CHARACTERIZATION OF RESPONSE SPECTRA BY PARAMETERS GOVERNING THE GROSS NATURE OF EARTHQUAKE SOURCE MECHANISMS

by

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SYNOPSIS

Prediction of response spectra for earthquake engineering purposes is considered from the point of view of the dislocation theory of earthquakes. It is shown that the traditional scaling of response spectra by the predicted peak acceleration should be limited to the high-frequency end of the spectrum, and that the peak acceleration in the near field is not strongly correlated with earthquake magnitude. The amplitude of the long-period end of the response spectrum at source to station distances greater than about 10 source dimensions should be correlated with seismic moment, while for distances less than about one source dimension this amplitude should be proportional to the permanent ground displacement. To reconcile the existing extensive data on seismicity of active regions based on the magnitude scale, it is shown that magnitude correlates approximately with seismic moment.

SIMPLE DISLOCATION MODEL

Although many parameters are required to describe fault motion in time and space, strong-motion studies of the source mechanism are, at the present time, mainly concerned with those parameters that have the most prominent influence on the nature of recorded ground motion. These are: (1) effective stress σ (Figure 1) and (2) the seismic moment M_o. The seismic moment $M_o \equiv \mu\bar{u}A$ combines information on rigidity in the source region μ, average dislocation $\bar{u}$ and the area of faulting A in one parameter. The significance of seismic moment(1) emerges from the fact that it specifies the amplitude of the long-period level ($\omega \to 0$) of the Fourier amplitude spectrum $\Omega_{FF}(\omega)$ of the far-field S-wave displacement through $\Omega_{FF}(\omega) \to M_o(4\pi\rho R\beta^3)^{-1}$ Here ρ is the material density, R is the source to station distance and $\beta = (\mu/\rho)^{1/2}$ is the shear wave velocity. However, near a fault the significance of the overall fault size and therefore its moment M_o are lost and the static displacement u after the earthquake becomes the only dominant factor for the long-period components of the near-field Fourier amplitude spectrum $\Omega_{NF}(\omega)$ which for $\omega \to 0$ tends to $\Omega_{NF}(\omega) \to u/\omega$.

As shown in Figure 1, an earthquake dislocation in an infinite space might be modeled by a plane surface area A whose size is measured by the radius r of an equivalent circular disc(2). For simplicity it can be assumed(1, 2) that the stress pulse σ (Figure 1) produced by a sudden drop of surface tractions on A during an earthquake is applied instantaneously over the whole fault surface. The approximate nature of the resulting S-wave displacement is then as shown in Figure 1 and the resulting Fourier amplitude spectra of ground displacement(1, 2) become

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$$\Omega_{NF}(\omega) = \frac{\sigma\beta}{\mu}\,\omega^{-1}(\omega^2+\tau^{-2})^{-1/2} \quad \text{(for near-field S-wave motion at P, Fig. 1)} \qquad (1)$$

and

$$\Omega_{FF}(\omega) = \frac{r}{R}\,\frac{\sigma\beta}{\mu}\,\frac{1}{\omega^2+\alpha^2} \quad \text{(for far-field S-wave motion at Q, Fig. 1)} \qquad (2)$$

The time constant τ can be shown to be related to the size of an earthquake r and shear wave velocity β through $\tau = \eta r/\beta$ where η is a dimensionless parameter depending on the fault geometry, elastic constants and the point of observation. Values of η range from 0 to η_{max}, which is given in Table I for several simple fault geometries. Parameter α, also called the "corner frequency" is given by $\alpha = 2.34\,\beta/r$[1, 2, 3].

