The Energy Release in Great Earthquakes

HIROO KANAMORI

Seismological Laboratory, California Institute of Technology, Pasadena, California 91125

The conventional magnitude scale M suffers saturation when the rupture dimension of the earthquake exceeds the wavelength of the seismic waves used for the magnitude determination (usually 5–50 km). This saturation leads to an inaccurate estimate of energy released in great earthquakes. To circumvent this problem the strain energy drop W (difference in strain energy before and after an earthquake) in great earthquakes is estimated from the seismic moment M_0. If the stress drop $\Delta\sigma$ is complete, $W = W_0 = (\Delta\sigma/2\mu)M_0 \sim M_0/(2 \times 10^4)$, where μ is the rigidity; if it is partial, W_0 gives the minimum estimate of the strain energy drop. Furthermore, if Orowan's condition, i.e., that frictional stress equal final stress, is met, W_0 represents the seismic wave energy. A new magnitude scale M_w is defined in terms of W_0 through the standard energy-magnitude relation $\log W_0 = 1.5M_w + 11.8$. M_w is as large as 9.5 for the 1960 Chilean earthquake and connects smoothly to M_s (surface wave magnitude) for earthquakes with a rupture dimension of about 100 km or less. The M_w scale does not suffer saturation and is a more adequate magnitude scale for great earthquakes. The seismic energy release curve defined by W_0 is entirely different from that previously estimated from M_s. During the 15-year period from 1950 to 1965 the annual average of W_0 is more than 1 order of magnitude larger than that during the periods from 1920 to 1950 and from 1965 to 1976. The temporal variation of the amplitude of the Chandler wobble correlates very well with the variation of W_0, with a slight indication of the former preceding the latter. In contrast, the number N of moderate to large earthquakes increased very sharply as the Chandler wobble amplitude increased but decreased very sharply during the period from 1945 to 1965, when W_0 was largest. One possible explanation for these correlations is that the increase in the wobble amplitude triggers worldwide seismic activity and accelerates plate motion which eventually leads to great decoupling earthquakes. This decoupling causes the decline of moderate to large earthquake activity. Changes in the rotation rate of the earth may be an important element in this mechanism.

INTRODUCTION

The energy release in earthquakes is one of the most fundamental subjects in geophysics. In most cases the amount of energy E released in seismic waves is estimated from the earthquake magnitude M through the magnitude-energy relation $\log E = 1.5M + 11.8$ developed by Gutenberg and Richter [*Gutenberg*, 1956*a*]. While this relation was very carefully calibrated through repeated revisions and is considered to give a reasonably accurate estimate of seismic wave energy for most earthquakes, the validity of this relation is questionable for great earthquakes. Here great earthquakes are those with a very large, 100 km or greater, rupture length. This arises from the fact that for such a great earthquake the magnitude M which is determined at the period of 20 s (or converted from m (body wave magnitude) determined at shorter periods) does not represent the entire rupture process of an earthquake. In fact, there is little correlation between M and the rupture length for great earthquakes. Thus the energy E estimated from M is very uncertain for great earthquakes. Yet it is such great earthquakes that contribute most to the seismic energy budget. In order to circumvent this difficulty we estimate in this paper the energy involved in great earthquakes on the basis of static source parameters such as the seismic moment and the area of the fault plane. Since the absolute level of stress involved in faulting is unknown, it is not possible to determine the change in the strain energy before and after an earthquake. However, it is possible to estimate the minimum strain energy drop which, under reasonable conditions, approximates the seismic wave energy. Since the static source parameters are very accurately determined for many great earthquakes, this method gives accurate estimates of energy for great earthquakes, which have the greatest contribution to the seismic energy budget. It is hoped that this method provides a more meaningful basis for various studies pertaining to global processes such as heat flow, Chandler wobble, and plate motions.

COMPILATION OF SEISMIC MOMENTS OF GREAT EARTHQUAKES

The seismic moment M_0, which is defined by $\mu\bar{D}S$ (μ is the rigidity; $\bar{D}$ is the average offset on the fault; and S is the area of the fault), is one of the most accurately determined seismic source parameters. For many great earthquakes, M_0 has been determined by using long-period body waves, surface waves, free oscillations, and geodetic data. A partial list is found in the work by *Kanamori and Anderson* [1975*b*]. For earthquakes for which no direct determination of M_0 has been made, we estimate it from the area of the fault plane S and/or the 100-s magnitude determined by *Brune and Engen* [1969].

