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# Use of the Mantle Magnitude $M_m$ for the Reassessment of the Moment of Historical Earthquakes.

## I: Shallow Events

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*Abstract*—The mantle magnitude $M_m$ is used on a dataset of more than 180 wavetrains from 44 large shallow historical earthquakes to reassess their moments, which in many cases had been previously estimated only on the basis of the earthquake's rupture area. We provide 27 new or revised values of $M_0$, based on the spectral amplitudes of surface waves recorded at a number of stations, principally Uppsala and Pasadena. Among them, and most significantly, we document a large low-frequency component to the source of the 1923 Kanto earthquake: the low-frequency seismic moment is $2.9 \times 10^{28}$ dyn-cm, in accord with geodetic observations. On the other hand, we revise downwards the seismic moment of the 1906 Ecuador event, which did not exceed $6 \times 10^{28}$ dyn-cm.

Finally, the study of the 1960 Chilean and 1964 Alaskan earthquakes whose exceptionally large moments are properly retrieved through $M_m$ measurements, serves proof that this approach performs flawlessly even for the very greatest earthquakes, and is therefore successful in its goal to avoid the saturation effects plaguing any magnitude scale measured at a fixed period.

**Key words:** Historical earthquakes, magnitudes, mantle waves, tsunamis.

### *Introduction and Purpose*

The purpose of this paper is to use the variable-period mantle magnitude $M_m$ in order to reassess the seismic moment of a number of large historical earthquakes. The concept of $M_m$, introduced initially for Rayleigh waves and later extended to Love waves (OKAL and TALANDIER, 1989, 1990; hereafter Papers I and II), is an attempt to define a magnitude scale firmly related to the seismic moment $M_0$ (and in particular avoiding the well-known saturation effects suffered by $M_s$ and other scales defined at constant periods), while at the same time retaining the basic philosophy of a magnitude scale, i.e., a quick, one-station measurement that does not require the knowledge of either the earthquake's focal geometry, or its exact depth.

In Paper I, we showed that the mantle magnitude $M_m$ could be defined as

$$M_m = \log_{10} X(\omega) + C_S + C_D - 0.90 \quad (1)$$

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where $X(\omega)$ is the spectral amplitude of a Rayleigh wave in $\mu$m-s, and $C_S$ a source correction described by the cubic spline

$$C_S = 1.61630\theta^3 - 0.833220\theta^2 + 0.428610\theta + 3.7411 \tag{2}$$

where $\theta = \log_{10} T - 1.8209$. $C_D$ is a frequency-dependent distance correction:

$$C_D = 0.5 \log_{10} \sin \Delta + (\log_{10} e) \cdot \frac{\omega a \Delta}{2UQ} \tag{3}$$

and the constant 0.90 predicts theoretically that $M_m$ should approach the quantity $\log_{10} M_0 - 20$. The second part of the correction $C_D$ can be regionalized to take into account the lateral variation of attenuation and dispersion with tectonic regime. In these equations, $T = 2\pi/\omega$ is the period of the wave in seconds, and $\Delta$ the distance traveled in radians; $a$ is the earth's radius in km, and $U$ the group velocity in km/s. $Q$ is the quality factor of the wave at period $T$. In the case of Love waves, Equation (1) still holds, but the correction $C_S$ must be adjusted to reflect the different excitation of Love waves:

$$C_S^{\text{Love}} = 0.802630\theta^3 + 0.135240\theta^2 + 0.285700\theta + 3.6122 \tag{4}$$

where $\theta$ is now $\log_{10} T - 2.2354$. Similary, the parameters $U$ and $Q$ in the distance correction (3) must be given values appropriate to Love waves.

On the basis of a large dataset of more than 500 records, we showed in Papers I and II that $M_m$ gave an estimate of its theoretical value ($\log_{10} M_0 - 20$) with average residuals $\bar{r}$ on the order of 0.15 unit of magnitude, and standard deviations on the order of 0.25 unit of magnitude, in the range $M_m = 5.9 - 8.3$. More recently, REYMOND *et al.* (1991) have shown that the concept can be extended without difficulty nor degradation of performance to moments as low as $10^{24}$ dyn-cm.

The mantle magnitude $M_m$ was developed as a means of extracting reliable moment information in real time, and therefore from the limited dataset available at a single station. Because of the occasional loss of seismograms over the years, or of uncertainty about instrument responses, the situation with historical earthquakes can be very similar, and thus $M_m$ should be a powerful tool in estimating their moments.

### *Historical Earthquakes*

The need to re-interpret historical data and in particular earthquake sources, in the framework of modern developments in seismological theory stems from several reasons.

The evaluation of seismic risk along individual subduction zones surrounding the Pacific Basin involves concepts such as "recurrence times" and "seismic gaps" (e.g., NISHENKO, 1991), which rely heavily on a precise understanding of historical earthquakes. In decoupled subduction zones, such as the Marianas, there is ample evidence that earthquakes do not reach the catastrophic levels they can attain in a more coupled environment (UYEDA and KANAMORI, 1979; RUFF and KANAMORI, 1980). However, the seismological classification of the degree of coupling of a given segment of subduction zone relies, to a large extent, on the maximum earthquake size reported at the site. In some instances, notably for the Samoa-Tonga-Kermadec region, no reliable moments are available, and the assessment of seismic risk is made difficult. In other areas, known to undergo major earthquakes, the estimation of a recurrence time runs into the problem of comparing a modern (well studied and reliably quantified) event with the previous one, often a historical earthquake, for which only a conventional magnitude is available. Estimating the moment of the historical earthquake is crucial to a correct inference of recurrence time.

Traditionally, the year 1963, when the WWSSN was implemented, has been used to separate "modern" from "historical" earthquakes; however, with further progress in theory and instrumentation, we can distinguish the following periods:

- For events postdating 1976, more than 9500 Centroid-Moment-Tensor solutions (CMT) are available from the Harvard project (DZIEWONSKI *et al.*, 1987a and subsequent quarterly updates).
- For events between 1963 and 1976, WWSSN data are generally available and abundant, and the moment and geometry (in modern lingo, a CMT solution) can be worked out, the only difficulty being the tedious task of digitizing the analog records. Indeed, most large shallow earthquakes of this period have been studied in detail individually (e.g., KANAMORI, 1970a,b; WU and KANAMORI, 1973). It is worth noting that this is not the case for intermediate and deep earthquakes: while their focal geometries have been extensively studied in the context of stresses inside subduction zones, the seismic moment information was generally not obtained in what were mostly first motion studies (e.g., ISACKS and MOLNAR, 1971), and is available only for a few isolated, and in practice large, events (MENDIGUREN, 1973; GILBERT and DZIEWONSKI, 1975; FURUMOTO and FUKAO, 1976). The present paper, however, is concerned only with shallow events ($h \leq 75$ km); a companion paper examines the case of intermediate and deep historical earthquakes (OKAL, 1992).
- For earthquakes predating 1963, the availability and quality of seismological data decreases rapidly with increasing age of the event. In general, the large events have been targeted for detailed studies, notably by H. Kanamori and co-workers (e.g., KANAMORI 1976; KANAMORI and CIPAR, 1974). However, for many earthquakes, no such study exists, in some instances because too few seismograms could be recovered.

In this framework, the present paper has several objectives:

*1. Evaluate if $M_m$ works for truly gigantic earthquakes.* Our prime motivation in developing the mantle magnitude $M_m$ was to avoid saturation for very large events.

In the course of the performance evaluation studies in Papers I and II, we showed that the regressed slopes of $M_m$ as a function of the [published] moment of the events were not statistically different from 1, and therefore concluded that we had successfully avoided saturation. Yet, the ultimate test is to compute $M_m$ values for truly gigantic events, say with $M_0 \geq 10^{29}$ dyn-cm. The study in Papers I and II were purposely limited to homogeneous datasets for which both broad-band records and CMT solutions were available. Since no truly great earthquake has happened in the past 27 years[2], it is clearly necessary to involve historical earthquakes to test the performance of $M_m$ at very large moments. In this particular framework, we will use events whose moment is well known from detailed seismological studies, and test $M_m$ against the published (and reliable) value of $M_0$.

*2. Re-assess events whose moment is either unavailable or questionable.* One of the best catalog of moments for very large historical earthquakes remains KANAMORI's (1977) compilation; it is, however, somewhat incomplete, in that a few very large events (notably the 1917–1919 shocks along the Samoa-Kermadec subduction zones) are assigned only a conventional magnitude $M_s$. In addition, we note that no fewer than 15 very large events (usually in the range $M_0 \geq 10^{28}$ dyn-cm) have moments derived from the size of the aftershock area. This procedure, which uses scaling laws relating the length $L$ or the area $S$ of faulting to the seismic moment $M_0$, can run into several problems. First and foremost, the accuracy of earthquake locations degrades significantly when going back in time. Relocations of aftershocks *relative* to the main shock can be even less reliable: for example WYSESSION *et al.* (1991) have shown that the aftershocks of large subduction zone events (e.g., the 03 February 1923 Kamchatka earthquake) could be mislocated by several degrees. Taking into account that $M_0$ grows as $L^3$, this becomes a serious source of error. Furthermore, the assumption of seismic scaling could be wrong when the surface of faulting has a pronounced aspect ratio (as for large strike-slip events on shallow faults), or finally, if the stress drop $\Delta\sigma$ was unusually high—or low.

