PREDICTION OF STRONG MOTION USING PHYSICAL MODELS OF EARTHQUAKE FAULTING

K. Aki (I)
Presenting Author: K. Aki

SUMMARY

Physical parameters of heterogeneous fault model determined by Papageorgiou and Aki[1] from acceleration spectra observed for several California earthquakes were discussed in comparison with scaling law of earthquake source spectra and fault behaviors obtained from geological and paleoseismological studies. The characteristic barrier interval and end zone size obtained from strong motion are consistent with the observed departure of scaling law from self-similarity for global earthquakes. Stable barriers and asperities may serve as a unifying physical model for strong motion simulation and the estimation of recurrence of earthquakes from a given fault segment.

INTRODUCTION

Strong motion seismology has a long history of development, owing mainly to the effort of earthquake engineers with emphasis on data collection and interpretation based primarily on empirical approaches. In the past decade or so, seismologists began to get involved in data collection and analysis, and several attempts have been made to develop quantitative models of the earthquake rupture process with the ultimate goal of predicting strong ground motion for a given potential earthquake fault on the basis of an understanding of basic physical laws governing fault mechanics. The problem is difficult because of the sensitivity of high-frequency waves to the details of fault plane heterogeneities. To overcome this difficulty, several attempts have been made to introduce a hybrid of deterministic and stochastic models, in which the gross features of rupture propagation are specified deterministically but the details of the rupture over a heterogeneous fault plane are described by stochastic models.[1-8]

MODELING A HETEROGENEOUS FAULT PLANE

There are two extreme ways to model such a heterogeneous fault plane as illustrated schematically in Fig. 1, which compares the state of a fault plane before and after an earthquake. The shaded region is stressed, and the blank region is slipped. The completely shaded fault plane at the upper left corner corresponds to a uniformly stressed fault, while the completely blank one at the lower right corner represents a smoothly slipped fault without any unbroken patches.

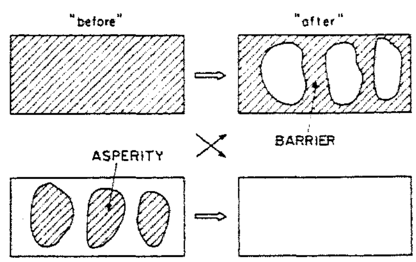

Fig. 1. Asperities and barriers as two extreme models of heterogeneous fault planes.

(I) Robert R. Shrock Professor in Earth and Planetary Sciences, Massachusetts Institute of Technology, Massachusetts,USA.

the upper right of Figure 1, we show the fault plane containing unbroken strong patches after an earthquake. Numerical experiments[9,10] showed that a shear rupture can propagate leaving behind unbroken patches. These patches are called "barriers" and are used for modeling strong motion observed for various earthquakes[11,12]. This model explains the occurrence of after-shocks as the release of stress concentration around barriers through static fatigue[13-15] and offers a physical mechanism for a stress-roughening process to explain the observed stationary frequency-magnitude relation.[16]

On the lower left of Figure 1, we show the fault plane containing strong patches under stress surrounded by a region where stress has already been released by pre-slips and foreshocks. These strong patches are called "asperities" by Kanamori and Stewart[17], and they considered the breaking of these asperities as the model of faulting during the Guatemala earthquake of 1976. In their model, the stress is heterogeneous before the mainshock and becomes homogeneous afterwards. Thus, this model represents the mainshock as a stress-smoothing process in contrast to the barrier model in which the mainshock is considered as a stress-roughening process.

Despite the difference of stress roughening vs. smoothing, the barrier model[1] and the asperity model[18] cannot be distinguished by the high-frequency seismic radiation, because, according to Madariaga[19], both give rise to a flat high-frequency spectrum of the type found by Hanks and McGuire[20] in most accelerograms.

Papageorgiou and Aki[1] considered a specific barrier model of a rectangular fault filled with circular cracks and determined the model parameters using the observed acceleration power spectra for several California earthquakes, as shown in Table 1. The results have turned out to be rather encouraging for strong motion prediction. The diameter of the circular crack 2ρ (or barrier interval) was found to be nearly proportional to the maximum slip Δu_{max} of an earthquake. This means that the "local stress drop" $\Delta\sigma$ inside the circular crack (or within a barrier interval) is a stable parameter, ranging from 200 to 400 bars for all the earthquakes studied.

