RECENT SEISMOLOGICAL DEVELOPMENTS RELATING TO EARTHQUAKE HAZARD

JAMES N. BRUNE * and
CINNA LOMNITZ**

RESUMEN

Se discuten los aportes recientes de la sismología en relación a la estimación de velocidades y aceleraciones máximas de terremotos, el desplazamiento potencial en diversos puntos de una falla, la actividad sísmica a largo plazo, la estadística de temblores y el riesgo sísmico.

ABSTRACT

Recent seismological evidence is discussed as it relates to estimating earthquake strong motion, potential slip at various points along a fault, long term seismic pattern, earthquake statistics, and earthquake hazard.

* *Institute of Geophysics and Planetary Physics, University of California.*

** *Instituto de Geofísica, UNAM.*

INTRODUCTION

The most important recent seismological development relating to earthquake hazard is the understanding of the role of stress in determining the character of near-source ground motion. This understanding allows us to make quantitative estimates of the peak accelerations and velocities near faults and also to predict how these motions are likely to vary with magnitude.

Another recent seismological development comes from the New Global Tectonics, the simplified tectonic model in which earth deformation results primarily from the interaction of a limited number of lithospheric plates moving about on a more or less weak substratum. This process has been called "plate tectonics". The knowledge of the long term motion of the plates combined with slip resulting from earthquakes allows a quantitative estimate of earthquake potential at various points along a fault or plate boundary. Sections which have been "locked" for a long time while adjacent sections are slipping present the greatest potential for earthquakes.

Also important in estimations of earthquake hazard are recent improvements in our understanding of what constitutes an "active" fault and of the time scale necessary for estimation of long term seismic behavior from a statistical study of earthquakes. The use of simple statistical models leads to reliable predictions concerning earthquake risk in a region for a given design period. The margin of uncertainty due to the restricted availability of data can also be estimated.

STRESS AND NEAR-SOURCE STRONG GROUND MOTION

Various studies over the past five years have led to better estimates of the stress changes involved in major earthquakes. It appears that for major earthquakes the stresses available for generating strong near-source ground motion are approximately the same as the stress drop which seismologists have been measuring at large distances from earthquakes and which also can be estimated from permanent ground deformation near the earthquake fault. These stresses may be quite variable but are commonly of the order of 100 bars (10^8 dynes/cm^2). Accepting this value for accelerating stress on the faults of large earthquakes, we can estimate the maximum values for ground velocity and acceleration. Following the results of Brune (1970) these are given as follows:

$$\dot{u}\ (\max) = \frac{\sigma}{\mu}\ v_s \qquad (1)$$

$$\ddot{u}\ (\max) = \frac{1}{\pi}\ \frac{\sigma}{\mu}\ v_s\ \omega_s \qquad (2)$$

In these equations u, $\dot{u}$, and $\ddot{u}$ are the particle displacement, velocity and acceleration respectively; σ is the stress, μ the shear modulus, v_s the shear wave velocity and ω_s the cutoff frequency above which the seismic energy is either attenuated or becomes incoherently scattered. Inserting values of π = 100 bars, v_s = 3 km/sec., $\mu = 3 \times 10''$ dynes/cm^2 and ω_s = 10 cps we obtain:

$$\dot{u}\ (\max) = 100\ \text{cm/sec} \qquad (3)$$

$$\ddot{u}\ (\max) \cong 2\ \text{g} \qquad (4)$$

If ω_s is 5 cps:

$$\ddot{u}\ (\max) = 1\text{g}, \qquad (5)$$

where g is the acceleration of gravity.

These results compare surprisingly well with existing strong motion observations. Peak velocities observed from the Parkfield earthquake of June 27, 1966 were 76 cm/sec and those from the recent San Fernando earthquake of February 9, 1971 were about 125 cm/sec. Similarly peak accelerations recorded from the Parkfield earthquake were 0.5g and those from the San Fernando Earthquake about 1.25g.

Thus, the strong motion observations seem consistent with predictions based on the stress pulse model and give confidence that we understand the forces involved in earthquakes.