Other estimates of $M_0$ for historical earthquakes have been inferred from BRUNE and ENGEN's (1969) 100-s magnitude measurements. Conceptually, their study comes closest to the mantle magnitude $M_m$, but retains the measurement of seismic amplitude at a *constant* (though admittedly long) period, and cannot therefore avoid eventual saturation. Finally, ABE (1983) has proposed to obtain a direct estimate of the seismic moment of an earthquake from a tsunami magnitude $M_t$, which he designed to be equivalent to KANAMORI's (1977) $M_w$.

In this framework, we will seek to obtain independent estimates of $M_0$ from the mantle magnitude $M_m$, and to compare them to previously published values.

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[2] As of the date of proof-reading, July 14, 1992.

## *Dataset*

### *Targeting the Stations*

Inherent in the success of $M_m$ is the use of very long mantle periods, around and possibly beyond 200 s. It is therefore necessary to secure records from instruments featuring good response at these periods.

Before 1930, the choice is basically between mechanical and electromagnetic seismographs, typified by, respectively, the Wiechert and Golitsyn instruments (WIECHERT, 1904; GOLITSYN, 1908). A significant network of both types of instruments existed even before World War I. For the study of mantle waves, electromagnetic instruments suffer from the fall-off of their response as $\omega^3$ at very long periods (as compared to $\omega^2$ for mechanical systems). According to MCCOMB and WEST's (1931) compilation, the typical periods used in Golitsyn systems were approximately 12 s, the longest reached 25 s. Figure 1 compares the response of such an instrument (dashed line) to that of a Wiechert with 10 s pendulum period

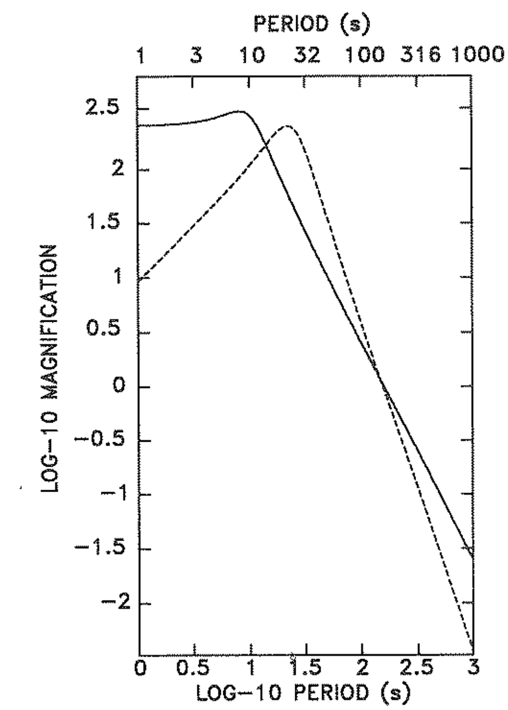


Figure 1

Typical response curves of a mechanical (Wiechert) instrument (solid line), compared to an electromagnetic (Golitsyn) system (dashed line). The Wiechert has the nominal characteristics of the Uppsala instrument; the Golitsyn is the longest-period version ($T_p = T_g = 25$ s). Note the faster fall-off of the Golitsyn, which eventually leads to poorer gains at long periods.

(solid line). It is clear that the advantage of the Golitsyn, while substantial in the range of crustal waves ($T = 20$ s), erodes fast with increasing $T$, and has totally vanished at the periods of mantle waves ($T = 200$ s). For that reason, we did not attempt to use any Golitsyn records in our study.

The Seismological Institute at the University of Uppsala (UPP) possesses one of the most complete and readily accessible collection of Wiechert seismograms. Also, the Uppsala Wiechert has a relatively long pendulum period $T \sim 10$ s (as opposed to 5 s at many other observatories); furthermore, the instruments are still being operated as of this day, allowing immediate comparison between historical events and modern earthquakes of comparable epicenters (KULHÁNEK, 1987). For all these reasons, we decided to use Uppsala records in the present investigation. We were able to obtain a total of 83 usable seismograms.

For reasonably recent (post-1930) earthquakes, we relied heavily on the archives of Caltech's Seismological Laboratory, in Pasadena (PAS). Choice instruments for our purpose included BENIOFF's (1935) strainmeter, whose response is basically equivalent to that of a mechanical instrument with a pendulum period equal to the strainmeter's galvanometer period ($T_g = 70$ s), and also Benioff's "1-90" seismometer, which combines the advantage of a large short-period magnification, and a very long $T_g$ (by historical standards), ensuring adequate gains around 100–200 s. Whenever possible, we also used records on the Press-Ewing instruments developed during the 1950s, and from the ultra-long period seismometers operated since the 1960s (GILMAN, 1960). Finally, the strainmeters and Benioff 1-90 instruments were preceded, during their development, by prototypes, such as a 35-s strainmeter and a Benioff "1-120." We also used these records, whenever we could recover information on their characteristics.

A number of records gathered for individual events on a more or less *ad hoc* basis complete our dataset, for a grand total of 185 wavetrains.

*Instrument Responses*

Paramount in any such study is the ability to reliably recover information on instrument response. The Seismological Bulletin of Uppsala University contains a detailed log of the annual (and often times, more frequent) calibration of the constants of the Wiechert instrument, from the very beginning of its operation (October, 1904) to 1960. After that date, the practice of regularly calibrating the Wiechert was discontinued, and there remains some uncertainty as to its response. We corrected each seismogram with the appropriate constants $T_p$, $V$, $\varepsilon$ for the particular date and component, and in the case of post-1960 records, we used the last available calibrations.

At Pasadena, we relied on existing computer software, and on available station logs and calibrations (H. Kanamori and D. I. Doser, pers. commun., 1990).

*Targeting the Events*

With our dual purpose in mind, we sought to gather as many records as possible of the very largest earthquakes which took place before the development of the WWSSN. A special emphasis was put on the Pacific Basin, since the development of $M_m$ was initially motivated by its application to tsunami warning (TALANDIER and OKAL, 1989). We also retained UPP records for a number of recent large earthquakes, in order to compare these shocks to some of the historical events having occurred at identical locations. In order to make sure that the uncertainty in instrument response after 1960 at UPP does not hinder the computation of $M_m$ and the estimation of $M_0$, we also processed a number of records from recent, large earthquakes (Mexico and Chile, 1985; Macquarie Ridge, 1989) for which accurate CMT solutions are available.

In addition, in a number of decoupled subduction zones (Vanuatu, Philippines), we targeted for study the largest historical earthquakes recorded in terms of conventional magnitude (principally the Pasadena magnitude assigned by GUTENBERG and RICHTER, 1954), in order to verify that these events remain moderate in terms of moment, by the standards of subduction zones.

## *Methods*

At Uppsala, original smoked-paper records were digitized directly, and later processed to remove pen curvature and to equalize time sampling to $\Delta t = 1$ s. At Pasadena, photographic recording allowed xeroxing of the originals for later digitizing and processing.

In most cases, we obtained horizontal records, which in principle should be rotated into radial and transverse polarization before further use. In some instances, however, only one of the records was available (or usable), in which case an adequate correction was effected. For standard seismometers, the correction is simply $-\log_{10}|\cos\beta|$ for Rayleigh waves, and $-\log_{10}|\sin\beta|$ for Love waves, where $\beta$ is the back-azimuth of the epicenter as seen from the station, relative to the direction of the particular component of ground motion recorded. For strainmeters, the correction is $-\log_{10}|\cos^2\beta|$ (Rayleigh) or $-\log_{10}|\cos\beta\sin\beta|$ (Love). An ellipticity correction was also included when using the horizontal component of Rayleigh waves.

The standard $M_m$ algorithm consists of Fourier-transforming the seismogram, computing expression (1) at each FFT period in the range of mantle waves, and keeping the largest such number. In this respect, the definition of the allowable frequency range is crucial: if it is chosen to be too narrow, the resulting $M_m$ could be underestimating the moment due to source finiteness effects; ultra-long periods, on the other hand, carry the danger of processing noise (either of telluric or digitizing nature) in a frequency band where the instrument response has fallen dramatically.