TABLE 1. SOURCE PARAMETERS OF BARRIER MODEL

EVENT	M_S	M_L	m_b	$M_o \times 10^{27}$ (dyn-cm)	L (km)	W (km)	$\Delta\sigma$ (bars)	2ρ (km)	d (km)	f_{max} (Hz)
Kern Country, 1952	7.7	7.2		2.0	70.	20.	350.	13.	1.0	2.5
San Fernando, 1971	6.6	6.35	6.2	0.12	20.	14.	300.	5.	0.5	5.0
Borrego Mtn., 1968	6.7	6.8	6.1	0.063	33.	11.	200-300.	2.-3.	0.6	4.0
Long Beach, 1933	6.25	6.43		0.028	30.	15.	220.	1.	0.6	4.0
Parkfield, 1966	6.5	5.5	5.9	0.014	35.	15.	200-300	1.-2.	0.5	5.0

The observed constancy or stability of local stress drop for a variety of earthquakes can be explained[1,11] by a slip-weakening model of fracture criterion in which the cohesive stress (= stress to break the bond) is roughly constant (within a factor of about 2) but the critical weakening slip (= slip necessary to break the bond completely) varies among earthquakes by orders of magnitude.

The relative constancy of local stress drop found by Papageorgiou and Aki is qualitatively consistent with the result obtained by Hanks and McGuire[20] using the data on root mean square acceleration (α_{rms}) and the method based on Brune's model[21] and developed by Hanks.[22]

The local stress drop (or its associated slip velocity) determines the level of the Fourier acceleration spectrum and the corresponding power spectrum. The evaluation of peak acceleration or rms acceleration, however, requires another parameter that defines the bandwidth of the acceleration spectrum. For example, if we use the model of Papageorgiou and Aki[1], which is based on the solution of Sato and Hirasawa[23] for a circular crack, the level of the acceleration power spectrum P_o observed at a distance r_o from an earthquake source can be written as

$$P_o \sim \text{const} \cdot W \cdot V \cdot v^2 \cdot \left(\frac{\Delta\sigma}{\mu}\right)^2 \left(\frac{1}{\beta r_o}\right)^2 , \qquad (1)$$

where W is the fault width, V is the velocity of the rupture front along the fault length (sweeping velocity), v is the velocity of rupture spreading within a circular crack, $\Delta\sigma$ is the local stress drop, μ is the rigidity, and β is the shear wave velocity.

If the acceleration spectrum is band-limited between f_o and f_{max},[24] the rms acceleration α_{rms} can be written[25] as

$$\alpha_{rms} = \{P_o \, 2(f_{max} - f_o)\}^{1/2} . \qquad (2)$$

If $f_{max} >> f_o$, we get approximately

$$\alpha_{rms} = \{P_o \, 2f_{max}\}^{1/2} . \qquad (3)$$

From (1) and (3), we find that two major factors controlling the acceleration amplitudes are the local stress drop $\Delta\sigma$ and the cut-off frequency f_{max}.

SCALING LAW OF EARTHQUAKE SOURCE SPECTRA

A direct study of the cut-off frequency f_{max} of acceleration spectra has been done only for a small number of earthquakes. As shown in Table 1, the values of f_{max} for several California earthquakes (M_L ranges from 5.5 to 7.2) are nearly constant at about 4~5 Hz, except for the largest one (Kern County) which showed f_{max} of 2.5 Hz. We would like to know if this apparent stability of f_{max} is a universal phenomenon.

One source of information relevant to this problem is the so-called "source spectra" and their scaling law--that is, how they change with earthquake size. The source spectra may be defined as the Fourier transform $\dot{M}_o(\omega)$ of the time-derivative of the seismic moment function $\dot{M}_o(t)$ of an equivalent point source of an earthquake. If the medium between the earthquake and the observer is homogeneous, the far-field body wave displacement will have the form of $\dot{M}_o(t)$. Thus the source spectrum can be considered as proportional to the Fourier transform of the far-field body wave displacement to be observed from the earthquake, if the earthquake source were placed in a homogeneous medium with properties similar to the source volume.

An extensive survey of the scaling law of earthquakes on a global scale was done by Gusev[26] who gave the result in terms of acceleration spectra. Fig. 2 shows Gusev's scaling law in terms of more familiar displacement source spectra. This result is based not only on the relation among various magnitude scales including 1 Hz band ChISS magnitude, m_{pv}^{sk}, m_{pv}^{skM}, M_{LH}, M_S and M_{JMA} [27], but also accelerogram spectra for the western United States compiled by Trifunac[28].

Fig. 2 demonstrates the existence of a pronounced knee in the source spectra for large (M ~ 8) earthquakes at a frequency around 0.2 Hz. At frequencies above 10 Hz, the observed accelerogram spectra fall down sharply. Thus, we see a bump in the source spectrum between about 0.2 and 10 Hz. This bump may be interpreted as the presence of strong heterogeneity on the fault plane with a range of scale length from about 20 km to a few hundred meters.

