Reports

The Frequency of Very Large Earthquakes

Abstract. *Observational data relating surface wave magnitude* M_s *to seismic moment* M_0 *are used to convert a well-known frequency-*M_s *plot into a frequency-*M_0 *relationship, which turns out to be remarkably linear. There is no evidence of an upper bound to* M_0*, on the basis of presently available evidence. The possibility exists that extremely large earthquakes* ($M_0 = 10^{31}$ *dyne-centimeters or greater) may occur from time to time.*

Apart from their intrinsic interest, there are several rather practical reasons why we would like to know more about how large earthquakes can be, and how often the very large ones occur. It turns out that both elastic energy release (*1*) and fault slip or plate movement (*2*) due to earthquakes are dominated by the very largest shocks that occur. Meaningful estimates of these quantities are therefore very difficult, unless we assume a well-defined upper bound to earthquake size. (We use the term "size" in a general sense. As we shall see, magnitude as it is normally measured is not an adequate quantity to define the source parameters of very large earthquakes.) In the matter of earthquake risk, we are also concerned with very large events. Even if the probability of such an event is very low, its potential for destruction may be very high, and the net effect is that these infrequent earthquakes must be considered in a realistic assessment of future seismic hazards.

Unfortunately, the amount of reliable information that is available concerning very large earthquakes is rather limited. An adequate global network of seismic stations has existed only for the past 15 years or so. Before this time, problems of station distribution and calibration were severe. And, of course, before the early 1900's there were essentially no instrumental data. We therefore have at most a 70-year record, and for much of this time magnitude determinations were not very reliable. Since all indications are that earthquake frequencies decrease with increasing earthquake size, there are clearly difficulties in attempting to determine the source parameters of earthquakes that occur, on the average, every 50 years or more. We have no alternative but to use all the available data (bypassing numerous questions of reliability) in an attempt to construct a relationship between size and frequency. Then we can discuss the implications of extrapolating this relationship beyond the known data.

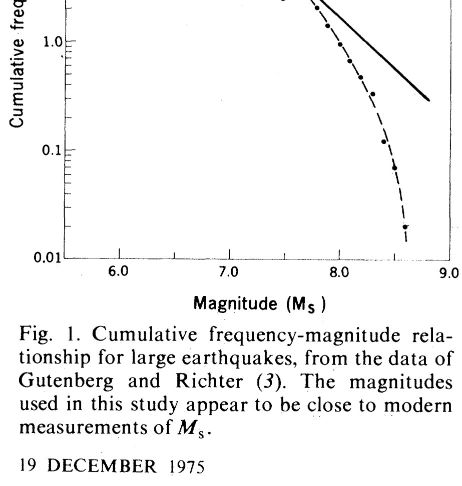

Fig. 1. Cumulative frequency-magnitude relationship for large earthquakes, from the data of Gutenberg and Richter (*3*). The magnitudes used in this study appear to be close to modern measurements of M_s.

The normal method for investigating earthquake frequencies is in terms of a frequency-magnitude plot. Figure 1 shows the cumulative frequency (number of earthquakes with magnitude greater than or equal to a given level) of large shallow earthquakes compiled in the classic study of Gutenberg and Richter (*3*). The data used consist of earthquakes of magnitudes 7.75 to 8.6 for the period 1904 to 1945, those of magnitude 7.0 to 7.7 for the period 1918 to 1945, and those of magnitude 6.0 to 6.9 for the period January 1932 to June 1935. All have been normalized to annual occurrence rates. Subsequent studies (*4*) have produced very similar results and have shown that the magnitudes used by Gutenberg and Richter are very close to modern-day observations of surface wave magnitude, M_s. The "revision" of the original magnitudes of the larger events by Richter (*5*) do not agree with M_s determinations.

The frequency-M_s plot is approximately linear for earthquakes with M_s less than about 7.0, and this linearity is confirmed by many other studies at smaller M_s values. At higher M_s values, however, the relationship departs from a straight line and trends toward the vertical in the general vicinity of $M_s = 8.6$. Many investigators have concluded from this result that earthquakes with M_s greater than 8.6 or so do not occur, and this may well be correct. However, it is not valid to argue that this result necessarily places an upper bound on earthquakes source parameters (fault length, width, and offset). In order to understand this point, we must examine what we mean by M_s and consider two effects that limit its usefulness for very large earthquakes.