A remarkable linearity between $\log M_0$ and $\log S$ has been noted by *Aki* [1972], *Thatcher and Hanks* [1973], *Kanamori and*

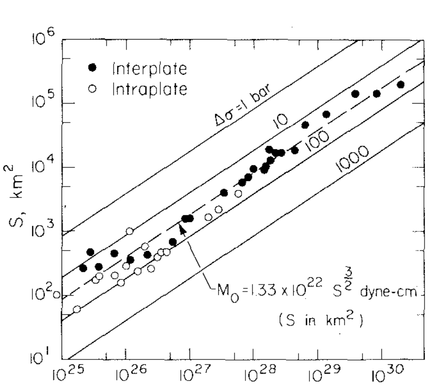

Fig. 1. The relation between the fault area and the seismic moment (modified from *Kanamori and Anderson* [1975*b*]). The dashed line gives the average M_0 versus S relation suggested by *Abe* [1975*a*].

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TABLE 1. Great Earthquakes

Date	Region	M_s	M_0, 10^{27} dyn cm	M_w	Source for M_0 Value*
June 25, 1904	Kamchatka	8.0			
June 25, 1904	Kamchatka	8.1			
April 4, 1905	East Kashmir	8.0			
July 9, 1905	Mongolia	$8\frac{1}{4}$	50	8.4	*Okal* [1977].
July 23, 1905	Mongolia	$8\frac{1}{4}$	50	8.4	*Okal* [1977].
Jan. 31, 1906	Ecuador	8.6	204	8.8	From the aftershock area.
April 18, 1906	San Francisco	$8\frac{1}{4}$	10	7.9	Estimated from fault length of 500 km, width of 15 km, and dislocation of 5 m.
Aug. 17, 1906	Rat Islands	8.0			
Aug. 17, 1906	Central Chile	8.4	29	8.2	From the aftershock area.
Sept. 14, 1906	New Britain	8.1			
April 15, 1907	Mexico	8.1			
Oct. 21, 1907	Afghanistan	8.0			
Jan. 3, 1911	Turkestan	8.4	4.9	7.7	*Chen and Molnar* [1977].
May 23, 1912	Burma	8.0			
May 1, 1917	Kermadec	8.0			
June 26, 1917	Samoa	8.3			
Aug. 15, 1918	Mindanao	$8\frac{1}{4}$			
Sept. 7, 1918	Kurile	$8\frac{1}{4}$			
April 30, 1919	Tonga	8.3			
June 5, 1920	Taiwan	8			
Sept. 20, 1920	Loyalty Islands	8			
Dec. 16, 1920	Kansu, China	8.5	6.6	7.8	*Chen and Molnar* [1977].
Nov. 11, 1922	Central Chile	8.3	69	8.5	From the aftershock area.
Feb. 3, 1923	Kamchatka	8.3	37	8.3	From the aftershock area.
Sept. 1, 1923	Kanto	8.2	8.5	7.9	
April 14, 1924	Philippine	8.3			
May 22, 1927	Tsinghai, China	8.0	3.0	7.6	*Chen and Molnar* [1977].
June 17, 1928	Guerrero, Mexico	7.8	12	8.0	From the aftershock area.
Dec. 1, 1928	Central Chile	8.0	3	7.6	From the aftershock area.
March 7, 1929	Fox Islands (Aleutian)	8.1	6.7	7.8	*Kanamori* [1972*b*].
Aug. 10, 1931	Sinkiang, China	8.0	12	8.0	*Chen and Molnar* [1977].
May 14, 1932	Molucca	8.0	1.5	7.4	From the 100-s magnitude.
June 3, 1932	Jalisco, Mexico	8.1	15	8.1	Average of value from the aftershock area and value from the 100-s magnitude.
March 2, 1933	Sanriku	8.5	43	8.4	
Jan. 15, 1934	India-Nepal	8.3	16	8.1	*Chen and Molnar* [1977].
July 18, 1934	Santa Cruz Islands	8.2	0.8	7.2	From the 100-s magnitude based on one station.
Feb. 1, 1938	Banda Sea	8.2	70	8.5	From the 100-s magnitude.
Nov. 10, 1938	Alaska	8.3	28	8.2	Average of value from the aftershock area and value from the 100-s magnitude.
April 30, 1939	Solomon Islands	8.0			
Dec. 26, 1939	Turkey	8.0			
May 24, 1940	Peru	8.0	25	8.2	From the aftershock area.
June 26, 1941	Andaman Islands	8.1	3	7.6	From the 100-s magnitude.
Nov. 25, 1941	North Atlantic	8.3			
Aug. 24, 1942	Peru	8.1	27	8.2	From the aftershock area.
April 6, 1943	Chile	7.9	28	8.2	From the aftershock area.
Dec. 7, 1944	Tonankai	8.0	15	8.1	
Nov. 27, 1945	West Pakistan	$8\frac{1}{4}$			
Aug. 4, 1946	Dominican Republic	8.1			
Dec. 20, 1946	Nankaido	8.2	15	8.1	
Jan. 24, 1948	Philippine	8.2			
Aug. 22, 1949	Alaska	8.1	15	8.1	From the aftershock area.
Aug. 15, 1950	Assam	8.6	100	8.6	Average of values from *Ben-Menahem et al.* [1974], *Chen and Molnar* [1977], and G. S. Stewart (personal communication, 1977).
Nov. 18, 1951	Tibet	8.0	1.9	7.5	*Chen and Molnar* [1977].
March 4, 1952	Tokachi-oki	8.3	17	8.1	
Nov. 4, 1952	Kamchatka	$8\frac{1}{4}$	350	9.0	*Kanamori* [1976*b*].
March 9, 1957	Aleutian Islands	$8\frac{1}{4}$	585	9.1	From the aftershock area.
Dec. 4, 1957	Mongolia	8.3	18	8.1	*Okal* [1976].
July 10, 1958	Alaska	7.9	29	8.2	From the aftershock area.
Nov. 6, 1958	Kurile Islands	8.7	40	8.3	Y. Fukao (personal communication, 1977).
May 4, 1959	Kamchatka	$8\frac{1}{4}$	26	8.2	From the aftershock area.
May 22, 1960	Chile	8.3	2000	9.5	
Oct. 13, 1963	Kurile Islands	8.1	67	8.5	
March 28, 1964	Alaska	8.4	820	9.2	
Feb. 4, 1965	Aleutian Islands	$7\frac{3}{4}$	125	8.7	
Oct. 17, 1966	Peru	7.5	20	8.1	
May 16, 1968	Tokachi-oki	7.9	28	8.2	
Feb. 28, 1969	North Atlantic	8.0	6	7.8	
Aug. 11, 1969	Kurile Islands	7.8	22	8.2	