The effective stress $\sigma = \sigma_o - \sigma_f$ (Figure 1) is therefore the principal parameter specifying the spectral displacement amplitudes at high frequencies. As $\omega \to \infty$, $\Omega_{NF}(\omega) \to (\sigma\beta)/(\mu\omega^2)$ for $\omega >> 1/\tau$ and $\Omega_{FF}(\omega) \to (r\sigma\beta)/(R\mu\omega^2)$ for $\omega >> \alpha$. Although these results imply that the acceleration spectra of S-waves would have constant amplitudes for high frequencies, attenuation and scattering reduce the observed amplitudes significantly. The precise nature of attentuation of high-frequency seismic waves is not well understood, partly because of the sparsity of recorded accelerograms. It may be assumed for the purpose of this qualitative discussion, however, that the attenuation of body waves is adequately described by $2\exp[-(\omega R)/(2Q\beta)]$ where the Q factor may range from about 50 to 500[2, 3], and 2 models the amplification due to the half-space boundary. The normalized spectra (1) and (2) corrected for attenuation are plotted versus period $T = 2\pi/\omega$ in Figure 2 for an arbitrary choice of $\alpha = 1/\tau = \pi/5$.

CHARACTERIZATION OF EARTHQUAKE RESPONSE SPECTRA

A method often used in scaling of the design spectrum amplitudes is based on an assumed empirical relationship between peak acceleration and earthquake magnitude. Figure 3 shows data on peak acceleration, plotted versus distance from the fault, and indicates no strong correlation between peak acceleration and earthquake magnitude. In fact, this figure shows that an earthquake with smaller magnitude can have higher peak acceleration at the same distance. This is consistent with the simplified theory in this paper which suggests that peak acceleration should be correlated with the effective stress σ, since the peak acceleration usually samples the high-frequency spectral amplitude for $\omega > \alpha$ or $\omega > 1/\tau$.

One must know μ, $\bar{u}$, and A to scale the long-period end of the spectrum by the moment M_o for intermediate and large distances (R>>r). Although these quantities might be estimated from geological and seismological investigations in the area under consideration, frequently the only statistics available are in terms of earthquake magnitude. There are several magnitude scales currently in use, but in this paper we will employ Richter's local magnitude M_L for shocks less than 6 and the surface wave magnitude M_s for shocks greater than 6. The magnitude gives an estimate of the amplitude of the long-period end of the spectrum if the frequency band of the recording instrument is centered between 0 and α. Therefore, since $\Omega_{FF}(0) = M_o(4\pi\rho R\beta^3)^{-1}$ a relationship should exist between the magnitude of an earthquake and the seismic moment M_o. Figure 4 presents such a relationship based on data from central and southern California[2, 3, 4]. Although there is an appreciable scatter a definite trend exists that is given approximately by $\log_{10} M_o = 1.45M + 16.0$, where M stands for $M_L \leq 6$ or $M_s \geq 6$.

For a transient, band-limited, ground acceleration the amplitude of the high-frequency end of the SD spectrum (relative displacement spectrum) tends to $(T/2\pi)^2$ times the peak ground acceleration, where T is the undamped natural period of an oscillator, while the long-period end of the SD spectrum tends to the maximum ground displacement. Depending on the degree of conservatism desired, one may take the average or the maximum of the observed accelerations at a given distance from a fault (Figure 3) to find the high-frequency amplitude of the pseudo relative velocity spectrum $PSV \equiv (2\pi/T)SD$. The long-period amplitude of the same spectrum would be $2\pi u/T$ for a station essentially at the fault. For a distant station using the approximate pulse shape shown in Figure 1 and approximating the dislocation surface A with a disc of radius r, it can be shown that for $R >> r$, $u_{max} = [0.069/(\mu rR)] \times 10^{1.45+16}$. Thus at large distances, where only the dynamic field contributes to the ground displacement this result can be used to approximate the maximum amplitude associated with S-wave motion.

The above characterization of earthquake spectra, of course, gives only the general spectral trends for body waves in an elastic homogeneous infinite space. Although this characterization may be used as the first order approximation of Fourier and response spectra of strong ground motion, it must be remembered that local geologic formations may change the spectral amplitudes appreciably through reflection, refraction, scattering, and focusing effects, and through wave-guide phenomenon leading to surface waves. Since destructive earthquake ground motion is largely involved with the near-field phenomenon the above model applies directly to a large group of practical problems.