There are some important consequences of this result as far as earthquake hazard is concerned:

1. Accelerations of greater than 1g at high frequencies ($\sim$10 cps) may be expected on hard rock ($v_s \sim 3$ km/sec) near faults.

Such accelerations could occur for quite small earthquakes although the volume affected by such high accelerations would be much smaller for small earthquakes than for large ones. Similarly, on thick deposits of lower-rigidity rocks, the upper limit on accelerations would be considerably smaller, perhaps 1/2 g. Of course, for hard rock a considerable distance away from the rock in which the fault occurs, the strong motion will be less than for adjacent softer rocks because of ground amplification.

2. Peak particle velocities near 100 cm/sec at periods near 1/2 to 1 sec may be expected to be quite common near large earthquakes.

3. The region affected by large ground motions is determined by the length of the rupture and a width determined by the depth of faulting, but the very highest accelerations will be expected near the fault plane.

PLATE TECTONICS AND EARTHQUAKE POTENTIAL

Assuming that, as a result of tectonic and geologic studies, we determine the long-term steady state velocities of two adjacent plates and consequently the long-term rate of slip along a given fault zone we may set up the following simplified equation for earthquake potential slip:

$$\text{Potential slip} = \text{Plate motion} - \text{Earthquake slip} - \text{Aseismic slip} \qquad (6)$$

The plate motion is given. Earthquake slip along a given zone may be determined by summing the moments of various earthquakes

(Brune, 1968). Moment is an easily determined parameter and is proportional to the area, A, of a fault zone multiplied by the average slip, $\overline{u}$, over that zone during the earthquake:

$$M_o = \mu \, A \, \overline{u} \approx \text{amplitude of long period waves} \tag{7}$$

Aseismic slip may only be determined by direct observation. Its importance has not been established in most shallow earthquake zones.

Given a long-term sample of seismicity we then estimate the potential slip along various sectors of a fault and thus estimate the seismic hazard. There are a number of areas in the world where a given section of a major fault has been "locked" for a long time and for which the long-term average motion is expected to be high and indeed for which adjacent sections of the fault have been moving quite rapidly. Such areas obviously have a high potential slip and constitute a high seismic hazard.

The major difficulty besetting such determinations of potential slip is the small statistical sample which the history of quantitative earthquake recording represents. For example, an area such as that described above which is calculated to have a high potential slip may in fact have had a very large but unrecorded earthquake just prior to the historic record and thus be considerably less hazardous than expected. It is easy to verify that for most fault zones the historic record of seismicity does not give an adequate view of long-term seismicity. Major faults tend to be relatively quiet for periods of the order of one to a few centuries and then have a period of high activity. This is illustrated by the study of Davies and Brune (1971). Figures 1 and 2 show sums of seismic moment taken from their paper for the world as a whole and also for specific areas. It is immediately obvious from these data that the historic record is not adequate to represent the long term seismicity either for the world as a whole or for the major fault zones. In the period near 1900 the worldwide activity was very high wheareas in recent years it has been very low. It is probable that 1000 years of seismicity data would be a minimum sample for many areas. It nevertheless was shown by Davies and Brune that the slip rates determined by the historic sample are in fairly good agreement with slip rates predicted by plate tectonics; this is evidence that the historic record of seismicity is not too uncharacteristic of the long term pattern of seismicity.

INFERRING EARTHQUAKE HAZARD FROM SEISMICITY AND FAULTING

Improved knowledge of faulting mechanism as well as better seismicity data allow us to state some general principles regarding the estimation of earthquake hazard in regions for which we do not have an adequate long-term seismic record.

Faulting

Although there are many faults that are inactive or "dead", faulting is necessary to produce earthquakes and, as far as we know, all severe earthquakes have ocurred on pre-existing faults. Logically, there must be occasions when earthquakes occur in fresh rock not previously faulted, but the odds against this are so great that it represents a negligible hazard. We can assume that the existence of faulting is a necessary but not sufficient condition for high earthquake hazard. Of course, there are many faults which are hidden by recent sediments or by water and we cannot necessarily assume, because no faults are mapped in a certain area, that none exist. The tasks of finding hidden faults and of determining which faults are active are of prime importance to the geologist working on the problem of earthquake hazard.