TABLE 1. (continued)

Date	Region	M_s	M_0, 10^{27} dyn cm	M_w	Source for M_0 Value*
May 31, 1970	Peru	7.8	10	7.9	
Jan. 10, 1971	West New Guinea	8.1			
Oct. 3, 1974	Peru	7.6	15	8.1	G. S. Stewart (personal communication, 1977).
May 26, 1975	North Atlantic	7.9	5	7.7	*Hadley and Kanamori* [1975].
July 27, 1976	China	8.0	2	7.5	Revised from *Stewart et al.* [1976].
Aug. 16, 1976	Mindanao	8.2	19	8.1	G. S. Stewart (personal communication, 1977).

The values of M_0 not referenced are taken from Table 1 of *Kanamori and Anderson* [1975*b*].

Anderson [1975*b*], *Abe* [1975*a*], and *Geller* [1976]. This linearity is interpreted in terms of constant average stress drop in earthquakes [*Chinnery*, 1964]. Figure 1 demonstrates this linearity for large and great earthquakes. *Abe* [1975*a*] and *Geller and Kanamori* [1977] suggest a relation

$$M_0 = 1.23 \times 10^{22} S^{3/2} \quad \text{dyn cm}$$

where S is in square kilometers, to represent the overall relation between S and M_0. In many cases the aftershock area defined at a relatively early stage of the aftershock sequence, usually 1 day after the main shock, is used for S. This procedure involves some ambiguity but is adequate for the present purpose. *Utsu and Seki* [1954], *Fedotov* [1965], *Mogi* [1968*a*, *b*], *Sykes* [1971], *Kelleher* [1972], and *Kelleher et al.* [1973] mapped aftershock areas and rupture zones of many large and great earthquakes, including those for which no direct determination of M_0 has been made. We estimate M_0 of these earthquakes by using (1) and the size of the rupture zones determined by these authors. Although not very essential, one adjustment is made. The rupture zones determined by these authors are based on the aftershock area at a relatively later stage, usually several months, after the main shock, while S used in (1) is determined from the aftershock area at a relatively early stage, usually 1 day. Comparison between these two sets of data suggests that the former is, on the average, 75% larger than the latter. Therefore in using (1) we divided the size of the published rupture zones by 1.75. The results of moment determinations by this method are listed in Table 1.