Ideally, M_s is a measure of the spectral amplitude at a period of 20 seconds of the seismic radiation from an earthquake, as observed at a standard station at a standard distance. In practice, however, M_s is measured in the time domain, and it is hard to obtain a meaningful estimate of the spectral amplitude when the source emits the radiation for more than one cycle. When an earthquake dislocation propagates for significantly longer than 20 seconds (that is, for fault lengths of 100 km or more), measured M_s values are underestimates of the true spectral amplitudes at 20 seconds, and this effect becomes more important as larger earthquakes are considered. A second and even more important factor arises from the typical shape of the spectrum of seismic waves. This spectrum contains a relatively flat portion at low frequencies, and then, beyond a "corner frequency," the spectral amplitudes fall off rapidly (*6*). The corner frequency is determined by the dimensions of the earthquake source and moves to lower and lower frequencies as the size of the source increases. When M_s is determined from the flat part of the spectrum, it is a good measure of the source dimensions. However, for earthquakes large enough that the cor-

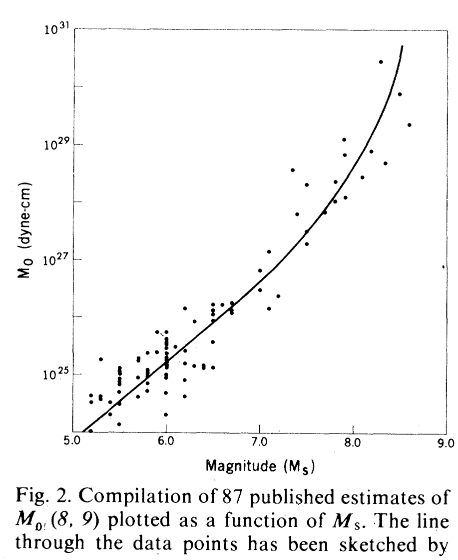

Fig. 2. Compilation of 87 published estimates of M_0 (*8, 9*) plotted as a function of M_s. The line through the data points has been sketched by eye.

ner frequency is less than the measurement frequency (corresponding to a period of 20 seconds for M_s), M_s becomes a very slowly changing measure of earthquake size.

The net result of these two effects is that there is likely to be an upper limit to M_s that will be observed, regardless of the true earthquake size. Clearly, it would be advantageous to measure the size of large earthquakes by some frequency-independent quantity. There is such a quantity, and it is called the seismic moment, M_0. It is related to the zero frequency asymptote of the seismic spectrum, and in geometrical terms it is the product of the fault area multiplied by the fault displacement multiplied by the rigidity modulus (*7*).

There are, as yet, relatively few earthquakes for which we have been able to obtain enough information to calculate M_0. Certainly we have no way to directly plot a frequency-M_0 graph. However, a series of recent papers have provided enough data to permit us to begin to establish the shape of the relationship between M_s and M_0 (*8, 9*). Figure 2 shows 87 determinations of M_s plotted as a function of M_0, for M_s greater than 5.0. Although the scatter is considerable, the general shape of the relationship seems clear, and a provisional curve has been sketched through the data. A roughly linear trend persists up to an M_s of 7.0 or 7.5, and then the curve bends sharply upward. The highest point (Chilean earthquake of 1960) and the next highest (Alaskan earthquake of 1964) have been studied extensively (*9*) and appear quite reliable. There is a definite suggestion that the curve may become vertical in the vicinity of M_s = 8.6 or 8.7, an indication that M_s values greater than this will not be observed, regardless of the value of M_0.

We can use the relationships in Figs. 1 and 2 to make the first attempt at the empirical construction of a frequency-M_0 curve. As shown in Fig. 3, a remarkably linear frequency-M_0 relation is obtained. We have, as yet, no convincing theoretical arguments that predict that this relation should be linear, although a linear relation was postulated by Wyss (*10*). Interestingly, the slope of the line in Fig. 3 (0.61) is identical to the slope of the corresponding relation for the aftershock sequence of the Parkfield earthquake (*10*), although this agreement may be coincidence. Perhaps we should phrase our result as follows: A linear frequency-M_0 relation is entirely consistent with the general shape of frequency-M_s data, given recent observations of M_0 as a function of M_s. It is also consistent with observations of m_b (body wave magnitude measured at a period of 1 second) [a detailed discussion of this point will be presented elsewhere (*11*)].

We may use Fig. 3 to make estimates of

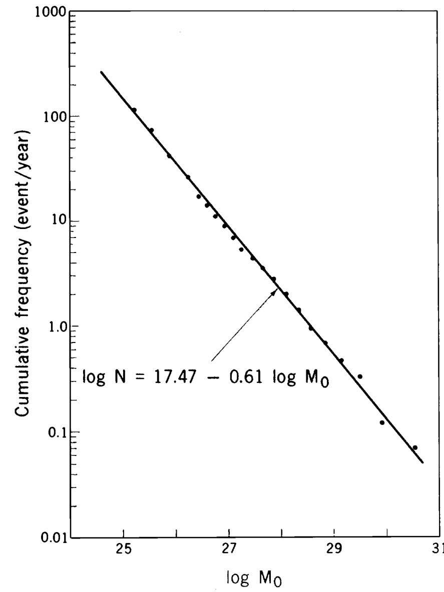

Fig. 3. Cumulative frequency-M_0 graph deduced from the frequency data in Fig. 1 and the M_0-M_s relationship in Fig. 2. The straight line is a least-squares fit to the data.

the frequencies of earthquakes of various M_0 values. It appears, for example, that events with M_0 of 10^{30} dyne-cm or greater occur, on the average, every 10 years or so, and this is entirely consistent with the observation that two large earthquakes with well-determined M_0 values have occurred within the past 15 years. Apparently, events such as the 1960 Chile earthquake ($M_0 = 2.5 \times 10^{30}$ dyne-cm) and the 1964 Alaska earthquake ($M_0 = 7.5 \times 10^{29}$ dyne-cm) are not at all unusual.