To guard against this possibility, we define "digitizing noise" as a signal of amplitude $n(t) = 0.25$ mm on the original records, and consisting of uncorrelated single sinusoidal arches, each being of a period significantly different from the previous one. As discussed by OKAL and TALANDIER (1987), under these conditions, the relationship between time-domain and spectral amplitudes is $N(\omega) = n(t)T/2$. This allows us to define a threshold of spectral amplitude below which measurements are contaminated by noise. Immediately after Fourier-transforming the record, and before deconvolving the instrument response, the spectral amplitude $X(\omega)$ is compared to $N(\omega)$; if it is smaller, that frequency is automatically rejected in the $M_m$ algorithm. Although such exclusions seldom took place, they provide some control mechanism against falling below noise levels. When interpreting $M_m$ values, we further consider that a measurement is reliable only if the spectral amplitude retained is at least twice as large as $N(\omega)$. In practice, and given a specific instrument and distance, this defines a minimum reliable $M_m$ at each period, as shown on Figure 2, in the case of a typical horizontal Wiechert record at Uppsala

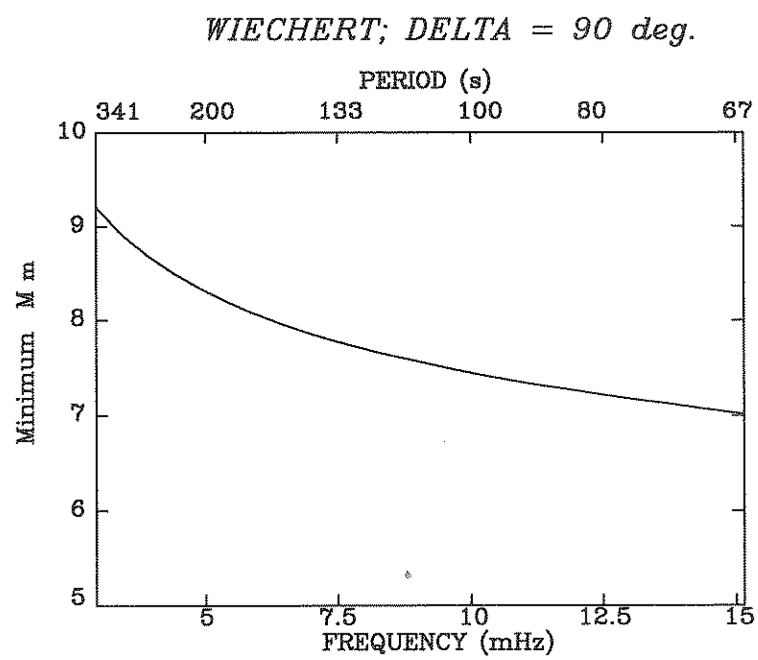


Figure 2
Minimum mantle magnitude $M_m$ measurable as a function of frequency on a horizontal component of the UPP Wiechert, at a typical distance of 90°. The solid line represents the magnitude equivalent to a noise level of 0.5 mm on the record.

($\Delta = 90°$; $T = 10$ s; $V = 200$; $\varepsilon = 4$). Any value of $M_m$ falling on or below the curve on Figure 2 can be considered noise.

Finally, an interesting rule of thumb applies to the Uppsala records: it takes a moment of approximately $10^{29}$ dyn-cm to produce visible multiple wavetrains $R_2$ or $G_2$. This empirical observation, obtained from many well-studied events can be justified by considering the formula for time-domain measurements of Rayleigh wave $M_m$;

$$M_m^{TD} = \log_{10}[a \cdot T] + C_S + C_D + 0.5 \log_{10} \frac{\Delta}{70°} - 1.20 \tag{5}$$

appropriate for distances larger than 120° (OKAL, 1989). An absolute minimum for the *visual* identification of $R_2$ can be taken as $a = 1$ mm. At the periods typical of $R_2$ (150 s), the gain of the Uppsala Wiechert is 0.9. At an average distance $\Delta = 270°$, an amplitude of 1 mm on a horizontal record (or 1630 $\mu$m of vertical ground displacement) leads to $M_m^{TD} = 9.03$, justifying our rule of thumb. Obviously, this result is for an average excitation, and an average distance. Nevertheless, it proved remarkably robust when assessing events of known moments.

At this stage, the nature of the further study depended fundamentally on the information available on the event's focal mechanism and seismic moment.

1. Events for which a detailed study (focal mechanism, depth, and moment) is available in the literature are listed in Table 1. In order to allow for easy comparison with our measurements, the published moment is reported as $M_m^P = \log_{10} M_0 - 20$, where $M_0$ is in dyn-cm. A range of $M_m^P$ is given when several studies list significantly different values. The range of $M_m$ values obtained from the various records considered is listed in Column 5 of Table 1.

   For these 25 events, we proceeded to compute the "corrected magnitude," $M_c$ as defined in Paper I, using the published focal geometry and exact depth. This computation amounts to correcting for the effect of using an average focal mechanism, rather than the exact source-station geometry, when estimating the excitation of the surface wave by the source. As discussed in Paper I, the latter procedure is the source of some systematic errors, and the use of $M_c$ on modern datasets results in a variance reduction of approximately 30–40% for the residuals $r$.

   At this point, and when warranted, we reassessed the value of $M_0$ using the corrected mantle magnitude $M_c$.

2. Fifteen events for which there exists a published seismic moment, but no detailed seismological study of the earthquake (in particular of its focal mechanism) are listed in Table 2. In general, the moment estimate has been obtained from either an assessment of the aftershock area or the evaluation of a tusnami magnitude. Another group consists of earthquakes for which BRUNE and ENGEN (1969) have given a 100-s spectral amplitude $X_{100}$. Following their algorithm, it is

Table 1

*Events for which a published moment and focal geometry are available from a detailed seismological study*

| Date | Epicenter | Published Moment | | | | Preferred $M_0$ ($10^{27}$ dyn-cm) | |
|---|---|---|---|---|---|---|---|
| | °N; °E | $M_m^p$ | Reference | $M_m$ | $M_c$ | Retained | Adjusted |
| 09 JUL 1905 | 49; 99 | 8.6 | a | 8.31–9.03 | 8.50–8.60 | | 20–40 |
| 23 JUL 1905 | 49; 98 | 8.6 | a | 8.33 | 8.30 | | 20–40 |
| 18 APR 1906 | 38; −123 | 7.97–8.00 | b, c | 8.54 | 8.13 | | 14 |
| 01 SEP 1923 | 35.25; 139.5 | 7.88 | d | 8.07–8.73 | 8.43–8.46 | | 29 |
| 07 MAR 1929 | 51; −170 | 7.84 | e | 8.38 | 8.02 | | 10 |
| 02 MAR 1933 | 39.25; 144.50 | 8.63 | f | 8.50–9.63 | 8.60–9.16 | | 95 |
| 22 AUG 1949 | 53.75; −133.25 | 8.06 | b | 8.76 | 8.37 | | 23 |
| 15 AUG 1950 | 28.5; 96.5 | 8.97–9.39 | g, h | 8.96–9.03 | 9.07–9.23 | | 140 |
| 18 NOV 1951 | 30.5; 91 | 7.28 | g | 7.54 | 7.43 | 1.9 | |
| 04 NOV 1952 | 52.75; 159.5 | 9.54 | i | 9.13–9.39 | 9.03–9.52 | 350 | |
| 04 DEC 1957 | 42.25; 99.4 | 8.26 | j | 8.02–8.52 | 8.03–8.40 | 18 | |
| 10 JUL 1958 | 58.6; −137.1 | 7.72–8.46 | b, c | 7.99–8.46 | 7.57–8.10 | | 6 |
| 06 NOV 1958 | 44.38; 148.58 | 8.64 | k | 7.92–8.78 | 8.02–8.78 | 44 | |
| 22 MAY 1960 | −38.5; −73.2 | 10.30–10.51 | l, m | 9.60–10.57 | 9.76–10.88 | 3200 | |
| 20 NOV 1960 | −6.72; −80.90 | 7.53 | n | 6.85–7.43 | 7.16–7.91 | 2.7 | |
| 13 OCT 1963* | 44.8; 149.5 | 8.83 | o | 8.31–8.80 | 8.68 | 67 | |
| 28 MAR 1964 | 61.04; 147.73 | 9.91 | p | 9.45–9.57 | 9.48–9.59 | 820 | |
| 04 FEB 1965 | 51.3; 178.6 | 9.10 | q | 8.91 | 8.95 | 125 | |
| 11 AUG 1969* | 43.54; 147.35 | 8.34 | r | 8.31–8.62 | 8.50 | 22 | |
| 19 AUG 1977* | −11.09; 118.46 | 8.56 | s | 9.21 | 8.82 | 36 | |
| 12 DEC 1979 | 1.6; −79.36 | 8.23–8.46 | t, u, v | 8.14 | 8.48 | 29 | |
| 03 MAR 1985* | −33.13; −71.87 | 8.01 | w | 7.82–7.96 | 7.88–7.90 | 10.2 | |
| 19 SEP 1985* | 18.19; −102.53 | 8.04 | x | 8.50 | 8.26 | 11 | |
| 07 MAY 1986 | 51.52; −174.78 | 8.02 | y | 7.83–8.64 | 8.09–8.40 | 10.5 | |
| 23 MAY 1989* | −52.24; 160.20 | 8.15 | z | 8.76 | 8.29 | 14 | |

* Event selected for an independent estimate of the stability of the UPP Wiechert after 1960.

References for Published Moments:

a: OKAL (1977); b: BEN-MENAHEM (1978); c: KANAMORI (1977) (from the aftershock area); d: KANAMORI (1971a); e: KANAMORI (1972a); f: KANAMORI (1971b); g: CHEN and MOLNAR (1977); h: BEN-MENAHEM *et al.* (1974)†; i: KANAMORI (1976); j: OKAL (1976); k: FUKAO and FURUMOTO (1979); l: KANAMORI and CIPAR (1974); m: CIFUENTES and SILVER (1989) (this figure represents only the "main shock"); n: PELAYO and WIENS (1990); o: KANAMORI (1970a; 1977); p: KANAMORI (1970b); q: WU and KANAMORI (1973); r: ABE (1973); s: DZIEWONSKI *et al.* (1987a); t: DZIEWONSKI *et al.* (1987c); u: KANAMORI and GIVEN (1981); v: BECK and RUFF (1984); w: DZIEWONSKI *et al.* (1985); x: DZIEWONSKI *et al.* (1986); y: DZIEWONSKI *et al.* (1987d); z: DZIEWONSKI *et al.* (1990).

