6°$ W). The seismic moment of the earthquake is estimated from both the body wave and the surface wave spectra.

BRIEF REVIEW OF THE THEORY

An earthquake occurs when two sides of a fault suddenly slip with respect to one another converting part of the stored elastic strain energy into seismic energy. The accumulation of the strain energy can be explained by relative motion of plates which constitute the upper part of the earth (the lithosphere). The slip is, in general, a complex function of time and space due to variation of frictional property, roughness, temperature, and tectonic stress on the fault plane.

In this review of the theory relating body-wave spectrum to source parameters, we will consider an infinite, homogeneous, isotropic, elastic earth. Earthquake models in a half space are mathematically very cumbersome (Langston and Helmberger, 1975; Levy and Mal, 1976).

Following Aki (1967) we define seismic moment M_o by

$$M_o = \mu A \bar{u} \tag{1}$$

where μ = rigidity, A = rupture area, and $\bar{u}$ = average slip over the rupture area. At long wavelengths and large distances as compared to the source dimension and at periods greater than the duration of the faulting, an earthquake source may be viewed as a shear dislocation at a point acting as a step function of time and can be described equivalently by a double couple point source without net moment (Burridge and Knopoff, 1964).

Since a point dislocation is equivalent to a double couple point source, the theory of point forces developed by Love (1927, p. 304) can be used. Under the conditions mentioned above the spectra of displacement components are given by (Keilis-Borok *et al.*, 1960; White, 1965, p.220):

$$\Omega_p(\omega) = \frac{1}{4\pi r} \frac{M_o}{\rho\alpha^3} R_p(\theta,\phi) \tag{2a}$$

$$\Omega_s(\omega) = \frac{1}{4\pi r} \frac{M_o}{\rho\beta^3} R_s(\theta,\phi) \tag{2b}$$

when referred to spherical polar coordinates having the conventional relationship to a cartesian frame with the fault plane at z = o and slip at the origin in the y-direction (Figure 1). In equation (2), M_o = seismic moment (equation 1); ρ = density of the medium; α, β = P and S wave velocities of the medium; r = distance from the point dislocation to the observation point P; $R_{p,s}(\theta, \phi)$ = radiation pattern factor (vector) of P and S wave.

$$R_p(\theta,\phi) = (\sin 2\theta \ \sin\phi, o, o) \tag{3a}$$

$$R_s(\theta,\phi) = (o, \cos 2\theta \ \sin\phi, \cos\theta \cos\phi) \tag{3b}$$

From equations (2) and (3) it is clear that P wave has only the radial component with $\sin 2\theta$ $\sin\phi$ as the radiation pattern factor whereas S wave has θ-component (SV motion) and ϕ-component (SH motion) with radiation pattern factors of $\cos 2\theta$ $\sin\phi$ and $\cos\theta$ $\cos\phi$, respectively.

Since the point source approximation in equation (2) is valid for wavelengths larger than the source dimensions, we note that the low frequency spectral amplitudes are independent of frequency and details of the faulting process. The seismic moment can be estimated directly from the low frequency amplitude level using equation (2). Determinations of M_o from P and S wave data, long-period surface wave data, free oscillation data and geodetic data give all reasonably consistent values (Hanks and Wyss, 1972; Kanamori and Anderson, 1975). M_o is one of the more reliably determined parameters since it is obtained from long-period waves which are less affected by structural complexities and anelastic attenuation than the short-period waves.

Static slip and stress drop $\Delta\sigma$ on the fault are related to each other. Stress drop is defined as the difference between the initial tectonic shear stress σ_1 and the final tectonic shear stress σ_2, i.e., $\Delta\sigma = \sigma_1 - \sigma_2$. If in a homogeneous shear stress field of σ_1 a circular shear crack of radius r_o is introduced over which the stress is σ_2, then it can be shown that (Eshelby, 1957; Keilis-Borok, 1959; Singh, 1977):

$$\Delta\sigma = \frac{7\pi}{16}\,\mu\,\frac{\bar{u}}{r_o} \tag{4}$$

and from equation (1)

$$\Delta\sigma = \frac{7}{16}\,\frac{M_o}{r_o^3} \tag{5}$$

Knowing M_o and r_o, $\Delta\sigma$ can be estimated from equation (5). Note that the estimate of $\Delta\sigma$ is dependent on the accuracy of the determination of equivalent radius r_o of the fault. We have assumed that the fault is circular and that stress drop on the fault is constant. A circular fault is an adequate approximation for the Orizaba earthquake since the focal depth was about 80 km. A variable stress drop would change the equation (5) slightly (Singh, 1977). Relationship between $\Delta\sigma$ and M_o for two dimensional faulting has been given by Starr (1928) and Knopoff (1958) and for a rectangular fault by Sato (1972).