Brune and Engen [1969] determined 100-s magnitude M_{100} for 21 great earthquakes. Since M_{100} is determined from the spectral amplitude of 100-s mantle surface waves, it can be used to estimate M_0 if the corner period is shorter than 100 s. In fact, there is a very good correlation between M_{100} and log M_0. For 7 out of the 21 events of *Brune and Engen* [1969], direct determination of M_0 is available. Comparison of M_{100} and log M_0 for these events leads to a relation

$$\log M_0 = 2.83 M_{100} + 4.83 \tag{2}$$

where M_0 is in dyne centimeters. This relation is used to estimate M_0 for the remaining 14 events. The results are listed in Table 1. Table 1 includes all shallow earthquakes of $M_s \geq 8.0$ since 1904 (when the magnitude refers specifically to the 20-s surface wave magnitude, it is denoted by M_s). These earthquakes are taken from *Gutenberg and Richter* [1954] for the period from 1904 to 1952, from the *Science Almanac* [*Tokyo Astronomical Observatory*, 1975, 1977] for the period from 1953 to 1975 and from the Preliminary Determination of Epicenters (PDE) cards of the U.S. Geological Survey for 1976. Nine earthquakes of $M_s < 8.0$ for which M_0 is known are included.

For the period from 1921 to 1976 the data are fairly complete; there are eight earthquakes for which M_0 is unknown, but only four of them have M_s larger than 8.1. It is notable that in terms of M_0, four earthquakes, the 1960 Chilean, 1964 Alaskan, 1957 Aleutian Islands, and 1952 Kamchatka earthquakes, dominate. For the period prior to 1920, Table 1 is very incomplete, except around 1905 and 1906.

MOMENT M_0, MINIMUM STRAIN ENERGY DROP W_0, AND A NEW MAGNITUDE SCALE M_W

The seismic moment M_0 is a very important earthquake parameter that measures the overall deformation at the source. In particular, it has a very important bearing on global phenomena such as plate motion [*Brune*, 1968; *Davies and Brune*, 1971; *Kanamori*, 1977], polar motion, and rotation of the earth [*Smylie and Mansinha*, 1968; *Dahlen*, 1973; *Anderson*, 1974; *Press and Briggs*, 1975; *O'Connell and Dziewonski*, 1976].

The seismic moment can be also interpreted in terms of the strain energy released in earthquakes. In the framework of the elastic stress relaxation model of an earthquake [*Knopoff*, 1958] the difference in the elastic strain energy W before and after an earthquake can be written as

$$W = \bar{\sigma} \bar{D} S \tag{3}$$

where $\bar{\sigma}$ is the average stress during faulting. If the stress drop is complete, the stress drop $\Delta\sigma$ is equal to $2\bar{\sigma}$, and

$$W = W_0 = \tfrac{1}{2} \Delta\sigma \bar{D} S = (\Delta\sigma / 2\mu) M_0 \tag{4}$$

Since $\Delta\sigma$ is nearly constant at 20–60 bars = 2–6 × 10^7 dyn/cm² for very large earthquakes (Figure 1) and μ = 3–6 × 10^{11} dyn/cm² under crust–upper mantle conditions, $(\Delta\sigma/\mu) \sim 10^{-4}$ and (4) becomes

$$W_0 \sim M_0 / (2 \times 10^4) \tag{4'}$$

Thus one can estimate W_0 by dividing the seismic moment by 2×10^4.

When the stress drop is partial, the situation becomes more complicated. We let σ_0 and σ_1 be the initial and final stresses, respectively. Then

$$W = \bar{\sigma} \bar{D} S = (\Delta\sigma/2) \bar{D} S + \sigma_1 \bar{D} S = W_0 + \sigma_1 \bar{D} S \tag{5}$$

Unless a substantial overshoot occurs, σ_1 is usually positive, so that W_0 gives the minimum estimate of the strain energy drop. We can attach more significance to W_0 if we introduce a model proposed by *Orowan* [1960]. We let σ_f be the frictional stress during faulting. Then

$$W = H + E$$

TABLE 2. Earthquakes of Large M_w

Event	Year	M_w
Chile	1960	9.5
Alaska	1964	9.2
Aleutian	1957	9.1
Kamchatka	1952	9.0
Ecuador	1906	8.8
Aleutian	1965	8.7
Assam	1950	8.6
Kurile Islands	1963	8.5
Chile	1922	8.5
Banda Sea	1938	8.5
Mongolia	1905	8.4
Mongolia	1905	8.4
Sanriku	1933	8.4
Kamchatka	1923	8.3
Kurile Islands	1958	8.3
Chile	1906	8.2
Alaska	1938	8.2
Kamchatka	1959	8.2
Tokachi-oki	1968	8.2
Peru	1940	8.2
Peru	1942	8.2
Alaska	1958	8.2
Chile	1943	8.2
Kurile	1969	8.2
Mexico	1932	8.1
Tonankai	1944	8.1
Nankaido	1946	8.1
Alaska	1949	8.1
Tokachi-oki	1952	8.1
Mongolia	1957	8.1
Peru	1966	8.1
India-Nepal	1934	8.1
Peru	1974	8.1
Mindanao	1976	8.1
Mexico	1928	8.0
China	1931	8.0
San Francisco	1906	7.9
Kanto	1923	7.9
Peru	1970	7.9