† These authors report $M_0 = 2.5 \times 10^{30}$ dyn-cm, the result of an obvious clerical error; they clearly mean $M_0 = 2.5 \times 10^{29}$ ($M_m^p = 9.39$).

Table 2

*Events for which a published moment is available based on a 100-second magnitude, a tsunami magnitude, or a study of the aftershock area*

| Date | Epicenter | Published Moment | | | $M_m$ | Reference for focal mechanism | $M_c$ | Proposed moment ($10^{27}$ dyn-cm)* |
|---|---|---|---|---|---|---|---|---|
| | °N; °E | $M_m^P$ | Nature * | Reference | | | | |
| 31 JAN 1906 | 1; −81.5 | 9.31 | C | a | 8.32–8.49 | b | 8.50–8.60 | 60 A |
| 11 NOV 1922 | −28.5; −70 | 8.83–9.15 | B, C | a, c | 8.56–8.61 | d | 8.61–8.63 | 42 A |
| 03 FEB 1923 | 54; 161 | 8.57–9.30 | B, C | a, c | 8.86–8.97 | e | 8.65–8.79 | 55 A |
| 17 JUN 1928 | 16.25; −98 | 8.08 | C | a | 7.76–8.30 | f | 8.37–8.45 | 23 A |
| 03 JUN 1932 | 19.5; −104.25 | 8.00–8.18 | A, C | a, g, h | 8.41 | i | 8.21 | 16 R |
| 01 FEB 1938 | −5.25; 130.5 | 8.63–8.85 | A | a, g | 8.50–9.04 | j | 8.60–9.08 | 75 R |
| 10 NOV 1938 | 55.5; −158 | 8.45 | A, C | a, g | 8.52–8.76 | k | 8.64–8.69 | 46 A |
| 24 MAY 1940 | −10.5; −77 | 8.40 | C | a | 7.73–8.28 | d, l, m | 8.20–8.37 | 25 R |
| 24 AUG 1942 | −15; −76 | 7.89–8.43 | A, C | a, g | 7.55–8.38 | d | 7.97–8.25 | 13 A |
| 06 APR 1943 | −30.75; −72 | 8.18 | C | a | 7.77–8.02 | d | 7.86–8.35 | 18 A |
| 20 DEC 1946 | 32.5; 134.5 | 8.18 | | n | 8.08–8.69 | o | 8.19–8.80 | 30 A |
| 04 MAR 1952 | 42.5; 143 | 8.20–8.58 | A, B | c, g, p | 7.93–9.10 | q | 8.72–8.93 | 60 A |
| 25 NOV 1953 | 33.9; 141.5 | 7.95 | | r | 7.84 | s | | 8.9 R |
| 09 MAR 1957 | 51.3; −175.8 | 9.29–9.77 | C | a, t | 8.11–8.62 | u | 8.49–8.97 | I |
| 04 MAY 1959 | 52.5; 159.5 | 8.42 | C | a | 7.49–8.35 | q | 7.77–8.09 | 9.5 A |

* Nature of Published Moment: (A): inferred from 100-s magnitude; (B): inferred from tsunami magnitude; (C): inferred from aftershock area. Proposed moment: (R): Retained; (A): Adjusted; (I): Insufficient data, see Appendix.

References for Published Moments, and Focal Mechanisms used in computing $M_c$

a: KANAMORI (1977); b: CMT solution (12 Dec 1979) (DZIEWONSKI *et al.*, 1987c); c: ABE (1983); d: Assumed local interplate motion; e: Assumed mechanism of 1952 event (KANAMORI, 1976); f: CMT solution (29 Nov 1978) (DZIEWONSKI *et al.*, 1987b); g: BRUNE and ENGEN (1969); h: ESPÍNDOLA *et al.* (1981); i: Average geometry for the area as given by SINGH *et al.* (1984); j: Mechanism adapted from nearby CMT solution (25 Jul 1988) (DZIEWONSKI *et al.*, 1989b); k: Adapted from nearby CMT solution (13 Feb 1979) (DZIEWONSKI *et al.*, 1987c); l: ABE (1972); m: DEWEY and SPENCE (1979); n: KANAMORI (1972b) (moment from a comparison with 1923 and 1933 events); o: mechanism slightly adjusted from KANAMORI (1972b); p: ABE (1975); q: DENHAM (1977); r: M. Ando, quoted by ABE and KANAMORI (1980); s: No attempt was made to compute an $M_c$ value; t: RUFF *et al.* (1985) (from long-period body-wave modeling); u: CMT solution (07 May 1986) (DZIEWONSKI *et al.*, 1987d).

possible to convert $X_{100}$ to $M_0$ through

$$\frac{M_0}{X_{100}} = 1.349 \times 10^{27} \text{ dyn/s.} \tag{6}$$

In many instances, several estimates of $M_0$, obtained by one or more of these methods at a number of stations, can be found in the literature. The corresponding range of $M_m^p$ is listed in Table 2. After the values of $M_m$ were listed in Column 6 of Table 2, an attempt was made to compute $M_c$ based on a mechanism representative of a possible earthquake geometry at the particular epicenter: for example, we analyze available CMT solutions for modern shocks in the area, or we assume that the event represents interplate motion along a plate boundary, as exemplified by later events with a published detailed study. In all cases, it must be emphasized that such an interpretation remains tentative, and occasionally speculative: CMT solutions for relatively small events (in the $10^{26}$ dyn-cm range) may not apply to larger shocks; also, when working with large earthquakes at plate boundaries, we cannot discard the possibility of a large normal faulting event expressing the decoupling of a segment of lithosphere, such as the 1933 Sanriku or 1929 Aleutian shocks (KANAMORI, 1971b, 1972a). When several measurements are available (either Love and Rayleigh at the same station, or at a number of stations), it is possible to put some constraints on the possible geometries. In some regions, such as the Banda Sea where the tectonic framework is very complex (as shown for example by a large variety of geometries for the CMT solutions), we did not compute a corrected magnitude $M_c$.

In a strongly nodal geometry, the focal mechanism correction computed to obtain $M_c$ can reach 1 to 1.5 units of magnitude. Yet, as shown on several examples in Papers I and II, the measured value of $M_m$ may not be significantly deficient, due either to lateral heterogeneity, which may channel energy outside the great circle path, or occasionally to a slight error in the focal mechanism itself. As a result, under circumstances when a focal mechanism is either published or estimated (mostly from the plate tectonics framework), we tried to avoid the use of stations located in pronounced nodes of the radiation pattern.

3. Finally, Table 3 lists eight earthquakes for which no moment was available in the literature, at least to the best of our knowledge. In such cases, we usually attempted to compute a value of $M_c$, based on a focal mechanism estimated along the lines explained above, and propose a value of $M_0$ for the event.

For earthquakes in the Pacific Basin, we also interpreted our results in the framework of the teleseismic tsunami generated by the event, as quantified by ABE (1983), or more generally described in SOLOV'EV and GO's (1984), and SOLOV'EV *et al.*'s (1986) monumental compilations[3].

[3] Unfortunately, SOLOV'EV and GO (1984) do not provide any information on Kamchatka tsunamis, including the large 1923 and 1952 events.

Table 3

*Events for which no published moment is available (to the best of our knowledge)*

| Date | Epicenter °N | Epicenter °E | $M_m$ | Reference for mechanism | $M_c$ | Proposed $M_0$ ($10^{27}$ dyn-cm) |
|---|---|---|---|---|---|---|
| 17 AUG 1906 | 51 | 179 | 7.99–8.55 | a | 8.03–8.33 | 15 |
| 01 MAY 1917 | −29 | −177 | 8.07–8.24 | b | 8.11–8.32 | 16 |
| 26 JUN 1917 | −15.5 | −173 | 7.89–8.48 | c | 8.11–8.13 | 13 |
| 30 APR 1919 | −18.3 | −173.1 | 8.22–8.81 | b | 8.15–8.63 | 25 |
| 14 MAY 1942 | −0.75 | −81.5 | 8.00 | d | 7.86 | 7.2 |
| 24 JAN 1948 | 10.5 | 122 | 7.55 | b | 7.60 | 4 |
| 02 NOV 1950 | −6.5 | 129.5 | 7.51–7.79 | e | | 5 |
| 02 DEC 1950 | −18.25 | 167.5 | 7.71–7.98 | b | 7.72–7.97 | 8 |

References for Mechanisms:
a: Assumed to represent decoupling failure of plate seaward of trench; b: Assumed local interplate motion; c: Similar to recent nearby CMT solution (01 Sep 1981) (DZIEWONSKI *et al.*, 1988); d: similar to recent nearby CMT solution (12 Dec 1979) (DZIEWONSKI *et al.*, 1987c); e: No attempt made to compute $M_c$; tentative value of $M_0$ derived from $M_m$.

## *Results and Discussion*

A total of 49 events were studied. Detailed results are listed in the Appendix, in chronological order. In the case when two events (mostly one historical, one modern) occurred at a similar location, these events are treated together. Figure 3 presents a map of all events studied, with special symbols keyed to the particular results obtained for each event.