Source dimensions can be obtained from field observation (for surface faulting), aftershock area, seismic directivity (Ben-Menahem, 1961) and body-wave spectra. A simple theory relating S-wave spectra to the source size was proposed by Brune (1970).

Brune considered a circular area in a shear field over which traction is suddenly lost. Lacking a dynamic solution, Brune constructed his theory on physical arguments and related the source dimension (radius r_o of the circular fault) to the 'corner frequency' of the S-wave spectrum using energy conservation principle. With his assumptions he obtained

$$r_o = \frac{2.34\beta}{2\pi f_{o,s}} \tag{6}$$

where $f_{o,s}$ is the corner frequency of S-wave spectrum and is defined as that frequency where the low frequency trend of the spectrum intersects the high frequency trend. At high frequencies the spectrum falls off as ω^{-2} for a complete stress drop on the fault. There are some advantages in using P-wave spectrum. Hanks and Wyss (1972) and Wyss and Hanks (1972) assumed, based on physical reasoning, that the relation between r_o and $f_{o,p}$, the P-wave spectrum corner frequency, is the same as given in equation (6) provided that β, the S-wave velocity, is replaced by α, the P-wave velocity:

$$r_o = \frac{2.34\alpha}{2\pi f_{o,p}} \tag{7}$$

Trifunac (1972 a, b) has derived equation (7) based on a slightly different assumption. From equations (6) and (7), it follows that

$$\frac{f_{o,p}}{f_{o,s}} = \frac{\alpha}{\beta} \tag{8}$$

Most observed spectra give $f_{o,p}/f_{o,s}$ between 1 and 2 (Molnar *et al.*, 1973), a value in reasonable agreement with equation (8). Determinations of r_o from equations (6) and (7) have been found to be consistent with field observations (Hanks and Wyss, 1972; Wyss and Shamey, 1975). Whether $f_{o,p}/f_{o,s} > 1$ is consistent with theory has been investigated by several authors (e.g., Sato and Hirasawa, 1973; Savage, 1974; Dahlen, 1974; Madariaga, 1976).

Since these authors consider sources with finite rupture velocities, $v_r < \beta$, they derive the source dimensions from equations of the form

$$r_o = c_s \frac{v_r}{f_{o,s}} \tag{6a}$$

$$r_o = c_p \frac{v_r}{f_{o,p}} \tag{7a}$$

where c_s and c_p are constants of order one. Since the rupture velocity is seldom known, we see from equations (6a) and (7a) that further uncertainties are introduced in the source dimension and stress-drop estimates if v_r varies between earthquakes. Furthermore, Madariaga (1976) concludes from his numerical solution of a circular dynamic shear crack that equation (6) overestimates r_o by a factor of about 2 (and thus $\Delta\sigma$ is underestimated by a factor of 8; see equation (5)). Calibrations of equations (6) and (7) for three deep earthquakes with aftershocks (Wyss and Shamey, 1975; Wyss and Lu, 1977) indicate that these equations overestimate the source dimensions only by approximately 25%. Therefore, we feel justified in using equations (6) and (7) in the present study, and we anticipate that the stress-drop determined in this way is probably too small by a factor of approximately 2.

The total energy E during rupture is given by (Wyss and Molnar, 1972):

$$E = (\sigma_1 + \sigma_2)\bar{u}\,A/2 \tag{9}$$

Part of this energy, E_s, is radiated as seismic waves and the rest, E_f, is lost as frictional heat; thus

$$E = E_s + E_f \tag{10}$$

$$E_f = \sigma_f \, \overline{u} \, A \tag{11}$$

where σ_f is average frictional stress on the fault. Thus E_s is given by

$$E_s = (\sigma_1 + \sigma_2 - 2\sigma_f)\overline{u} \, A/2 \tag{12}$$

Note that σ_2 may not be equal to σ_f. We can write

$$E_s = \eta E \tag{13}$$

where η = seismic efficiency is a quantity which is less than 1. Remembering the definition of seismic moment M_o in equation (1) it follows that