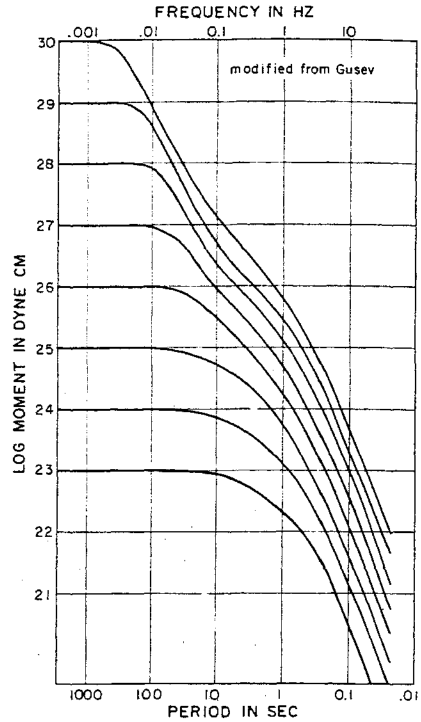

Fig. 2. Scaling law of source spectra based on global data of various magnitude scales. Adapted from Gusev (1983).

There are two independent observations supporting the above interpretation. One is the observed frequency dependence of the apparent attenuation of Q^{-1} of S waves in the lithosphere[29,30]. The Q^{-1} curve peaks at a frequency between 0.1 and 1 Hz, depending on the region. A tectonically active region appears to show a greater Q^{-1} peak at a relatively higher frequency. The observed frequency dependence of Q^{-1} has been explained by Sato[31] as a loss due to scattering in a heterogeneous random medium. The heterogeneities causing the attenuation of S waves by scattering may be related to those in the fault responsible for strong motion.

The other relevant observation is the tendency for the corner frequency of very small earthquakes ($M < 2 \sim 3$ or $M_0 < 10^{19-20}$) in several areas to become constant at around 10 Hz over a considerable range of seismic moment. This has been observed by Chouet et al.[32] for various regions of the earth, and in fact was one of the motivations for developing the barrier model of Aki et al.[12]. Similar results have been observed also by Rautian and Khalturin[33] for earthquakes in the Garm region, U.S.S.R., by Tucker and Brune[34] for San Fernando aftershocks, by Bakun et al.[35] for central California earthquakes, by Fletcher[36] for Oroville, California aftershocks, by Spottiswoode and McGarr[37] for South Africa mine tremors, and by Archuleta et al.[38] for the Mammoth Lakes, California, earthquake sequence.

The corner frequency cannot remain constant indefinitely for smaller earthquakes as we know from the pioneering work of Asada[39] that ultra-microearthquakes contain waves with frequencies around 100 Hz. So, the corner frequency must eventually increase beyond 10 Hz as the earthquake size decreases in any of the above regions. But the fact that it remains constant over a considerable range of seismic moment is consistent with our interpretation that the fault heterogeneity responsible for seismic generation decreases sharply for scale-lengths less than a few hundred meters (corresponding to about 10 Hz). This limiting scale-length may be related to the cut-off frequency f_{max} for acceleration spectra of large earthquakes. The cohesive force hypothesized by Aki[11] in the end zone of a fault rupture front may act over relative motions among blocks a few hundred meters in size or even larger. This suggests an interesting possibility that the detailed features of fault traces on the scale of a few hundred meters observed on the surface such as splaying and branching may play an important role in determining the rupture process relevant to strong motion.

PALEOSEISMOLOGY AND STRONG MOTION

Recent progress in Paleoseismology[40,41] has led Schwartz and Coppersmith[42] to propose an earthquake recurrence model called the "characteristic earthquake model." Earthquake recurrence has usually been modelled by the Gutenberg-Richter frequency-magnitude relation. The data from paleoseismology, however, suggests that a given fault segment generates characteristic earthquakes having a very narrow range of magnitudes and that the Gutenberg-Richter relation does not apply to individual fault segments. A similar conclusion was obtained by Wesnousky et al.[43], who compared the 400-year historical record of seismicity with geologically determined slips on Quaternary faults in Japan.

The characteristic earthquake model may be interpreted in terms of the stability of asperities and barriers on faults. Recent seismological observations suggest that there may be two distinct types of characteristic earthquakes, which Japanese seismologists call by a different name, "earthquake family." In one of them, both the fault length and the slip are stable parameters of a given fault segment. In the other, the slip varies by more than an order of magnitude while the length is stable.