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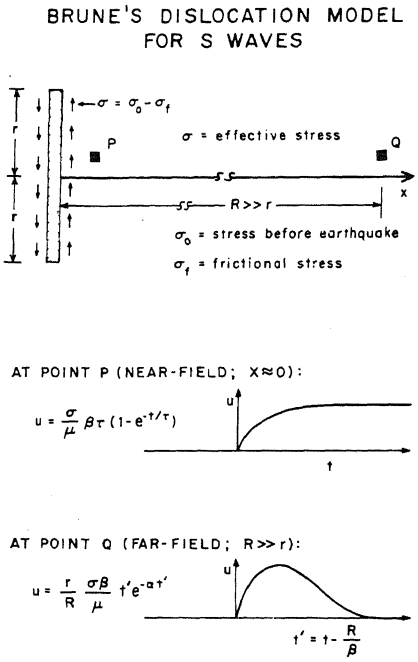
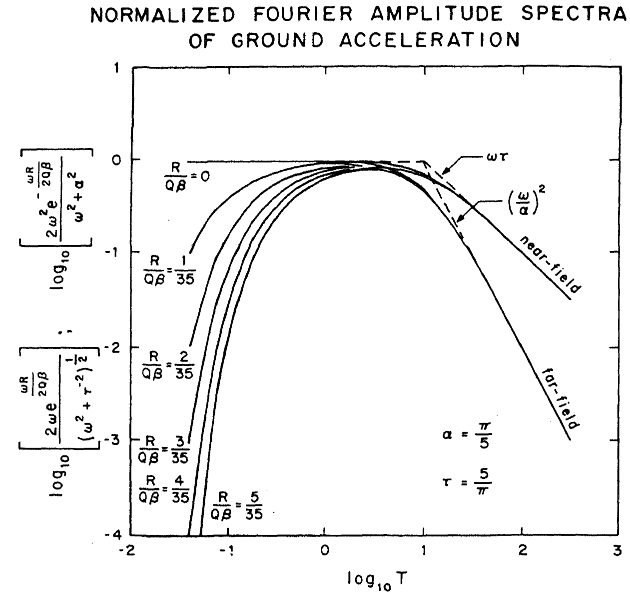
TABLE I

Type of faulting and fault geometry	η_{max}	r represents
Dip-slip displacement along an infinitely long narrow strip in a uniform shear field	$\frac{3}{4}$	fault width
Infinitely long vertical surface fault with strike slip displacement	2	fault width
Circular fault plane in an infinite medium	$\frac{12}{7\pi}$	radius of circular dislocation
Elliptical fault plane $x^2/a^2 + y^2/b^2 \leq 1$; ν depends on the ratio a/b, direction of faulting and elastic constants, $1 < \nu < 2$.	$\frac{2}{\nu}$	$r = b$

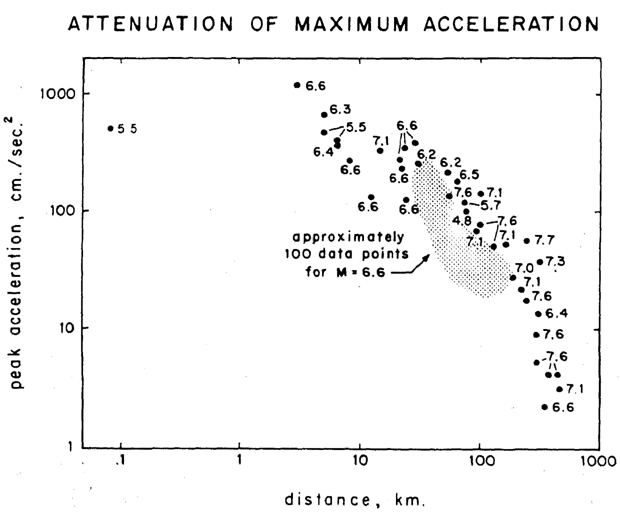
(figure 1)



(figure 2)



(figure 3)



(figure 4)