### *Wiechert Performance at UPP after 1960*

As explained above, periodic calibration of the Wiechert instrument at UPP ceased in 1960. In order to draw any comparisons between pre- and post-1960 records, we need an independent assessment of the stability of the instrument after 1960, and of the practice of using the last available calibration for post-1960 records. For this purpose, we consider the 6 large events flagged by an asterisk in Table 1, and shown as open squares on Figure 3. Each of these shocks has been the subject of several individual studies, and their published moments can be considered reliable. In order not to affect any further discussion, we purposely reject several earthquakes whose study is detailed later in the paper (e.g., Alaska 1964; Ecuador, 1979). The residuals $r_c$ for the dataset of 6 events have an average $\bar{r} = 0.09$ and a standard deviation $\sigma = 0.18$. These numbers are typical of the performance of an individual station for modern records (see Papers I and II), and suggest that no serious drift of the instrument response took place at UPP after 1960.

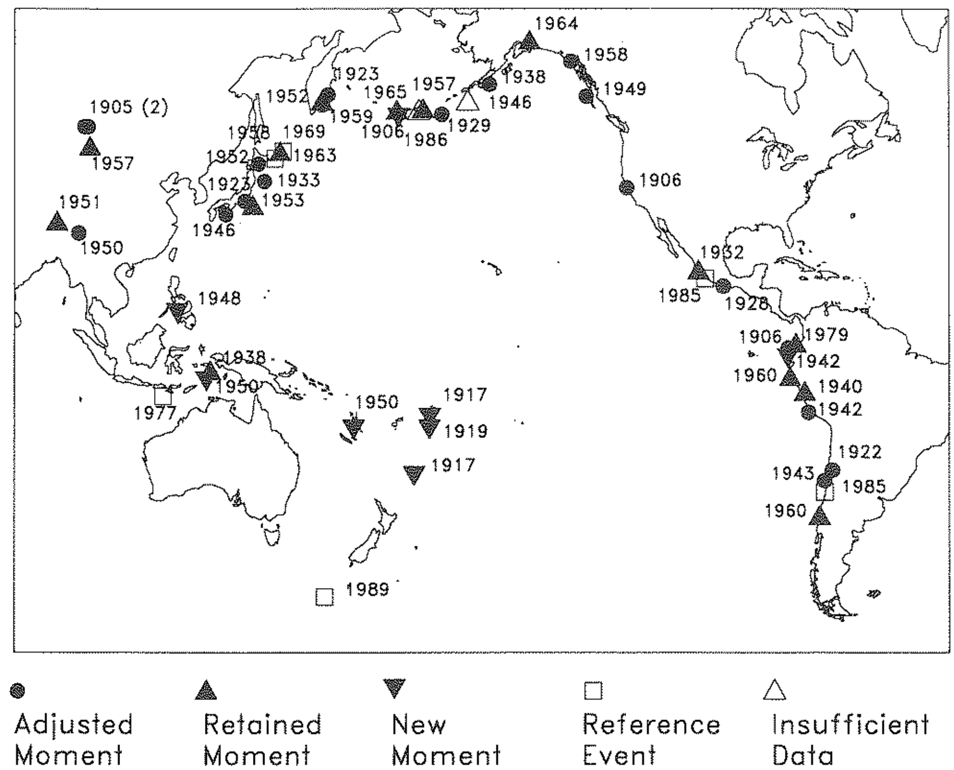


Figure 3

Map of all 49 events studied in this paper. Different symbols are used depending on the eventual result of the study (see text and Appendix for details). The year of each event is also plotted next to its location.

*Events with a Detailed Study: Performance of $M_m$ for Gigantic Earthquakes*

Figure 4 presents the measured values of $M_m$ and $M_c$ as a function of the published moment, for 19 earthquakes with available detailed studies, after we eliminated the six events used in the verification of the post-1960 UPP response (see above). This figure is comparable to Figures 8 (Paper I) or 6 and 7 (Paper II). Furthermore, we conducted some statistical computations on this dataset of 80 measurements.

1. The average value of the residual $[r = M_m - \log_{10} M_0 + 20]$ is $\bar{r} = -0.11$ and its standard deviation $\sigma = 0.45$. For the corrected values, $\bar{r}_c = -0.05$ and $\sigma_c = 0.32$. These numbers deserve some comments. First, the quality of the average residuals is excellent. The fact that it is actually better than in our original studies of modern data is obviously not significant, in view of the larger standard deviations. Not unexpectedly, the latter express, at least partly, the general

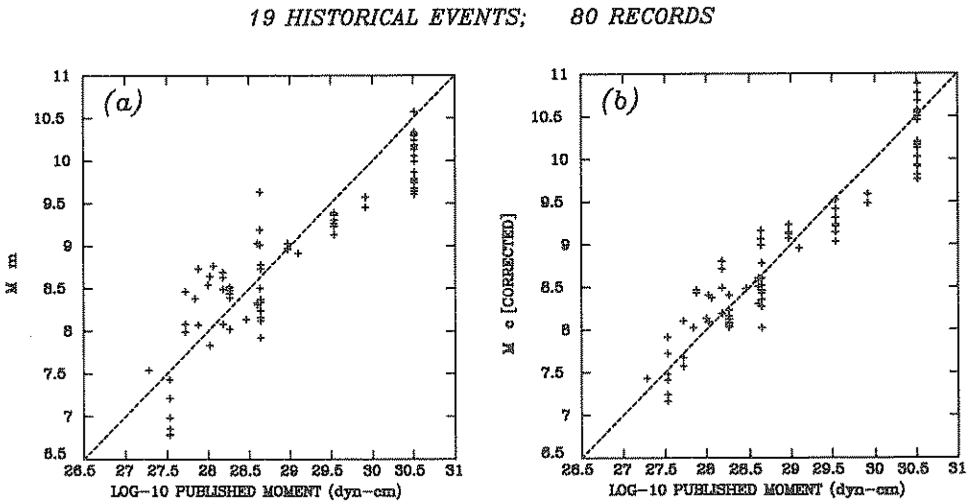


Figure 4

*Left* (a): $M_m$ values measured on individual records as a function of published moment, for those 19 earthquakes which have a detailed study published. The dashed line is the theoretical curve $M_m = \log_{10} M_0 - 20$. *Right* (b): Same as Left, for the corrected values $M_c$.

degradation of the quality of the historical data. But is should be emphasized that large (positive or negative) residuals can also reflect inaccuracy in the published moment. The Appendix details a few such cases, notably the 1923 Kanto and 1933 Sanriku earthquakes, and to a lesser extent the 1949 Queen Charlotte Islands events, which we think were underestimated in the literature. The cluster of points with strong negative residuals on Figure 4a pertains to the 1960 Peruvian earthquake, for which most available records are from nodal stations, and is thus an artifact of focal radiation (see Appendix for details). Given these remarks, we suggest that $M_m$ performs very well for historical events. This will be the rationale for the estimation of moments in the following sections.

2. The regression of $\log_{10} M_0$ as a function of $M_m$ yields a slope of 1.04. This figure clearly indicates that $M_m$ keeps growing with seismic moment, even for the largest events, and satisfactorily avoids saturation effects due to source finiteness. Individual $M_m$ values for the 1960 Chilean and 1964 Alaskan earthquakes stand out alone in the 9.5 to 10.5 range. The regressed slope of $\log_{10} M_0$ vs. $M_m$ for the data subset $M_0 \geq 10^{29}$ dyn-cm is 0.95, with $M_m$ correctly ranking the Chilean event as larger than the Alaskan, in contrast to BRUNE and ENGEN's (1969) $M_{100}$, which started to saturate for earthquakes of this size. A more detailed study of these events, from an extended dataset, and notably a discussion of time-domain estimates $M_m^{TD}$, can be found elsewhere (OKAL and TALANDIER, 1991), but the crux of the matter remains that the mantle magnitude $M_m$

recognizes the Alaskan and Chilean events as truly gigantic, even despite the generally unfavorable focal geometries for the few records we were able to gather. In the context of tsunami warning, it is probable that an $M_m$ measurement on early surface wave passages ($R_2$ to $R_4$) from the Chilean event, would have anticipated its exceptional tsunami generation and the extreme danger to be expected at all Pacific Basin sites.

Events for which we confirmed the published seismic moment (whether obtained from a detailed study or from an estimate of the source area) are shown as solid upward triangles on Figure 3.

*Reassessment of Old Moments*

We propose to adjust the seismic moments of 19 historical shocks, ranging in date from 1905 to 1959, and shown as circles on Figure 3. Eight of them were previously studied in detail, but occasionally within a narrower frequency band (e.g., the 1923 Kanto earthquake, for which we now propose $2.9 \times 10^{28}$ dyn-cm), or with a possibly inappropriate focal mechanism (e.g., the 1950 Assam event, for which we prefer CHEN and MOLNAR's (1977) geometry, leading to $1.4 \times 10^{29}$ dyn-cm). On the other hand, the remaining 10 had their moments estimated from their aftershock area, and our study provides a direct estimate of their true size from their seismic waves.