$$E_s / M_o = \frac{\eta}{\mu} < \sigma > \tag{14}$$

where $\eta < \sigma > = \eta(\sigma_1 + \sigma_2)/2$ is defined as the apparent average stress. Seismic energy can be obtained from spectral analysis of seismograms (Wyss, 1970) or from the Gutenberg energy-magnitude formula (Gutenberg and Richter, 1956)

$$\log E_s = 5.8 + 2.4 \ m_b \tag{15}$$

Thus an estimate of apparent average stress $\eta < \sigma >$ at the focus can be obtained. However it is not possible to determine the seismic efficiency η and the average stress $< \sigma >$ without knowing the *in situ* stress or energy lost in heat.

If we assume that $\sigma_2 = \sigma_f$ (this is called the Orowan (1960) model) then from equations (9) to (13) we get

$$\eta_f < \sigma > = \Delta\sigma/2 \tag{16}$$

where η_f is the seismic efficiency assuming $\sigma_2 = \sigma_f$.

In this paper the basic parameters determined are the seismic moment M_o and corner frequency of the P-wave spectra $f_{o,p}$. Equation (15) is used to determine the seismic energy E_s. Using these parameters, other source parameters such as source dimension r_o, stress drop $\Delta\sigma$, apparent average stress $\eta <\sigma>$ and $\eta_{\bar{f}} <\sigma>$ have been estimated from the relations given above. Seismic moment M_o was determined from P wave spectral level at long-wave length as well as from spectral density of 50 to 200 second surface waves using the tables given by Ben-Menahem *et al.* (1970).

CORRECTIONS AND CHOICE OF DATA

The theory presented above is for an idealized earth. Seismic waves propagating through the real earth encounter various discontinuities, suffer attenuation due to internal friction and are modified as they pass thorough the crust and are recorded at a free surface. In order to use the theory the data needs to be carefully chosen and properly corrected. We must select data at such epicentral distance Δ that the P or S wave pulse has a wide window. Such phases as pP, PcP, sS, ScS limit the window length. It is desirable to take the longest possible window since the spectral amplitudes at periods greater than the window length cannot be determined. Since for $\Delta < 35°$ upper mantle discontinuities cause large amplitude variations it is safer to take data at $\Delta > 35°$. However for large Δ, PcP and ScS merge with P and S. Thus the best Δ for analysis is between 35° to 70° (Wyss, 1973).

Corner frequency of most earthquakes with $m_b \leqslant 6$ is near the high resolution limit of long period instruments of the World Wide Standard Seismograph Network (WWSSN). Therefore the spectra from both long period as well as short period records are patched together for the analysis.

To obtain seismic moment M_o from equation (2) several corrections need to be applied to the spectra $\Omega_{p,s}(\omega)$. The simplest are the geometrical spreading r in layered spherical earth which can be calculated in a straightforward manner (Julian and Anderson, 1968) and the radiation pattern correction $R_{p,s}(\theta, \phi)$ which depends on the fault plane orientation and the location of the station with respect to the fault plane. The fault plane can be obtained from first motion studies and radiation pattern correction at any station can then be easily calculated (Ben-Menahem *et al.*, 1965).

It would be very useful to correct each spectrum for the crustal structure below each station but this is only possible if the structure is well known. Also this correction is not indispensable since (a) we need not explain each trough and peak of the spectrum but only its overall shape and (b) we will take the average of several stations and it will be sufficient to correct for average crustal amplification. A factor of 2.5 is considered adequate for crustal structure and free surface amplification and amplitudes at all frequencies are divided by this factor.

Along a ray path amplitude decreases due to internal friction and this attenuation is not well known. The decrease in amplitude can be roughly approximated by

$$A(f, t) = A_o e^{-\pi f t/Q} = A_o e^{-\pi S f/Qv}$$

where A_o = initial amplitude, f = frequency, t = travel time, Q = attenuation coefficient, S = distance and v = wave velocity (α or β). It follows that at high frequencies and large distances the shape of the spectrum would change. However for events with $m_b \geqslant 5.5$, the corner frequency is $\leqslant .1$ Hz. Only at frequencies higher than this will the attenuation effect start to exceed 20% for P and S waves (Julian and Anderson, 1968). Thus the data from WWSSN can be used to determine the corner frequency for events with $m_b \geqslant 5.5$ but not the seismic energy which depends strongly on the high frequency amplitudes.