where $H = \sigma_f \bar{D}S$ is the frictional loss and E is the wave energy. Using (3), we have

$$E = \bar{\sigma}\bar{D}S - \sigma_f\bar{D}S = (\Delta\sigma/2)\bar{D}S + \bar{D}S(\sigma_1 - \sigma_f) = W_0 + \bar{D}S(\sigma_1 - \sigma_f) \quad (6)$$

Thus if *Orowan*'s [1960] condition $\sigma_1 = \sigma_f$ is met, W_0 is not only the minimum estimate of W but also is equal to the wave energy [see also *Savage and Wood*, 1971].

Whether the earthquake stress drop is complete or partial is presently unresolved. *Brune et al.* [1969] argued, on the basis of lack of heat flow anomaly along the San Andreas fault, that frictional stress is very small. In this case the stress drop is nearly complete, and W_0 represents the actual strain energy drop. On the other hand, evidence for a very high (~1 kbar) tectonic stress has been suggested primarily from the analysis of the deformation of the oceanic lithosphere [*Hanks*, 1971; *Watts and Talwani*, 1974; *Caldwell et al.*, 1976]. If this high stress is representative of the tectonic stress that causes earthquakes, then the stress drop may be partial. Although this problem remains unresolved, W_0 is still a useful parameter in that it gives the minimum strain energy drop in earthquakes.

Furthermore, results of *Trifunac* [1972], *Kanamori* [1972*a*], *Abe* [1975*b*], *Kanamori and Anderson* [1975*b*], and *Geller* [1976] suggest that the stress drop is approximately equal to the effective stress; i.e., *Orowan*'s [1960] condition $\sigma_1 = \sigma_f$ is satisfied. Then (6) means that W_0 determined by (4′) is equal to the wave energy E.

For a more conventional measure of the 'size' of great earthquakes it is convenient to use a magnitude scale. To this end, we define a new magnitude scale for great earthquakes in terms of W_0 by using the Gutenberg-Richter magnitude-energy relation, $\log E = 1.5M + 11.8$. We use W_0 calculated from M_0 for E in this equation, calculate M, and denote it by M_w. The results are listed in Table 1. Table 2 lists the 39 largest earthquakes on this scale. The 1960 Chilean earthquake has the largest M_w, 9.5. The 1964 Alaskan ($M_w = 9.2$), 1957 Aleutian Islands ($M_w = 9.1$), and 1952 Kamchatka ($M_w = 9.0$) earthquakes follow. It is interesting to note that M_w agrees very well with M_s for many earthquakes with a rupture length of about 100 km (e.g., 1944 Tonankai, 1946 Nankaido, 1952 Tokachi-oki, 1966 Peru, 1923 Kanto, and 1970 Peru). This agreement may suggest that the Gutenberg-Richter magnitude-energy relation, $\log E = 1.5M + 11.8$, gives the correct value of seismic wave energy for earthquakes up to this size, i.e., a rupture dimension of $\lesssim$100 km. Thus the M_w scale can be used as a natural continuation of the M_s scale for great earthquakes. The saturation of the M_s scale for great earthquakes [*Kanamori and Anderson*, 1975*b*; *Geller*, 1976; *Chinnery and North*, 1975] has been an inconvenient and sometimes a confusing element in the conventional magnitude scale. The use of M_w eliminates this saturation.

TEMPORAL VARIATION OF ENERGY RELEASE IN EARTHQUAKES

As shown in the previous section, $W_0 = (\Delta\sigma/2\mu)M_0$ represents the minimum strain energy drop in an earthquake, and under the condition $\sigma_1 = \sigma_f$ (i.e., *Orowan*'s [1960] condition, or the condition that effective stress equal stress drop) it is equal to the seismic wave energy. The condition $\sigma_1 \sim \sigma_f$ has been verified experimentally for several earthquakes.