The concept of "earthquake families" or "characteristic earthquakes" defies the self-similarity of earthquake phenomena which has been known to hold over a large range of magnitude for the ensemble of earthquakes sampled without regard to specific faults or fault zones. As elaborated by Andrews[5] self-similarity should apply to earthquake phenomena because there is no unique scale length describing the heterogeneity of faults. For a large ensemble of earthquakes originating from a variety of faults, therefore, we expect that self-similarity will hold. For earthquakes from a specific fault or fault zone, however, there may be a departure from self-similarity because of the existence of heterogeneous scale lengths such as the barrier interval.

Thus, there is an exciting possibility that both recurrence behavior and strong motion may be explained by a unifying fault model, and that geological and seismological information about recurrence behavior may be used to predict strong motion or vice versa.

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REFERENCES

1 Papageorgiou, A. and K. Aki, A specific barrier model for the quantitative description of inhomogeneous faulting and the prediction of strong ground motion, Part I Description of model, Bull. Seis. Soc. Am., 73, 693-722, Part II Applications of the model, Bull. Seis. Soc. Am., 73, 953-978, 1983.

2 Blandford, R.R., A source theory for complex earthquakes, Bull. Seis. Soc. Am., 65, 1385-1405, 1975.

3 Boore, D.M. and W.B. Joyner, The influence of rupture incoherence on seismic directivity, Bull. Seis. Soc. Am., 68, 283-300, 1978.

4 Hanks, T.C., b-value and $\omega^{-\gamma}$ seismic source models: implications for tectonic stress variations along active crustal fault zones and the estimation of high-frequency strong ground motion, J. Geophys. Res., 86, 2235-2242, 1979.

5 Andrews, D.J., A stochastic fault model, 2. Time-dependent case, J. Geophys. Res., 86, 10821-10834, 1981.

6 McGarr, A., Analysis of peak ground motion in terms of a model of inhomogeneous faulting, J. Geophys. Res., 86, 3901-3912, 1981.

7 Boatwright, J., A dynamic model for far-field accelerations, Bull. Seis. Soc. Am., 72, 1049-1068, 1982.

8 Faccioli, E., Measures of strong ground motion derived from a stochastic source model, Soil Dynamics and Earthquake Engineering, 2, 135-149, 1983.

9 Das, S. and K. Aki, Fault planes with barriers: a versatile earthquake model, J. Geophys. Res., 82, 5648-5670, 1977.

10 Mikumo, T., and T. Miyatake, Dynamic rupture process on a three-dimensional fault with non-uniform frictions and near-field seismic waves, Geophys. J. Roy. astr. Soc., 54, 417-438, 1978.

11 Aki, K. Characterization of barriers on an earthquake fault, J. Geophys. Res., 84, 6140-6148, 1979.

12 Aki, K., M. Bouchon, B. Chouet, and S. Das, Quantitative prediction of strong motion for a potential earthquake fault, Annali di Geofisica, 30, 341-368, 1977.

13 Mogi, K., Study of elastic shocks caused by the fracture of heterogeneous material and its relation to earthquake phenomena, Bull. Earthq. Res. Inst., Tokyo Univ., 40, 125-173, 1962.

14 Otsuka, M., A simulation of earthquake occurrence, 5. An interpretation of aftershock phenomena (in Japanese), Zisin, Ser. 2, 29, 137-146, 1976.

15 Das, S. and C.H. Scholz, Theory of time dependent rupture in the Earth, J. Geophys. Res., 86, 6039-6051, 1981.

16 Andrews, D.J., Coupling of energy between tectonic processes and earthquakes, J. Geophys. Res., 83, 2259-2264, 1978.

17 Kanamori, H. and G.S. Stewart, Seismological aspects of the Guatemala earthquake of Feb. 4, 1976, J. Geophys. Res., 83, 3427-3434, 1978.

18 Rudnicki, J. and H. Kanamori, Effects of fault interaction on moment, stress drop, and strain energy release, J. Geophys. Res., 86, 1785-1793, 1981.

19 Madariaga, R., High frequency radiation from dynamic earthquake fault models, Annales Geophysicae, 1, 17-23, 1983.

20 Hanks, T.C., and R.K. McGuire, The character of high frequency strong ground motion, Bull. Seis. Soc. Am., 71, 2071-2096, 1981.

21 Brune, J.N., Tectonic stress and spectra of seismic shear waves from earthquakes, J. Geophys. Res., 75, 4997-5009, 1970.