Seismicity

Almost all of the world's earthquakes occur in narrow belts. Outside of these belts the earthquake hazard is low. We can assume that where there have been large earthquakes before there will be large earthquakes again, and conversely, where there have been no earthquakes for millions of years there will be no earthquakes in the near future. Unfortunately, the historic earthquake record is not long enough to establish the pattern of seismicity in most areas. In some countries, such as Japan, a fair record exists for the last 1 400 years, but in most regions the historic earthquake record is only a few hundred years old, and precise observations on magnitude, location and faulting have been available for only several decades. Consequently, we cannot conclude, because an area has had no historic earthquakes, that there is no seismic hazard.

Use of smaller earthquakes

Always associated with large earthquakes are much more numerous small ones, roughly so that earthquakes one magnitude unit smaller, occur ten times more frequently. In many areas we can record more than twenty small earthquakes per day. This suggests the possibility that we might determine the seismic hazard of an area by counting the number of very small earthquakes and thus not having to wait for the much more infrequent large ones. Unfortunately this does not work out because the pattern of seismicity is not stable on the time scale of the instrumental seismic record. For example, the San Andreas fault near Palmdale is one of the quietest seismic areas in California for small earthquakes although in 1857 one of the largest earthquakes in California occurred there. This earthquake was certainly followed and possibly preceded by a high rate of occurrence of small earthquakes.

We cannot assume that large earthquakes will be preceded by an increase in seismic activity. Although this sometimes happens, many times it does not happen. The recent San Fernando, California, earthquake occurred without any warning. The same is true for the much larger Alaska earthquake of 1964. The area had been quiet for over a hundred years prior to the earthquake. The great Chilean earthquake of 1960, although preceded a few hours and days by large foreshocks, had been anomalously quiet for preceding years. In conclusion, although high historic seismicity is probably a sufficient condition for high seismic hazard, it is not a necessary one.

There seems to be fair evidence that in general, the recurrence probability is decreased by a large earthquake. That is, once a great earthquake has occurred in a given area, it is less likely that another will occur soon after, because the earth will not have had enough time to store up the tremendous amounts of energy necessary; but our knowledge is not yet quantitative enough to place great faith in such reasoning. Of course, immediately after a great earthquake, the seismic hazard remains high for some time because of the occurrence of large aftershocks.

Statistical approaches to earthquake occurrence

The main difficulties in applying statistics to earthquake occurrence have been:

1. Lack of a physical model which can be formulated in statistical terms.

2. Lack of sufficient historical record and sufficient global coverage, especially for small earthquakes. For this reason there is little assurance that statistical parameters derived from the present historical record may be extrapolated into the future.

Despite these limitations, many statistical properties of the earthquake process have been investigated. Recent advances in our understanding of the earthquake mechanism and of plate tectonics promise to overcome many of the difficulties previously inherent in such studies. With this in mind we discuss some recent results of statistical approaches to earthquake occurrence.

Statistical fluctuations of tectonic stress

The distribution of earthquake magnitudes for a given sample of events may be written

$$f(M) = \beta \, e^{-\beta M} \tag{8}$$

Fluctuations in β have been observed in connection with the occurrence of large earthquakes. These fluctuations have been interpreted in various ways. Suyehiro (1966) pointed out that the β–value increased after the 1960 earthquake; he attributed this effect to the production of many small aftershocks, while the rate of incidence of larger earthquakes was about the same before and after the main shock. Mogi (1962) had suggested that β depended on the "heterogeneity" of the material, thus explaining the stationarity of β within a given regions; while Scholz (1968) connected β with the state of stress in a sample, thus emphasizing the variability of β. These observations may be explained by stating that β is the reciprocal value of $\overline{M}$, the mean magnitude of the distributions (8).