Figure 2 shows W_0 for great earthquakes as a function of year plotted from Table 1. The solid curve shows the annual average of W_0 obtained by taking a 5-year running average (taken at the center of the interval) of the data in Table 1. In the computation of the annual release curve, earthquakes for which the seismic moment M_0 is not known are inevitably

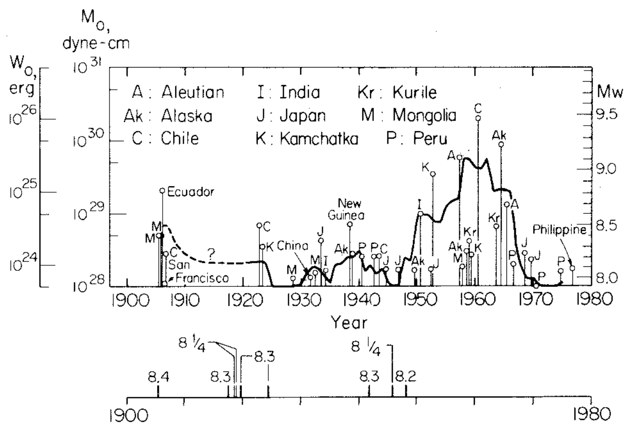

Fig. 2. The minimum strain energy drop W_0 (equal to the seismic wave energy if *Orowan*'s [1960] condition is met) in great earthquakes as a function of year. The solid curve shows unlagged 5-year running average (in ergs per year) taken at the center of the interval. The ordinate is given in three scales, the seismic moment M_0, W_0, and M_w. Large earthquakes for which M_0 has not been determined are plotted at the bottom with the surface wave magnitude M_s.

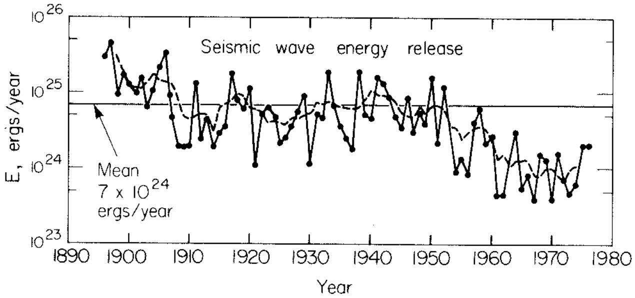

Fig. 3. Seismic wave energy released in earthquakes computed from the surface wave magnitude M_s through the Gutenberg-Richter energy versus magnitude relation. The dashed curve shows the unlagged 5-year running average.

ignored. However, since 1921, only four events of $M_s \geq 8.2$ are missing, and it is unlikely that the omission of these events affects the energy release curve drastically. For the period prior to 1920, Table 1 is very incomplete except around 1906. The annual average of W_0 for the period from 1920 to 1976 is 4.5×20^{24} ergs/yr.

It is remarkable that during the 15-year period from 1950 to 1965 the annual average of W_0 is more than an order of magnitude larger than that during the periods from 1920 to 1950 and from 1965 to 1976. Another peak is suggested around the turn of the century, but its confirmation must await further studies.

As mentioned earlier, W_0 represents the minimum strain energy drop, and the actual strain energy drop can be larger than this, if the stress drop in great earthquakes is only partial. Even then, if the fractional stress drop is about the same for all earthquakes, Figure 2 still gives the correct trend of the relative strain energy release.

Correlation Between W_0, Gutenberg-Richter Energy, Number of Events, and Polar Motion of the Earth

It is instructive to compare the temporal variation of W_0 with the conventional energy release curve computed from the magnitude. *Gutenberg* [1956*b*] calculated the annual energy release for the period from 1896 to 1955 by using the earthquake magnitude and the energy versus magnitude relation log $E = 1.5M + 11.8$. We extended this calculation to 1975 by using the catalog of earthquakes listed in the *Science Almanac* [*Tokyo Astronomical Observatory*, 1975, 1977] and to 1976 by using the PDE cards of the U.S. Geological Survey and Caltech determinations. The energy E calculated by this method, here called the Gutenberg-Richter energy, refers to the seismic wave energy radiated by earthquakes. As was discussed earlier, however, because of the saturation of the ordinary magnitude scale this relation tends to underestimate the wave energy of great earthquakes. Thus the annual energy curve computed by this method can be considered to approximate the wave energy radiated by earthquakes of up to moderate to large size. Figure 3 shows the variation of E as a function of year.

Another measure of seismic activity is the number of earthquakes. Figure 4 shows the annual number N of earthquakes of $M_s \geq 7.0$ taken from the catalog of the *Science Almanac* [*Tokyo Astronomical Observatory*, 1975, 1977]. Since 96.7% of these earthquakes have M_s between 7 and 8, the temporal variation of N is more representative of the activity of moderate to large earthquakes.

Since the estimate of the Gutenberg-Richter energy E based on the magnitude-energy relation can be greatly affected by errors in the magnitude of the few larger earthquakes, the number of events N is more representative of the global activity of moderate to large earthquakes than E.