More importantly, the linear trend in Fig. 3 continues through the points corresponding to the largest values of M_0 so far measured. There are good reasons for believing that there must be an upper bound to earthquake M_0 values, due to the geometry of seismic zones and the strength of crustal material. However, the data presented here show no indication of where this upper bound might be. Also, since we have accurate M_0 data only for very recent earthquakes, it does not seem reasonable to suppose that this linear trend stops immediately beyond the last data point. If the trend is extrapolated beyond the data, an event of $M_0 = 10^{31}$ dyne-cm is predicted to have a mean return period of 50 years. It is not clear that the record of large earthquakes during the last 100 years is sufficiently detailed that the occurrence of such a catastrophic event can be ruled out. The U.S. Environmental Data Service (*12*) has listed 151 earthquakes with M_s greater than or equal to 8.0 during the period 1897 to 1972, and very few of these have been studied in detail.

Certainly, if earthquakes with M_0 values much larger than 10^{30} dyne-cm do occur, this could have a significant effect on global estimates of seismic energy release and plate motion due to earthquakes, and such large events may well cause a considerable excitation of the Chandler wobble (*13*).

MICHAEL A. CHINNERY
ROBERT G. NORTH
Applied Seismology Group, Lincoln Laboratory, Massachusetts Institute of Technology, Cambridge 02142

References and Notes

1. M. Bath, in *Physics and Chemistry of the Earth*, L. H. Ahrens, F. Press, S. K. Runcorn, H. C. Urey, Eds. (Pergamon, New York, 1966), vol. 7, pp. 115–165.
2. J. N. Brune, *J. Geophys. Res.* **73**, 777 (1968).
3. B. Gutenberg and C. F. Richter, *Seismicity of the Earth and Associated Phenomena* (Princeton Univ. Press, Princeton, N.J., 1954).
4. V. I. Bune and N. V. Golubeva, *Izv. Acad. Sci. USSR Phys. Solid Earth* **7**, 17 (1972); J. P. Rothe, *The Seismicity of the Earth, 1953–1965* (UNESCO, New York, 1969); J. F. Evernden, *Bull. Seismol. Soc. Am.* **60**, 393 (1970).
5. C. F. Richter, *Elementary Seismology* (Freeman, San Francisco, 1958).
6. J. C. Savage, *J. Geophys. Res.* **77**, 3788 (1972).
7. T. Maruyama, *Bull. Earthquake Res. Inst. Tokyo Univ.* **41**, 467 (1963).
8. R. G. North, *Nature (Lond.)* **252**, 560 (1974); A. Ben-Menahem, E. Aboodi, R. Schild, *Phys. Earth Planet. Interiors* **9**, 265 (1975); K. Abe, *ibid.* **7**, 143 (1973); *ibid.* **5**, 367 (1972); K. Aki, *Geophys. J.* **31**, 3 (1972); H. Kanamori, *Phys. Earth Planet. Interiors* **6**, 346 (1972); *Tectonophysics* **12**, 187 (1971).
9. H. Kanamori and D. L. Anderson, *J. Geophys. Res.* **80**, 1075 (19[illegible] H. Kanamori, *ibid.* **75**, 5029 (1970).
10. M. Wyss, *Geophys. J.* **31**, 341 (1973).
11. M. A. Chinnery and R. G. North, in preparation.
12. Earthquake Data File, 1900–1972, available from the U.S. National Oceanic and Atmospheric Administration, Environmental Data Service, Boulder, Colo.
13. F. A. Dahlen, *Geophys. J.* **32**, 203 (1973).
14. Sponsored by the Advanced Research Projects Agency of the Department of Defense.

15 August 1975

Model for Disseminated Cutaneous Leishmaniasis

Abstract. Leishmania *infection of a skin site with interrupted lymphatic drainage results in widespread cutaneous metastases. This model may provide a method for the study of disseminated cutaneous leishmaniasis in man.*

Disseminated cutaneous leishmaniasis is a manifestation of cutaneous leishmaniasis characterized by diffuse skin nodules of parasitized macrophages, anergy to intradermal tests with specific antigen, and resistance to treatment. Although the underlying mechanism responsible for dissemination is largely unknown, it is generally believed to result from specific deficiency of cell-mediated immunity of the host (*1, 2*). With this in mind, several investigators have tried to reproduce disseminated cuta-