All individual cases are discussed in detail in the Appendix. The amount of adjustment ranges from $-0.51$ units of magnitude (Ecuador, 1906) to $+0.60$ (Tokachi-Oki, 1952). Apart from these events, the most significant adjustments were for the 1959 Kamchatka earthquake, which we think was significantly overestimated, and the 1933 Sanriku and 1949 Queen Charlotte Islands earthquakes, which on the contrary were probably greater than previously published.

*New Moments*

We list in Table 3 eight new moments, in the case of earthquakes of which only $M_s$ estimates were previously published. These earthquakes, shown as downward triangles on Figure 3, are of two types: very old (pre-1920) and relatively large, or more recent (post-1940) and smaller. Among them the 1917 and 1919 earthquakes in Tonga-Kermadec range between 1.6 and $2.5 \times 10^{28}$ dyn-cm. In their characterization of the level of coupling of those subduction zones, RUFF and KANAMORI (1980) used $M_s$ values (8.1 and 8.3, respectively), and interpreted them as $M_w$ (equivalent $M_m^p$ values would be 8.20 and 8.55). Our determinations indicate that the resulting error was negligible ($+0.05$ and $+0.15$ units of magnitude respectively), but this verification was nevertheless warranted.

On the other hand, we confirm that some of the largest earthquakes reported (in terms of conventional magnitudes $M_s$ or $M_{\mathrm{PAS}}$) at weakly coupled subduction zones, such as the Philippines and Vanuatu, are indeed relatively small in terms of moment (4 and $8 \times 10^{27}$ dyn-cm respectively).

Finally, and as detailed in the Appendix, in two instances, we could not propose a value of $M_0$; these earthquakes are shown as open upward triangles on Figure 3. In the case of the 01 April 1946 Aleutian earthquake, there is some evidence that the moment grows significantly with period, but most of the records are on relatively insensitive electromagnetic instruments; in addition, it has been suggested that the event may not be properly described by a double-couple (KANAMORI, 1985). In the case of the Aleutian earthquake of 09 March 1957, only one record (at Abuyama) gives an $M_c$ (9.27) approaching the lower bound of published values. It is probable that this event was exceptionally slow (LANE and BOYD, 1990), and its low-frequency source spectrum could not be reliably retrieved due to the noise level on the available records.

## *Conclusion and Possible Directions of Future Research*

The conclusions of this study can be summarized as follows:

1. The performance of the magnitude $M_m$ for truly gigantic events (Chile, 1960; Alaska, 1964) is excellent: the extreme character of these events is perfectly recognized and their seismic moments estimated with an accuracy comparable to that achieved for smaller events. This point is of prime importance since it demonstrates that $M_m$ does not saturate with increasing moment. It validates the whole endeavor, and should have important implications for rapid tsunami warning.
2. Valuable moment information can be retrieved from datasets of as few as one record and for events going as far back as 1905. When several records are available, it is often possible to draw some inferences on focal geometries.
3. By broadening considerably the range of frequencies where measurements are taken, a better estimate of the static moment $M_0$ can be obtained, as compared to previously published values. We propose to adjust 19 published values of $M_0$ for historical earthquakes. Most interesting are the cases of the 1923 Kanto earthquake, for which we show a significant long-period component of moment release, thus reconciling its value with geodetic observations, and the 1906 Ecuador earthquake, which, while significantly larger than its 1979 counterpart, remains smaller than published on the basis of its rupture area. Table 4 gives a revised list of the largest events in terms of seismic moment measured from spectral amplitudes of seismic waves.
4. The study of "maximum size" earthquakes in subduction zones for which no previous maximum moments had been published (Tonga, Kermadec) confirms that the values extrapolated from $M_s$ were adequate, even though this conclusion

Table 4

*Shallow earthquakes with moments greater than $5 \times 10^{28}$ dyn-cm*

| Date | Region | $M_0$ ($10^{27}$ dyn-cm) | Retained (R) or Adjusted (A) |
|---|---|---|---|
| 22 MAY 1960 | Chile | 3200† | R |
| 28 MAR 1964 | Alaska | 820 | R |
| 04 NOV 1952 | Kamchatka | 350 | R |
| 15 AUG 1950 | Assam | 140 | A |
| 04 FEB 1965 | Aleutian | 125 | R |
| 02 MAR 1933 | Sanriku | 95 | A |
| 01 FEB 1938 | Banda Sea | 75 | R |
| 13 OCT 1963 | Kuriles | 67 | R |
| 31 JAN 1906 | Ecuador | 60 | A |
| 04 MAR 1952 | Hokkaido | 60 | A |
| 03 FEB 1923 | Kamchatka | 55 | A |

† This represents only the "main shock" part of the event (CIFUENTES and SILVER, 1989).

*Note:* The 1957 Aleutian earthquake is not included for lack of sufficient constraints on its moment, but it is clear that it should be part of this dataset.

was by no means foregone. In other areas such as Vanuatu and the Philippines, we confirm that the largest reported events remain small by subduction zones standards (an illustration of the decoupled nature of the subduction), and this despite very large values of $M_s$.

5. Occasional discrepancies between Love and Rayleigh components at the same station, or between stations, suggest that some large earthquakes may have been normal faulting decoupling events. Such suggestions remain very speculative, and it is clear that more work is needed to fully understand those shocks. These and other candidates for further study include the 1952 Tokachi-Oki earthquake, the 1938 Banda Sea earthquake, and the 17 August 1906 pair in the Aleutians and Chile.
6. We fail to quantify only two events: the 1946 Aleutian earthquake, which may not be correctly represented by a double-couple (KANAMORI, 1985), and the 1957 Aleutian event, whose low-frequency source spectrum cannot be properly retrieved from the instruments available at the time (LANE and BOYD, 1990).

*Acknowledgements*

A study of this nature owes considerably to many persons: first and foremost, to the staff of seismological observatories worldwide, many of these individuals now bygone and their names forgotten, who, for so many decades, have diligently operated the stations, and provided the necessary calibrations. Next, it is a pleasure to acknowledge the contribution of the present curators of seismological collections: at the Caltech archives, I am grateful to Paul Roberts, for copying and filing countless seismograms, and to Hiroo Kanamori for answering many questions regarding instrument responses. At the University of Uppsala, Ota Kulhánek opened for me the doors of their superb collection and provided excellent facilities for *in situ* digitizing. Jim Taggart welcomed me repeatedly at the USGS Center in Denver. A few records were also obtained from Michel Cara at Strasbourg, Bob Herrmann at Saint Louis University, Kunihiko Shimazaki and Katsuyuki Abe at the University of Tokyo; and second-hand from Doug Wiens at Washington University in Saint Louis and Tom Boyd at the Colorado School of Mines; Inés Cifuentes shared her expertise on the response of IGY Press-Ewings, and provided a useful review of the original draft of the paper. I am also grateful to Göran Ekström and Gretchen Zwart, who provided regular updates of their computerized catalog of CMT solutions in advance of formal publication. Finally, Todd Retzlaff digitized a large part of the paper records copied at Pasadena; a few more were digitized by Marianne Okal.

The concept of $M_m$, the variable-period mantle magnitude, was developed in the course of many years of collaboration and friendship with Jacques Talandier and the present study obviously owes much to him.

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## *Appendix: Discussion of Individual Events*

• *09 and 23 July 1905; Mongolia*

Records processed: UPP

These earthquakes were studied by OKAL (1977), who proposed moments of about $5 \times 10^{28}$ dyn-cm ($M_m^p = 8.7$) for both events, on the basis of surface-wave modeling of the records at Göttingen, but stressed that these numbers were merely orders of magnitude. We obtained both NS and EW usable records at Uppsala for the first event, but only a NS record (mostly Love polarization) for the second one. These records yield $M_m = 8.31$ (Love) and 9.03 (Rayleigh) for the first event, and $M_m = 8.33$ (Love) for the second one.

Assuming the mechanism was pure strike-slip along the Bolnai Fault ($\phi = 100°$; $\delta = 90°$; $\lambda = 0°$), we obtain $M_c = 8.30$ (Love) and 8.66 (Rayleigh) for the first event, and $M_c = 8.30$ for the second one. The disparity between the two estimates of $M_c$ for the first event suggests that it may have occurred on a conjugate fault, possibly the Tsetserleg Fault branching away from the Bolnai Fault in a NE trend at about

98.3°E. This suggestion would be corroborated by reports of extensive damage during Event I North of the Bolnai Fault (OKAL, 1977). We therefore computed $M_c$ values for Event I (8.50 for Rayleigh and 8.60 for Love) for a normal faulting mechanism along the Tsetserleg Fault ($\phi = 55°$; $\delta = 45°$; $\lambda = 270°$). The fact that these values are less scattered than for the strike-slip mechanism would generally support associating the event with the Tsetserleg Fault, although this interpretation remains highly speculative.

At any rate, we confirm the general order of magnitude of these events' moments ($M_0 = 2 - 4 \times 10^{28}$ dyn-cm).

• *31 January 1906 and 12 December 1979; Colombia-Ecuador*

Records processed: UPP

It is particularly important to reassess the size of the 1906 earthquake in order to compare it quantitatively to the 1979 event at the same location. The 1906 shock was given a moment of $2.04 \times 10^{29}$ dyn-cm ($M_m^p = 9.31$) by KANAMORI (1977) on the basis of its aftershock area, although KELLEHER (1972) indicated that his estimate of the extent of rupture was from "marginal evidence." The 1979 earthquake has a CMT moment of $1.7 \times 10^{28}$ dyn-cm ($M_m^p = 8.23$). KANAMORI and GIVEN (1981) suggested a moment of $2.9 \times 10^{28}$ dyn-cm ($M_m^p = 8.46$), while BECK and RUFF (1984) obtained a value of $2.0 \times 10^{28}$ ($M_m^p = 8.32$).