For a detailed discussion on the choice of data and corrections to the spectra we refer to Wyss (1973). Here we summarize the steps required to estimate source parameters from body-wave spectral analysis:

(a) Determine the focal mechanism of the earthquake.

(b) Choose well recorded data at Δ between 35° to 70° with wide azimuthal coverage.

(c) Isolate P and S wave phases recorded by long and short period (alternatively broad band) instruments, digitize, and obtain spectrum at each station and correct for instrumental response.

(d) From the focal mechanism obtain radiation correction factor.

(e) Apply crustal and free surface amplification correction to the spectra. Correct for geometrical spreading, radiation pattern correction and attenuation correction.

(f) Obtain seismic moment M_o using equation (2) and the long period spectral level. Estimate corner frequency and compute source dimension r_0 from equations (6) and (7).

(g) Compute average of M_o and r_o from their value at each station.

(h) Compute stress drop $\Delta\sigma$ from equation (5), average dislocation $\bar{u}$ from equation (1), apparent average stress from equations (14) to (16).

THE ORIZABA EARTHQUAKE

On 28 August 1973 an earthquake occurred at a depth of about 80 km near the city of Orizaba in the state of Veracruz, Mexico. National Earthquake Information Service (NEIS) gives its location as 18.3°N, 98.6°W and its body wave magnitude m_b as 6.8. The magnitude m_b as reported by International Seismological Centre (ISC) is 6.6. The earthquake was located in the Mexican volcanic belt.

Focal mechanisms of a few other earthquakes in this region reported by Molnar and Sykes (1969) show normal faulting. Strong shaking during 28 August 1973 earthquake was felt in an area of about 350 000 km^2 and resulted in 500 deaths and extensive structural damage (De Valle, 1973; Meehan, 1973). No aftershocks were recorded (C. Lomnitz, personal communication); therefore source dimension and stress drop have to be estimated from seismic signals. The near source surface reflection, pP, arrives about 20 sec after the direct P wave. This limits the analysis window to 20 sec, which means that the combined body wave spectra of the long and short period WWSSN instruments will furnish data between 20 sec and 0.5 sec period only (frequency 0.05 to 2 cps). Since the corner frequency is expected to lie near 0.1 cps, the P–wave data will define the high frequency part of the spectrum well; however, the low frequency level will only be defined approximately. The long period spectral level is most easily derived from surface waves. Therefore our approach will be to abtain the spectra of surface waves and body waves. We will correct both kinds of spectra back to the source and express them in moment M_o as a function of frequency. Then we will combine the information to get an average source spetrum which will have the units of moment and which will be defined from 0.005 to 2 cps (periods 200 sec to 0.5 sec). From this combined spectrum we will be able to determine the moment and corner frequency with confidence, because of the wide range over which the spectrum is defined.

All available P-waves that fit the selection criteria outlined above were digitized, and their spectra are shown in Figure 2. The lower curves are the uncorrected station spectra. The anelastic attenuation correction resulted in the upper curves shown. The low frequency amplitudes remain unchanged, whereas the high frequency amplitudes are increased by this correction. As expected the P-spectra define very well the high frequency fall off, but the corner frequency cannot be estimated with confidence. Our estimate of the approximate long period amplitude level, Ω_p, is given for all stations in Table 1. From Ω_p we calculate approximately the moment (Table 1) using equation (2), after applying all the corrections mentioned previously.

The moment estimated from surface waves, also given in Table 1, is considered more reliable. In the average (bottom Table 1) all moment determinations agree very well. The surface waves digitized are shown in Figure 3. The spectral amplitude of the G2 signals is shown in Table 1, for the four periods 50, 100, 150 and 200 sec. The spectra amplitudes given were determined from one horizontal component only. After appropriate correction considering the back azimuth at each station M_o was calculated using the tables by Ben Menahem *et al.* (1970). Even though the moments of individual stations vary a great deal the average surface wave moments for different periods agree well with each other and with the P-wave moment.