22 Hanks, T.C., Seismological aspects of strong motion seismology, Proc. 2nd U.S. National Conference on Earthquake Engineering, Earthquake Engineering Institute, pp. 898-912, 1979.

23 Sato, T. and T. Hirasawa, Body wave spectra from propagating shear cracks, J. Phys. Earth, 21, 415-431, 1973.

24 Hanks, T.C., f_{max}, Bull. Seis. Soc. Am., 72, 1867-1880, 1982.

25 Aki, K. and P.G. Richards, Quantitative Seismology, Vol. 1, 493, 1980.

26 Gusev, A.A., A descriptive statistical model of earthquake source radiation and its application to an estimation of short-period strong motion, Geophys. J. Roy. astr. Soc., 74, 787-808, 1983.

27 Koyama, J., M. Takemura, and Z. Suzuki, A scaling model for quantification of earthquakes in and near Japan, Tectonophysics, 84, 3-16, 1982.

28 Trifunac, M.D., Preliminary analysis of the peaks of strong earthquake ground motion, dependence of peaks on earthquake magnitude, epicentral distance and recording site conditions, Bull. Seis. Soc. Am., 66, 189-219, 1976.

29 Aki, K. Scattering and attenuation of shear waves in the lithosphere, J. Geophys. Res., 85, 6496-6504, 1980.

30 Aki, K., Attenuation and scattering of short-period seismic waves in the lithosphere, in Identification of Seismic Sources - Earthquake or Underground Explosion (eds. E.S. Husebye and S. Mykkeltveit), 515-541, D. Reidel Publishing Co., 1981.

30 Aki, K., Attenuation and scattering of short-period seismic waves in the lithosphere, in Identification of Seismic Sources - Earthquake or Underground Explosion (eds. E.S. Husebye and S. Mykkeltveit), 515-541, D. Reidel Publishing Co., 1981.

31 Sato, H, Attenuation of S waves in the lithosphere due to scattering by its random velocity structure, J. Geophys. Res., 87, 7779-7785, 1982.

32 Chouet, B., K. Aki and M. Tsujiura, Regional variation of the scaling law of earthquake source spectra, Bull. Seis. Soc. Am., 68, 49-79, 1978.

33 Rautian, T.G. and V.I. Khalturin, The use of coda for determination of the earthquake spectrum, Bull. Seis. Soc. Am., 68, 923-948, 1978.

34 Tucker, B.E. and J.N. Brune, S wave spectra and source parameters for aftershocks of the San Fernando earthquake of Feb. 9, 1971, in Geological and Geophysical Studies, Vol. 3, San Fernando Earthquake of Feb. 9, 1971, National Oceanic and Atmospheric Administration, Washington, D.C., 1973.

35 Bakun, W.H., C.G. Bufe and R.M. Stewart, Body wave spectra of central California earthquakes, Bull. Seis. Soc. Am., 66, 439-459, 1976.

36 Fletcher, J.B., Spectra from high-dynamic range digital recordings of Oroville, California, aftershocks and their source parameters, Bull. Seis. Soc. Am., 70, 735-755, 1980.

37 Spottiswoode, S.M., and A. McGarr, Source parameters of tremors in a deep-level gold mine, Bull. Seis. Soc. Am., 65, 93-112.

38 Archuleta, R.J., P. Spudich, E. Cranswick, and C. Mueller, Source parameters of the 1980 Mammoth Lakes, California, earthquake sequence, J. Geophys. Res., 4595-4608, 1982.

39 Asada, T., Observations of nearby microearthquakes with ultrasensitive seismometers, J. Phys. Earth, 5, 83-113, 1957.

40 Wallace, R.E., Active faults, paleoseismicity and earthquake hazards in the western United States, in Earthquake Prediction: An International Review, Maurice Ewing Series, Vol. 4, ed., D.W. Simpson and P.G. Richards, American Geophysical Union, pp. 279-289, 1981.

41 Sieh, K.E., A review of geological evidence for recurrence times of large earthquakes, in Earthquake Prediction: An International Review, Maurice Ewing Series, vol. 4, ed. D.W. Simpson and P.G. Richards, American Geophysical Union, pp. 181-207, 1981.

42 Schwartz, D.P., and K.J. Coppersmith, The characteristic earthquake model, J. Geophys. Res., in press, 1983.

43 Wesnousky, S.G., C.H. Scholz, K. Shimazaki, and T. Matsuda, The occurrence rate of earthquakes on Quaternary faults in interplate Japan (abstract), AGU Chapman Conference on fault behavior and the earthquake generation process, Snowbird, Utah, October, 11-15, 1982.