For the 1906 event, we find $M_m = 8.32$ (Love) and 8.49 (Rayleigh) at Uppsala. We were unable to identify the multiple passages which could be expected for an event in the range $M_0 \geq 10^{29}$ dyn-cm. For the 1979 event, the EW record is extremely faint, and only the Love waves could be used, yielding $M_m = 8.14$. Assuming the 1979 CMT focal mechanism applies to both events, $M_c$ would range between 8.68 (Love) and 8.83 (Rayleigh) for 1906; 8.48 for 1979, the latter in excellent agreement with KANAMORI and GIVEN's (1981) figure. The comparison of the UPP records from the same instruments (Figure A-1) upholds this conclusion: the 1906 event is certainly larger than the 1979 one, but not by the full order of magnitude previously reported. This is supported by the fact that the 1906 tsunami, while catastrophic locally, was only moderate in Hawaii (about 20–30 cm on the various islands), and is not reported in Japan (SOLOV'EV and GO, 1984). We suggest that the 1906 event had a moment not larger than $6 \times 10^{28}$ dyn-cm.

• *18 April 1906; San Francisco*

Records processed: UPP

We do not use Love waves, expected to be close to nodal at Uppsala, and measure $M_m$ only from rotated Rayleigh waves, yielding $M_m = 8.54$; in the predicted geometry, $M_c$ would fall to 8.13. This figure is slightly higher than

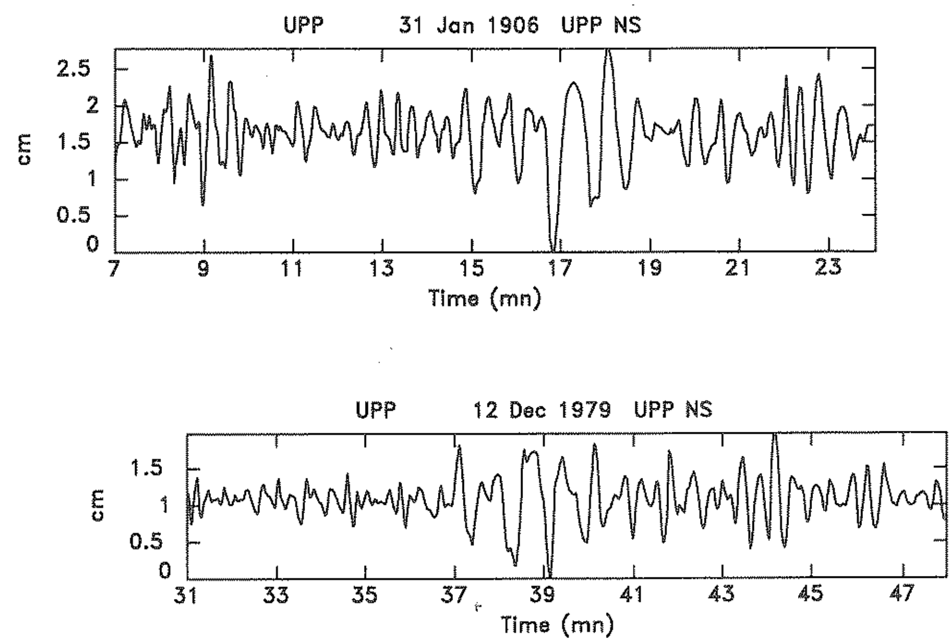


Figure A-1
Comparison of the Love wavetrains $G_1$ of the 1906 and 1979 Ecuador-Colombia earthquakes, as recorded on the NS component of the Uppsala Wiechert. The records are plotted on the same scale, with the abcissæ offset so as to align the $G_1$ wavetrains, thus allowing a direct comparison of their relative sizes. Note that while the 1906 earthquake is undoubtedly the larger of the two, it cannot have a moment 10 times larger than the 1979 event.

KANAMORI's (1977) estimate based on surface rupture ($M_m^p = 8.00$), and the value ($M_m^p = 7.97$) inferred from BEN-MENAHEM's (1978) analysis of the event's "potency," which, incidentally, was restricted to frequencies greater than 0.01 Hz.

• *17 August 1906 and 04 February 1965; Aleutian Islands*

Records processed: UPP

A very large event took place in Chile, only 30 minutes after the 1906 Aleutian event. Only the first (Aleutian) earthquake could be studied; signals from the second (Chilean) event fall within the coda of the Aleutian record. It is worth noting, however, that no multiple passages could be identified at Uppsala from either event, thus suggesting that none of the two shocks reached $10^{29}$ dyn-cm. The second (Chilean) event was proabably the stronger of the two since the arrival times of the tsunami at Pacific stations fit distances from Chile rather than from the Aleutians (SOLOV'EV and GO, 1984). The tsunami was very damaging locally, but only moderate at teleseismic distances. $M_m$ for the Aleutian event ranges between

8.03 (Rayleigh) and 8.55 (Love). A significant problem with this event is that the expected subduction geometry (as given, for example by the mechanism of the 1965 earthquake [WU and KANAMORI, 1973]) worsens, rather than reconciles, the discrepancy between Rayleigh and Love waves at Uppsala ($M_c = 7.75$ and 9.27, respectively). In this situation, we speculate that the 1906 earthquake may actually be a decoupling event, similar to the 1977 Indonesian shock. In such a geometry ($\phi = 290°$, $\delta = 45°$, $\lambda = -110°$), we obtain $M_c$ values of, respectively 8.03 and 8.33, which are more mutually compatible than for the subduction mechanism. This geometry would also require a more southern epicenter, by about 1°, a distance that the precision of 1906 locations cannot resolve. This model, which remains speculative, would call for a moment of $1.5 \times 10^{28}$ dyn-cm.

Only the NS component could be used for the 1965 event, for which $M_m$ is a disappointing 8.6 with $M_c$ very comparable (8.53). This probably reflects the very slow character of this event, as demonstrated by BECK and CHRISTENSEN (1991): with a rise time of 160 s, $M_m$ measurements should be taken at or beyond 300 s, to avoid source finiteness effects. Indeed, if we attempt to retrieve $M_m$ at 340 s, a value of 8.91 is obtained, but the corresponding spectral amplitude is at the noise level on the record. The corresponding $M_c$ would be 8.95. At any rate, Figure A-2 clearly shows that the 1965 Rat Island event is significantly larger than the 1906 one.

• *01 May 1917; Kermadec Islands*

Records processed: UPP

$M_m$ reaches 8.07 (Love) and 8.24 (Rayleigh). We have no information on the possible mechanism of the event. Local earthquakes, including large ones, can depart significantly from the expected overthrusting between the two plates, as demonstrated by the event of 20 October 1986 (LUNDGREN *et al.*, 1989). Assuming however that the earthquake did represent interplate motion, we obtain $M_c = 8.11$ (Love) and 8.32 (Rayleigh). The corresponding moment ($M_0 = 1.6 \times 10^{28}$ dyn-cm) would be in line with the relatively small tsunami at teleseismic distances. SOLOV'EV and GO (1984) indicate that they doubt reports of gigantic waves on Samoa, more than 1500 km away.

• *26 June 1917; Samoa*

Records processed: UPP

This event was one of the strongest ever felt in Samoa, where it generated a destructive tsunami. The teleseismic tsunami was not, however, very large (SOLOV'EV and GO, 1984). $M_m$ values at Uppsala are significantly different for Love (7.89) and Rayleigh (8.48) waves. We can only speculate as to the nature of the focal mechanism of the event, since available CMT solutions for large shocks in the

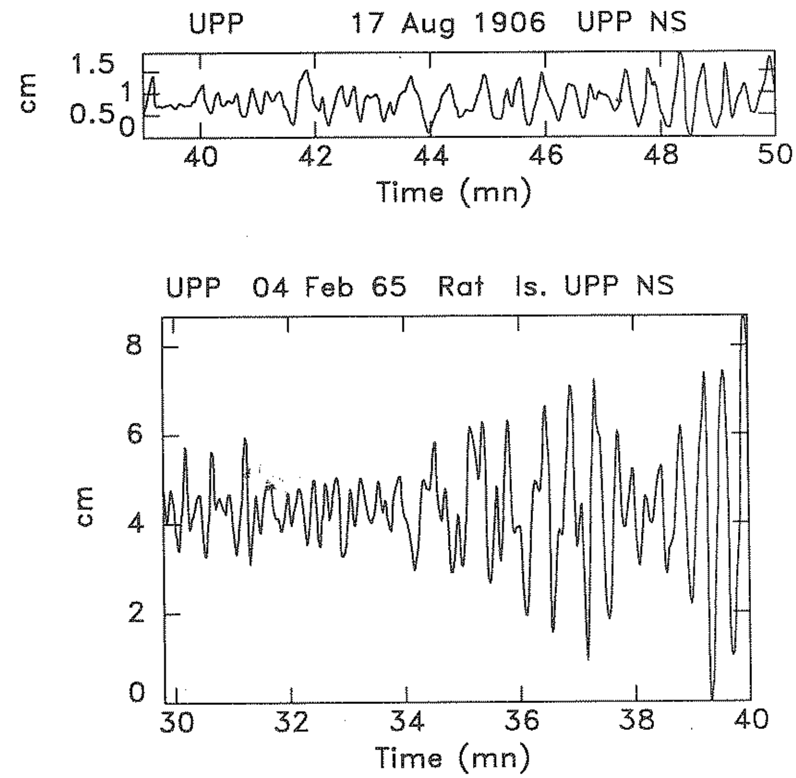


Figure A-2
Same as Figure A-1 for the $R_1$ wavetrains of the 17 Aug 1906 and 04 February 1965 events in the Aleutians. Note this time the much larger size of the recent event.

region show many kinds of geometries. However, it is possible to reconcile Love and Rayleigh measurements with a mechanism ($\phi = 273°$; $\delta = 40°$; $\lambda = 270°$) very close to that of the largest recent event, the normal faulting earthquake of 01 September 1981. We obtain values of $M_c$ of 8.13 (Love) and 8.11 (Rayleigh).