The combined source spectrum for this earthquake is shown in Figure 4. This spectrum contains the RMS averages of all the P- and surface wave spectra from Figure 2 and Table 1 respectively. In other words the P-station spectra were corrected to obtain source spectra as outlined in the beginning of this paper. Then the spectra of all azimuths were averaged. The result with attenuation correction is shown as solid line in Figure 4, without attenuation it is shown by triangles. The average surface wave moments at the four selected periods are shown as solid dots. From this combined source spectrum where the frequency range is more than two orders of magnitude we can estimate M_o, and $f_{o, p}$ with confidence. Some subjectivity will still enter in the choice of corner frequency. If the spectrum is interpreted by the dashed asymptotes shown in Figure 4, the moment is estimated at $4.8 \cdot 10^{26}$ dyne cm, and the corner frequency is 0.11 cps. Alternatively one might chose $f_{o, p} = 0.09$, a small difference indeed. The source dimension r_o and stress drop $\Delta\sigma$ calculated by equations (7) and (5) are 17 km and 43 bars, respectively. As mentioned earlier it seems that equation (7) overestimates r_o by 25%. Thus our best estimate for source radius is 13 km which gives, from equation (5), a stress drop of 96 bars. In the following we list the source parameters of the Orizaba earthquake. A body wave magnitude of 6.7 has been taken in the calculations.

Seismic moment, $M_o = 4.8 \cdot 10^{26}$ dyne cm
P-wave corner frequency, $f_{o, p} = 0.1$ cps
Radius of the fault, $r_o = 13$ km
Stress drop, $\Delta\sigma = 96$ bars
Apparent average stress, $\eta \langle\sigma\rangle = 10.5$ bars
Apparent average stress for
the Orowan model, $\eta_f \langle\sigma\rangle = 48$ bars
Average dislocation, $\bar{u} = 133$ cm

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TABLE 1. MOMENTS OF THE ORIZABA EARTHQUAKE DETERMINED FROM P-AND SURFACE-WAVES

Station	P-waves		Surface waves (G2)								Distance, in degrees	Azimuth, in degrees
	Ω_p, in cm sec	$M_0 \times 10^{26}$, dyne cm	Spectral amplitudes				$M_0 \times 10^{26}$, in dyne cm					
			T = 50	T = 100	T = 150	T = 200	T = 50	T = 100	T = 150	T = 200		
ADE	---	---	.030	.040	.15	.30	2.8	0.76	1.6	2.1	128	240
AFI	.012	1.2	.011	.120	.21	.10	2.4	3.9	3.4	1.0	81	252
AKU	.011	6.2	---	---	---	---	---	---	---	---	69	26
ANT	---	---	.001	.074	.10	.15	5.3	4.2	2.9	2.5	49	148
ARE	.013	3.0	.027	.15	.18	.30	9.9	6.1	3.5	3.2	42	143
BUL	---	---	.026	.20	.30	.35	2.2	3.5	3.1	2.4	128	102
COL	.022	8.8	---	---	---	---	---	---	---	---	58	337
COP	---	---	.007	.056	.11	.14	3.5	4.3	4.0	3.0	85	32
COR	.032	5.8	---	---	---	---	---	---	---	---	35	325
ESK	.012	9.4	---	---	---	---	---	---	---	---	77	36
GDH	.020	7.8	---	---	---	---	---	---	---	---	57	17
KIP	.020	2.2	---	---	---	---	---	---	---	---	58	284
NAT	.012	node	---	---	---	---	---	---	---	---	65	105
NUR	---	---	.016	.089	.13	.17	13.6	14.4	11.2	8.8	88	25
PTO	.010	7.6	.016	.046	.05	.09	4.0	1.6	.8	.9	77	51
RAB	---	---	.021	.090	.10	.09	9.2	7.2	4.8	2.5	111	273
TRI	---	---	.007	.029	.05	.06	4.1	2.6	2.2	1.8	91	41
TRN	.006	2.8	.034	.10	.30	.32	(21.)	7.3	10.0	6.9	35	98
WES	.011	1.3	---	---	---	---	---	---	---	---	32	230
Average	---	5.7	---	---	---	---	5.7	5.1	3.5	3.2	---	---