Stationarity of $\overline{M}$ within a region was first demonstrated by Lomnitz (1960, 1966), and confirmed by observations made in different regions (Hamada and Hagiwara, 1967; Hamilton, 1966; Drakopoulos, 1971; López-Arroyo and Udías, 1972). Fluctuations in $\overline{M}$ must be attributed largely to fluctuations in tectonic stress, since the distribution of fault sizes in a region may be considered invariant. Hence the observed increase in the β–value after a large earthquake follows directly from the assumption that the regional tectonic stress drops to a lower value. We may therefore use the fluctuations of the mean magnitude of earthquakes in a region as a convenient measure of tectonic stress.

Gibowicz (1973) has found an important difference in the pattern of fluctuations of β for swarms and aftershock sequences. In swarms the β–value decays continuously with time; in aftershock sequences it fluctuates about a stationary value, decaying before each large aftershock and increasing again afterwards. These differences must correspond to entirely different stress patterns in time.

As a first approximation, large earthquakes are independent random events. However, for purposes of predicting the earthquake hazard at a specific location it is important to understand the influence of the occurrence of previous events in the region and in neighboring regions, on the probability of occurrence of future earthquakes.

A suitable linear model of the earthquake process in a region is given by the Boltzmann Process:

$$\text{Prob}\,(t, t+1) = K\,\sigma(t) + \int_{-\infty}^{t} \varphi\,(t-\tau_i)\;d\sigma\;(\tau_i) \tag{9}$$

where σ (t) is the tectonic stress in the region. The memory function φ (t) has the general form

$$\varphi\,(t) \cong H\,(t)\,e^{-\alpha t} \tag{10}$$

Where H is the Heaviside function, and τ_i is the time of occurrence of the i-th event.

The tectonic stress $\sigma(t)$ may be modelled by a birth-and-death process with continuous increments and discrete decrements (Fig. 3). The decrements are the stress drops from earthquakes, which are obtained through a process derived from (8) with β a function of $\sigma(t)$.

If the interoccurrence time between events is very large their interaction may be neglected. In this case one obtains a quasi-stationary Poisson process, with increments given by eq. (8). The probability of ocurrence of an extreme event of magnitude M in a period of D years is given by

$$\text{Prob}\,(M, D) = 1 - \exp\,(-\alpha\,De^{-\beta M}). \tag{11}$$

The use of such statistical models, in good agreement with the observed recurrence patterns of earthquakes, may lead to new insights into the problem of estimating seismic hazards (Lomnitz, 1974). On the other hand, the question of estimating the time and place of the next large earthquake in a region in a deterministic sense must be attacked in a different manner. A combined statistical-mechanical approach may represent the most promising avenue of research in problems of earthquake prediction.

CONCLUSIONS

1. Recent seismological developments relating to earthquake hazard have indicated that near-source accelerations of near 1 g and particle velocities near 100 cm/sec are to be expected near faults of large earthquakes, depending on rock type and geological conditions, and that the region affected by these strong motions is closely related to the dimensions of the faulting.

2. Plate tectonic theory offers the possibility of eventual estimation of potential slip and thus potential hazard along a given section of a fault.