Despite the large uncertainty in E (Figure 3) the general trends of the curves of E and N are very similar to each other. In particular, both E and N show a very steady decrease since the middle 1940's. It is quite remarkable that during this period there was a very pronounced increase in W_0. The correlation is shown in Figure 5. Although the energy release curve itself may be subject to considerable uncertainty, it is certain that the number of earthquakes of $M_s \geq 7.0$ decreased very sharply during the period when many great earthquakes with a very large rupture dimension (500–1000 km) occurred from 1952 to 1965. This complementary occurrence of great earthquakes and moderate to large earthquakes is a very intriguing feature, suggestive of a causal relationship between these two groups of earthquakes.

In Figure 5 is also plotted the temporal variation of the amplitude (envelope) of the Chandler wobble taken from *Anderson* [1974] (for the period from 1900 to 1960) and *O'Connell and Dziewonski* [1976] (for the period from 1960 to 1970). The variation of the wobble shows a trend very similar to that of W_0 for the period from 1920 to 1970. A peak in the wobble curve around 1910 may be correlatable to a peak in W_0 suggested around the turn of the century. Although the data presented in this paper are not complete for this period, it is notable that many large earthquakes occurred all over the world around the turn of the century, e.g., Alaska, Tibet, the

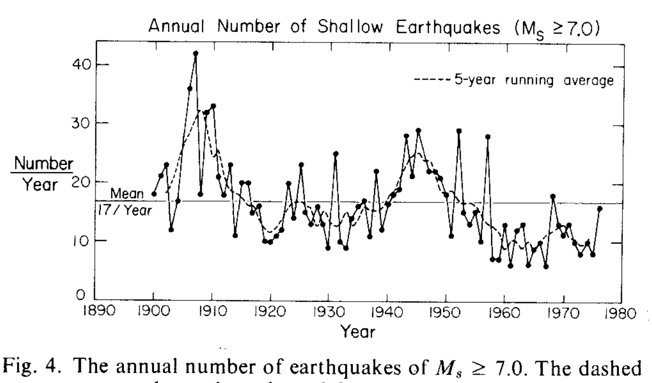

Fig. 4. The annual number of earthquakes of $M_s \geq 7.0$. The dashed curve shows the unlagged 5-year running average.

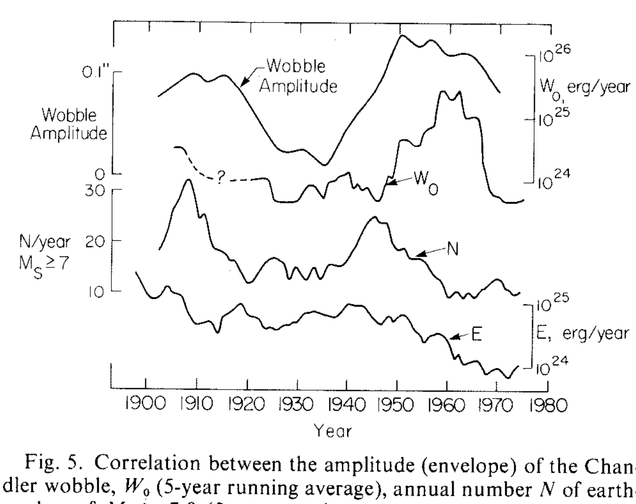

Fig. 5. Correlation between the amplitude (envelope) of the Chandler wobble, W_0 (5-year running average), annual number N of earthquakes of $M_s \geq 7.0$ (5-year running average), and the Gutenberg-Richter energy E (5-year running average).

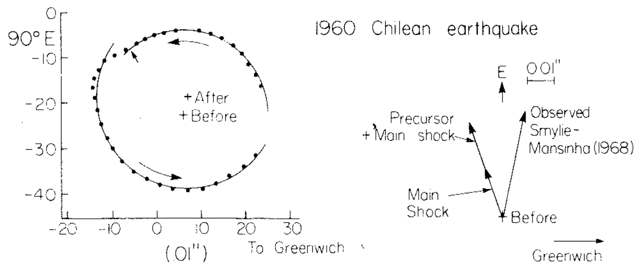

Fig. 6. The polar motion before and after the 1960 Chilean earthquake (left) inferred by *Smylie and Mansinha* [1968]. The center of the polar motion before and after the earthquake is shown by a plus sign. Comparison of the observed and computed polar shift is shown on the right. The computation is made by using *Dahlen*'s [1973] expression for the source parameters determined by *Kanamori and Cipar* [1974].