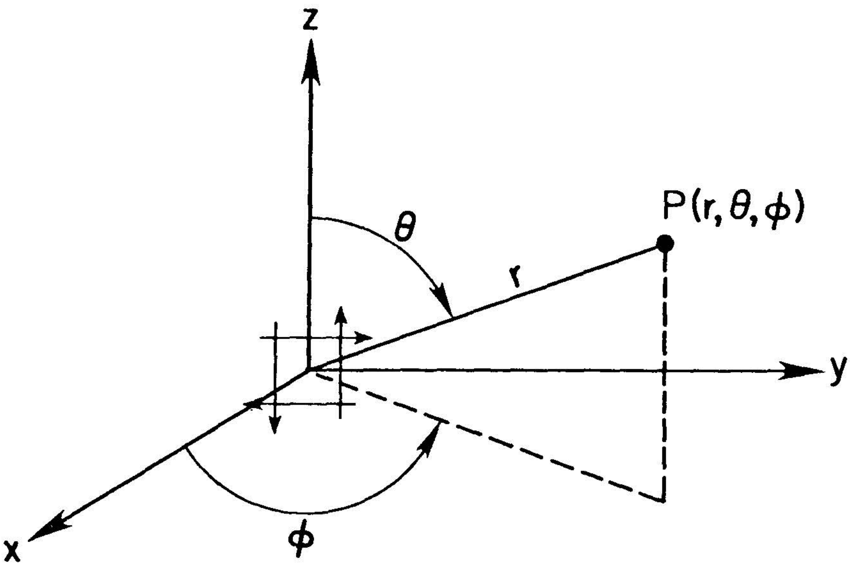

Figure I. A point slip (dislocation) at the origin in z =o plane and y-direction is equivalent to a distribution double couple point forces without moment. The forces would be located in the x =o plane at the origin. In the far-field and at long wave-lengths a point double couple source ia a good approximation to an carthquake.