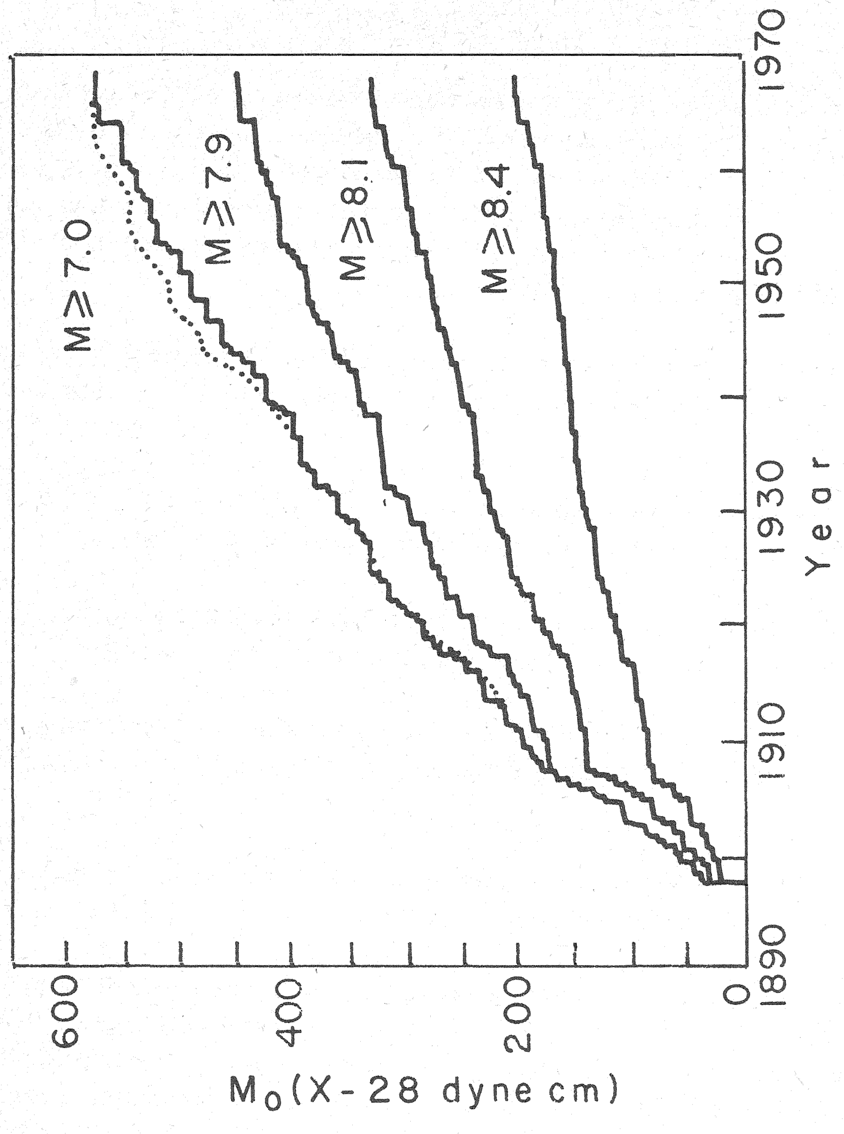
3. For most areas the historic sample of seismicity, by itself, is not sufficient to determine the long term pattern of seismicity.

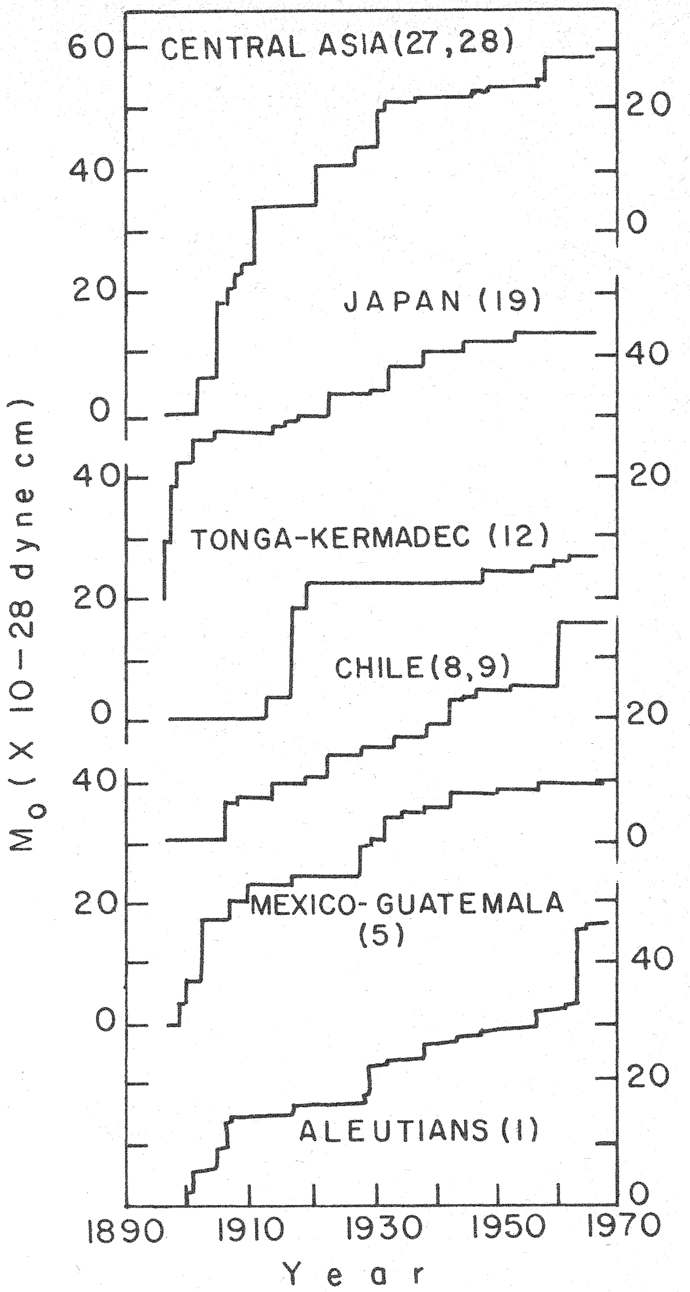
4. In estimating earthquake hazard in areas where the historic sample in not sufficient to establish a long term pattern we conclude:

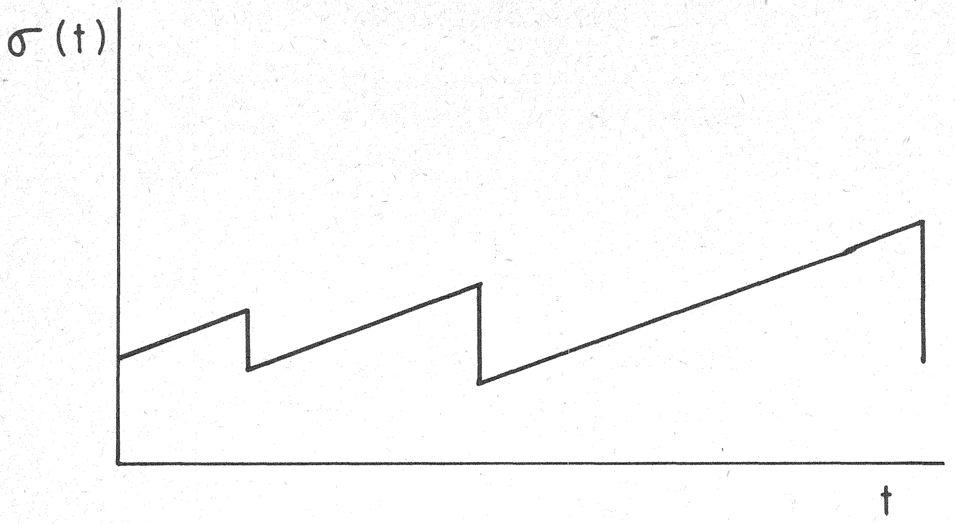
   a) Faulting is a necessary but not sufficient requirement for high seismic hazard. However, not all possibly active faults are known.

   b) Historical high seismicity is a sufficient but not necessary condition for high seismic hazard.

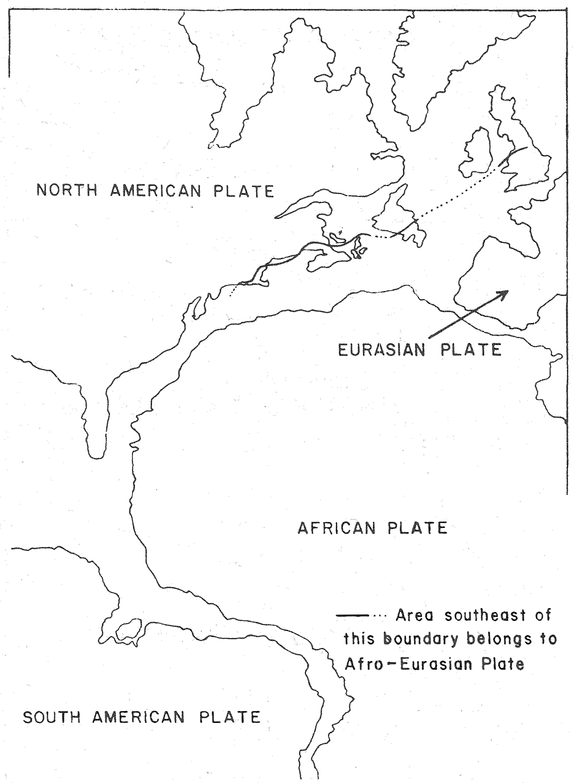
5. Application of probability concepts may be used to determine the statistical parameters of the earthquake process in a region. These statistics may lead to estimates for the earthquake risk within a given design period. A combined statistical-mechanical approach to earthquake hazard may represent a promising avenue of research in earthquake prediction.