Philippines, Mexico, New Zealand, Santa Cruz Island, Russia, the Caribbean, Loyalty Island, Guatemala, and Java.

Discussion and Conclusion

As shown in the previous section, W_0 represents the energy release in great earthquakes, while E or N represents that in moderate to large earthquakes. Therefore if there is a causal relation between the wobble and earthquake activity at all, it is more reasonable to compare the wobble with W_0 than with N or E.

Anderson [1974] discussed several possible mechanisms that would explain such a correlation. The first possibility is that the deformation caused by a great earthquake excites the Chandler wobble. The second is that a change in the polar motion caused by other factors, such as atmospheric changes, affects the plate motion, thereby triggering great earthquakes and other major earthquakes. Combination of these two mechanisms is also possible. Regarding the first possibility, many investigations have been made, those by *Smylie and Mansinha* [1968], *Dahlen* [1973], *Israel et al.* [1973], *Press and Briggs* [1975], and *O'Connell and Dziewonski* [1976] to mention a few. One problem is that the deformation caused by even a great earthquake is not large enough to excite the Chandler wobble unless a large aseismic slip is assumed [*Dahlen*, 1973; *O'Connell and Dziewonski*, 1976; *Kanamori*, 1976*a*]. Only the 1960 Chilean earthquake, the largest of all in M_w, can account for the shift of the pole position when the preseismic anelastic deformation reported by *Kanamori and Cipar* [1974] and *Kanamori and Anderson* [1975*a*] is included (Figure 6). Existence of large aseismic deformation has been suggested for very large tsunami earthquakes such as the 1896 Sanriku earthquake and the 1946 Aleutian Islands earthquake [*Kanamori*, [1972*b*], for the 1906 San Francisco earthquake [*Thatcher*, 1974], for the 1952 Kamchatka earthquake [*Kanamori*, 1976*b*], and for a Japanese earthquake [*Fukao and Furumoto*, 1975]. Also, disparity between seismic slip and plate motion provides evidence for such aseismic deformation [*Kanamori*, 1977]. Thus the first possibility still remains valid.

The second possibility is very intriguing. Recent analysis of *Wilson* [1975] suggests that atmospheric motions can maintain the Chandler wobble. In this context, *Anderson* [1975] notes that the temporal variation of global temperatures, one climatic indicator, is very similar to that of the wobble. It is quite possible that the increase in the amplitude of the Chandler wobble caused by such effects accelerates global plate motions, thereby triggering great earthquakes at plate boundaries. Figure 5 indicates that the sharp increase in W_0 around 1960 began very shortly after the amplitude of the wobble became maximum in 1950. This coincidence may be suggestive of the second possibility. It is remarkable that the annual number of earthquakes N increased toward 1945 and then decreased very sharply since then. One possibility is that when the wobble amplitude increases, the world seismic activity increases, and plate motion may be accelerated. However, once major plate boundaries are decoupled in great earthquakes, moderate to large earthquake activity declines owing to decrease in intraplate and interplate stresses as a result of plate decoupling.

It is equally possible that changes in the rotation rate of the earth are responsible for accelerated plate motions which in turn cause the variation in the Chandler wobble and great earthquakes. The change in the rotation rate of the earth correlates very well with the Chandler wobble [*Anderson*, 1974]. Since the rotational energy of the earth is so much greater than the energy involved in plate motions and earthquakes, even a small perturbation in the rotation can have a significant effect on earthquakes and plate motion.

The conclusions are as follows: (1) The minimum estimate of the strain energy drop in earthquakes, W_0, which can be estimated from the seismic moment M_0 can be considered to represent, under *Orowan*'s [1960] condition, the seismic wave energy release. (2) Since W_0 can be estimated accurately for great earthquakes, it provides a more accurate picture of the seismic energy budget. (3) A new magnitude scale M_w is defined in terms of W_0. It is as large as 9.5 for the 1960 Chilean earthquake and connects smoothly to M_s for moderate to large earthquakes. Therefore M_w provides a convenient magnitude scale which does not saturate. (4) The temporal variation of W_0, the energy release in great earthquakes, is very different from that in moderate to large earthquakes. The activity of moderate to large earthquakes was very low when W_0 was largest during the period from 1950 to 1965. (5) The amplitude of the Chandler wobble seems to correlate very well with W_0, with a slight indication of the former preceding the latter. (6) One possible mechanism that accounts for the correlation between the wobble, W_0, and the activity of moderate to large earthquakes is that an increase in wobble amplitude triggers worldwide seismic activity and accelerates plate motion, which eventually leads to great decoupling earthquakes. This decoupling causes the decline of moderate to large earthquake activity.

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