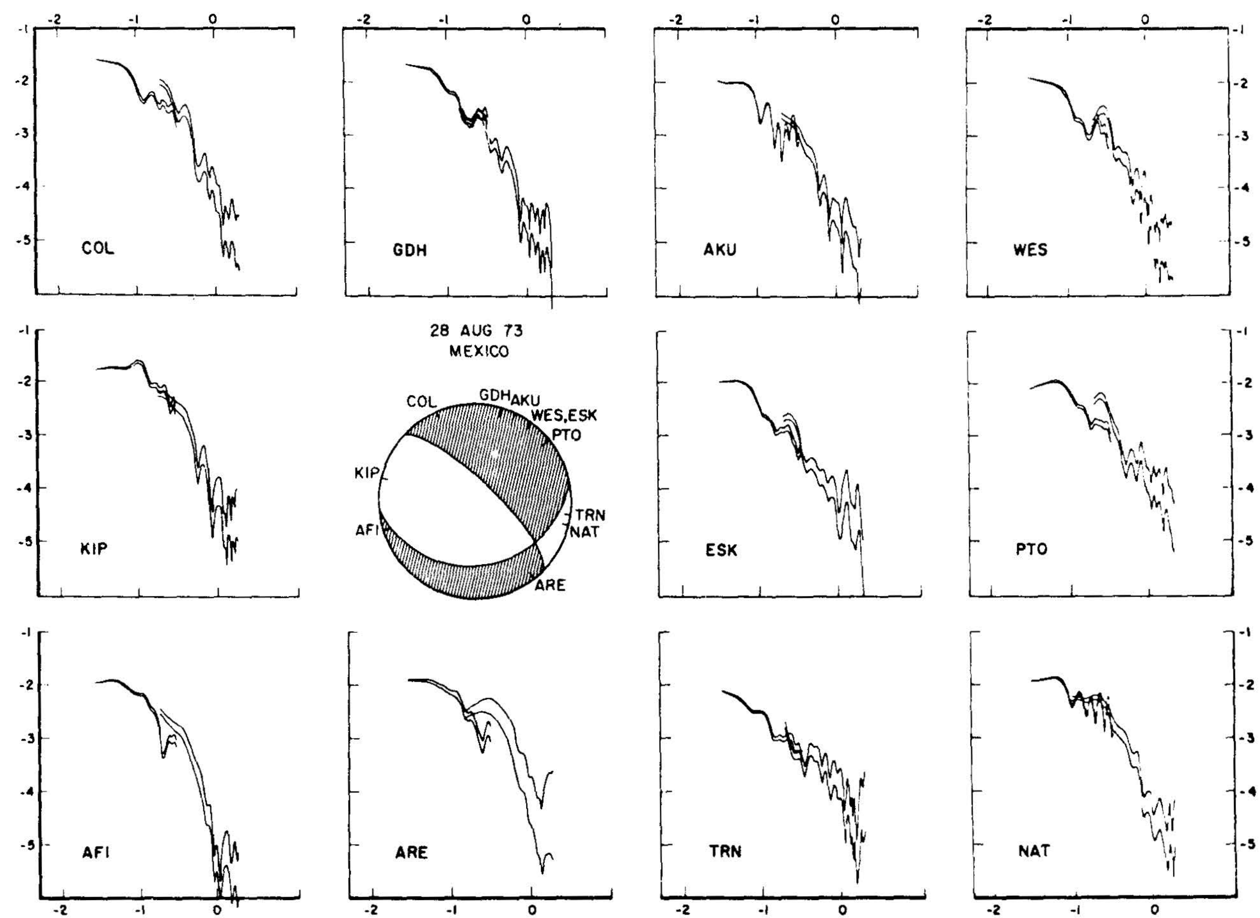

Figure 2. P-wave spectra as a function of azimuth. At the center the focal mechanism determined from first motions is shown with shaded quadrant compressional. Station abreviations are these of the WWSSN network. The azimuth to the stations is indicated on the focal sphere. Vertical axis is the logarithm of spectral amplitude in cm sec, horizontal axis is the logarithm of frequency in cps. The long and short period data are shown jointly for each station with the lower curves being the uncorrected station spectra, the upper curves are corrected for the anelastic attenuation using the results of Julian and Anderson (1968).