M ≥ 7.0
M ≥ 7.9
M ≥ 8.1
M ≥ 8.4
1890
1910
1930
1950
1970
Year
0
200
400
600
Mo(X-28 dyne cm)


CENTRAL ASIA (27, 28)
JAPAN (19)
TONGA-KERMADEC (12)
CHILE (8, 9)
MEXICO-GUATEMALA (5)
ALEUTIANS (1)
60
40
20
0
40
20
0
40
20
0
20
0
40
20
20
0
20
0
40
20
0
M_o (X 10 − 28 dyne cm)
1890
1910
1930
1950
1970
Year



σ (t)
t



NORTH AMERICAN PLATE
EURASIAN PLATE
AFRICAN PLATE
Area southeast of this boundary belongs to Afro-Eurasian Plate
SOUTH AMERICAN PLATE

BIBLIOGRAPHY

BRUNE, J. N., 1968. Seismic moment, seismicity, and rate of slip along major fault zones. *J. Geophys. Res.* 73: 777-784.

BRUNE, J. N., 1970. Tectonic stress and the spectra of seismic shear waves from earthquakes. *J. Geophys. Res.* 75: 4997-5009.

DAVIES, G. F., and J. N. BRUNE, 1971. Regional and Global fault slip rates from seismicity. *Nature Physical Science,* 229: 101-107.

DRAKOPOULOS, J. C., 1971. A statistical model on the occurrence of aftershocks in the area of Greece. *Bull, Int. Inst. Seismol. Earthq. Eng. Tokyo,* 8: 17-39.

GIBOWICZ, S. J., 1973. Variation of the frequency-magnitude relation during earthquake sequences in New Zealand. *Bull. Seis. Soc. Am.,* 63: 517-528.

HAMADA, K., and T. HAGIWARA, 1967. High sensitivity tripartite observations of Matsushiro earthquakes, Part 4. *Bull. Earthq. Res. Inst.,* 45: 159-196.

HAMILTON, R. M., 1966. The Fiordland earthquake sequence of 1960 and seismic velocities beneath New Zealand. *N. Z. J.Geol. Geophys.,* 9: 224-238.

LOMNITZ, C., 1960. A study of the Maipo Valley earthquakes of September 4, 1958. *Proc. 1st. Conf. Earthq. Eng., Tokyo,* 1: 501-520.

LOMNITZ, C., 1966. Magnitude stability in earthquake sequences. *Bull. Seis. Soc. Am.,* 56: 247-249.

LOMNITZ, C., 1974. *Global Tectonics and Earthquake Risk* (Elsevier, Amsterdam, 320 pp).

LOPEZ-ARROYO, A., and A. UDIAS, 1972. Aftershock sequences and focal parameters of the February 28, 1969 earthquake of the Azores-Gilbraltar fracture zone. *Bull. Seis. Soc. Am.,* 62: 699-720.

MOGI, K., 1962. Study of the elastic shocks caused by the fracture of heterogeneous materials and its relation to earthquake phenomena. *Bull. Earthq. Res. Inst.,* 40: 125-173.

SUYEHIRO, S., 1966 Difference between aftershocks and foreshocks in the relationship of magnitude to frequency of occurrence for the great Chilean earthquake of 1960. *Bull. Seis. Soc. Am.,* 56: 185-200.