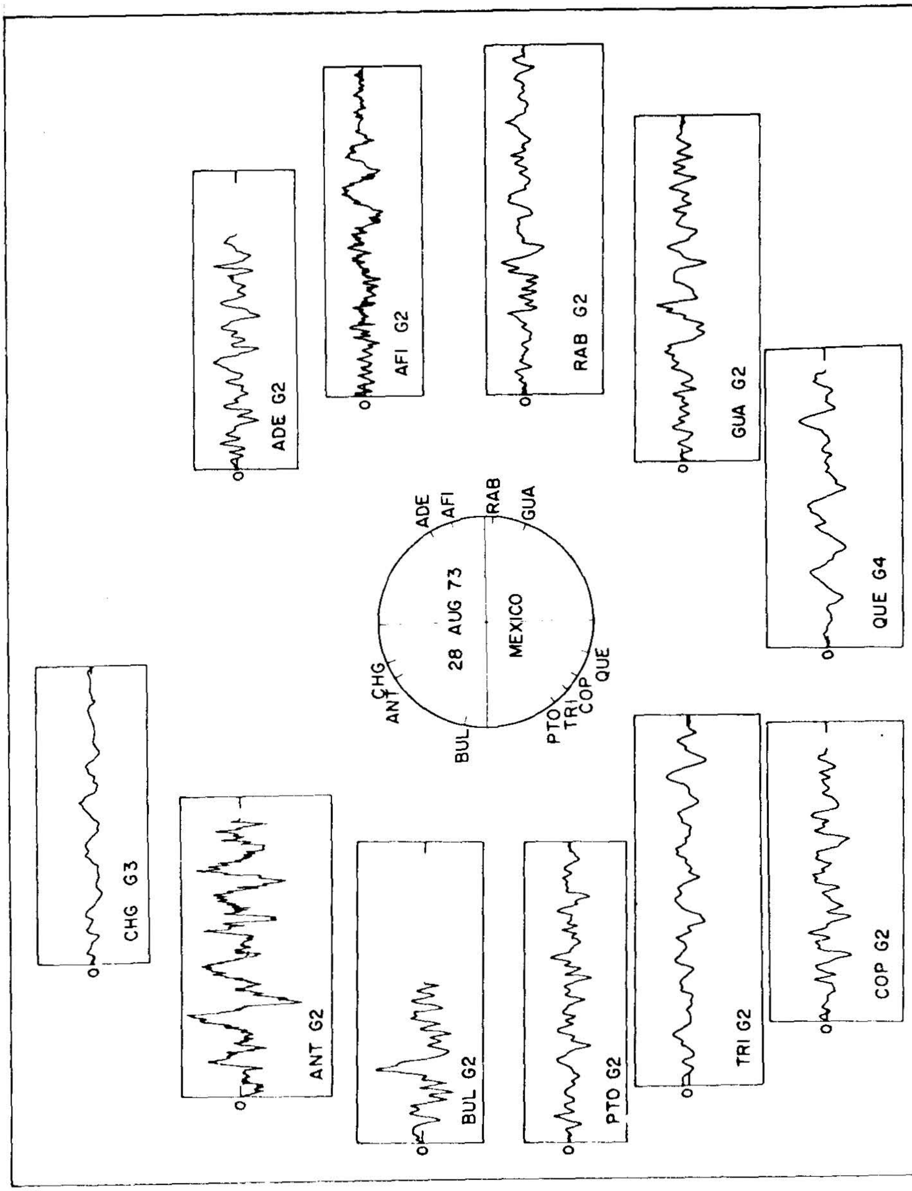

Figure 3. Surface wave traces as a function of azimuth. The length of the boxes is 560 seconds.

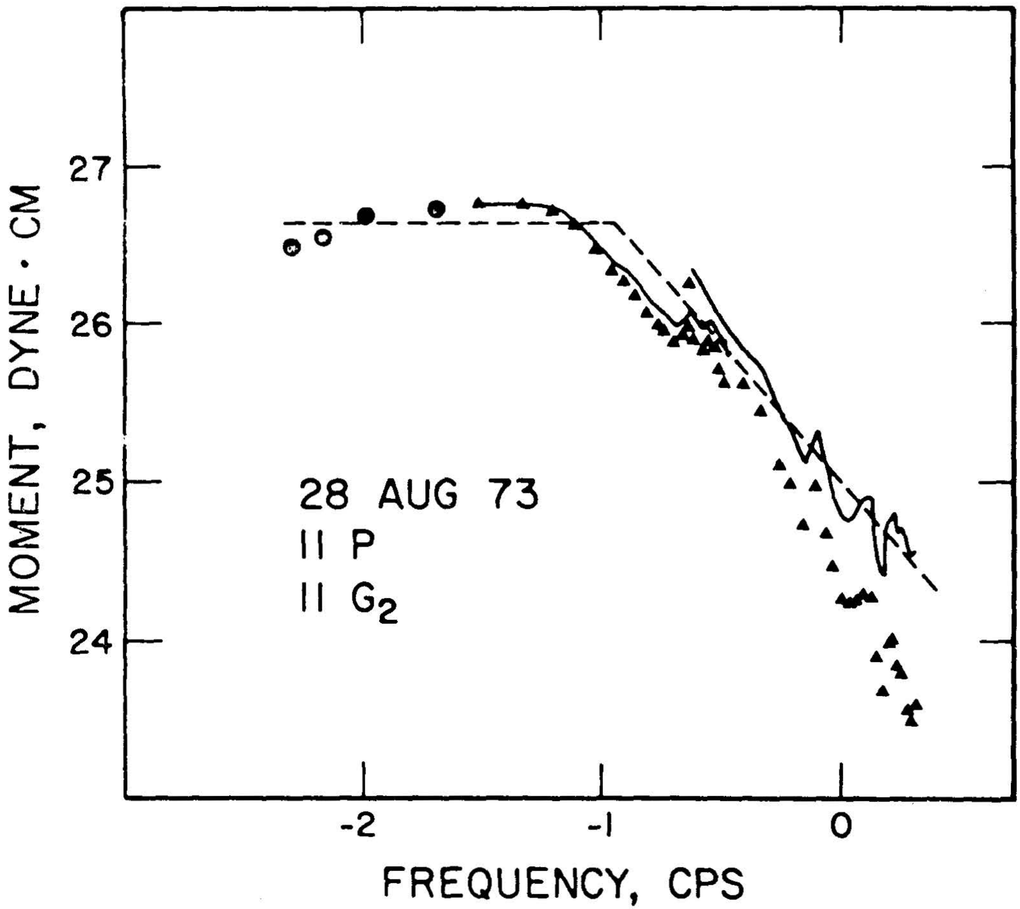

Figure 4. Average source spectrum of the Orizaba earthquake. All P-wave and surface wave data (Table 1) are corrected back to the source, averaged and expressed in units of moment as a function of frequency. 11 P-signals and 11 G2 records were used.

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