• *30 April 1919; Tonga*

Records processed: UPP, TOK

There is some uncertainty as to the exact epicenter of this earthquake. The ISS lists it at 21.2°S, 172.5°W, while Apia suggests 19.5°S; it is difficult to ignore the reported impulsive arrivals at Apia; our own relocation converges on 18.2°S, 173.1°W.

$M_m$ values are 8.53 (Rayleigh) and 8.22 (Love) at Uppsala; 8.83 (Rayleigh) and 8.81 (Love) at Hongo. Available focal mechanisms in the area vary widely; the largest CMT solution (02 April 1977) has a moment of only $10^{27}$ dyn-cm, and may not be representative of the geometry of a truly great earthquake, such as the 1919 event. For the latter, we tentatively adopt a reverse thrust mechanism derived from the local geometry of the subduction zone: $\phi = 196°$; $\delta = 43°$; $\lambda = 90°$; the resulting values of $M_c$ are 8.15 (UPP, Love), 8.38 (UPP, Rayleigh), 8.43 (TOK, Rayleigh), and 8.63 (TOK, Love). The moment of the earthquake is most probably around $2.5 \times 10^{28}$ dyn-cm.

• *11 November 1922; Chile*

Records processed: UPP

Our results are $M_m = 8.61$ (Love) and 8.56 (Rayleigh). They are practically unaffected by the presumable focal geometry ($\phi = 9°$; $\delta = 20°$; $\lambda = 110°$), with $M_c$ values of, respectively 8.61 and 8.63. They agree well with KANAMORI's (1977) estimate from the aftershock area ($M_m^p = 8.83$), but fall somewhat short of ABE's (1983) estimate ($M_t = 8.7$, equivalent to $M_m^p = 9.15$).

• *03 February 1923; Kamchatka*

Records processed: UPP, STR

Both Rayleigh and Love waves at Uppsala give consistent results, $M_m = 8.97$ (Rayleigh) and 8.94 (Love), also in agreement with our pilot study at Strasbourg (8.86). $M_c$ values computed for an interplate mechanism similar to that of the nearby great 1952 event ($\phi = 214°$; $\delta = 30°$; $\lambda = 100°$) are a little lower (8.76, 8.79 and 8.65, respectively). This earthquake seems somewhat larger than estimated ($M_m^p = 8.57$) from its aftershock area (KANAMORI, 1977), but smaller than ABE's (1983) estimate from tsunami data ($M_t = 8.8$, equivalent to $M_m^p = 9.3$).

• *01 September 1923; Kanto, Japan*

Records processed: UPP

$M_m$ values are significantly different for Love (8.07) and Rayleigh (8.73). However, based on KANAMORI's (1971a) focal geometry, $M_c$ values are in much better agreement (8.46 and 8.43, respectively). Thus, this event is significantly larger than estimated in KANAMORI's (1971a) seismological study ($M_0 = 7.6 \times 10^{27}$ dyn-cm).

We note that our $M_m$ measurements are retained at the long-period end of the available spectrum (170 s), and that spectral amplitudes have a tendency to increase sharply with period. Spectral amplitudes around 80–100 s would give results in generally good agreement with KANAMORI's (1971a) values, measured at 80 s

($M_m^p = 7.88$). On the basis of the UPP seismograms, we would tend to prefer a value around 8.45 ($M_0 = 2.9 \times 10^{28}$ dyn-cm). We note that KANAMORI (1971a) observes a significant deficiency (of as much as a factor of 3) between his inferred seismic slip and the result of geodetic measurements. This discrepancy disappears if our value of $M_0$ is used. Thus, it is probable that the Kanto earthquake had a significant component of slow strain release.

• *17 June 1928; Oaxaca, Mexico*

Records processed: UPP

$M_m$ values are 7.76 (Love) and 8.30 (Rayleigh). The Love wave deficiency is easily explained by a subduction mechanism comparable to that of the 1978 Oaxaca event ($\phi = 130°$; $\delta = 82°$; $\lambda = 90°$), yielding $M_c$ values of 8.37 (Rayleigh) and 8.45 (Love). We prefer the Rayleigh value, since the nodal character of the Love wave renders the correction unstable. The resulting moment ($2.3 \times 10^{28}$ dyn-cm) is somewhat larger than KANAMORI's (1977) estimate ($1.2 \times 10^{28}$ dyn-cm) based on the aftershock area.

• *07 March 1929; Aleutian Islands*

Records processed: UPP

Only the NS (Rayleigh) record could be studied. The $M_m$ value is 8.38. KANAMORI (1972a) showed that this earthquake is actually a normal faulting rupture of the sinking slab, and modeled it in the geometry of a nearby 1965 shock studied by STAUDER (1968). We adopt a slightly modified mechanism, still satisfying all body-wave constraints given by KANAMORI (1972a) ($\phi = 265°$; $\delta = 45°$; $\lambda = -90°$) and obtain $M_c = 8.02$. While this value is slightly higher than KANAMORI's estimate ($M_0 = 6.7 \times 10^{27}$ dyn-cm or $M_m^p = 7.84$), it clearly confirms that this earthquake is not very large, despite its very high conventional magnitude ($M_{\mathrm{PAS}} = 8.6$).

• *03 June 1932; Jalisco, Mexico*

Records processed: UPP

Only the East-West component (mostly Rayleigh polarization) is available for processing. The resulting $M_m$ is 8.41, and $M_c = 8.21$ using the average focal mechanism suggested for this region by SINGH *et al.* (1984) ($\phi = 310°$; $\delta = 14°$ $\lambda = 90°$). The resulting moment ($M_0 = 1.6 \times 10^{28}$ dyn-cm) is in agreement with various estimates ranging from 1.0 to $1.5 \times 10^{28}$ dyn-cm (BRUNE and ENGEN, 1969; KANAMORI, 1977; ESPÍNDOLA *et al.*, 1981).

• *02 March 1933; Sanriku*

Records processed: UPP, PAS

The Sanriku earthquake is the first one for which adequate data exist at Pasadena, in the form of a record from Benioff's early strainmeter prototype, on which the phases $R_2$, $G_2$ and $R_4$ are clearly identifiable. The characteristics of the instrument are given in BENIOFF (1935); even though the instrument was in its developmental stages, and the constants were adjusted frequently at the time (H. Kanamori, pers. commun., 1991), the period of the galvanometer was apparently kept relatively constant (between 28 and 35 s) prior to May, 1937. As for magnification and damping, we can only speculate that they were not altered. We verified that the original constants published by Benioff ($T_g = 35$ s; $V = 100$; $h = 1$) give the correct standard magnitude ($M_s = 7.62$, as compared to Gutenberg's estimate of $M_{\mathrm{PAS}} = 7.7$) for the Panama earthquake of 18 July 1934, whose record is published in Benioff's paper. On this basis, we assume that the response of the instrument as described by BENIOFF (1935) is adequate for the Sanriku record, and obtain very strong $M_m$ values, ranging from 9.19 ($G_2$) to 9.63 ($R_2$ and $R_4$).

At Uppsala, we could use only the EW record, which contains both Love and Rayleigh wavetrains. We obtain $M_m$ values of 8.53 (Love) and 9.01 (Rayleigh).

This event was studied by KANAMORI (1971b), who obtained a moment $M_0 = 4.3 \times 10^{28}$ ($M_m^p = 8.63$), based on 100-second spectral amplitudes. His mechanism ($\phi = 347°$; $\delta = 46°$; $\lambda = 257°$) yields a group of very stable $M_c$ values ranging from 8.99 to 9.16. Only UPP (Love) has $M_c = 8.60$. These values, corresponding to $M_0 = 9.5 \times 10^{28}$ dyn-cm, are significantly larger than Kanamori's, but they are generally obtained at longer periods, and would help account for the destructive teleseismic tsunami (IIDA *et al.*, 1967).

• *01 February 1938; Banda Sea*

Records processed: UPP, PAS, COL

This event is clearly very large, as documented by the multiple passages apparent on the UPP Wiecherts. $M_m$ values are 9.04 ($R_1$), 8.97 ($G_1$), 8.76 ($G_2$) and 9.00 ($G_3$) at Uppsala, 8.50 ($G_1$), 8.76 ($G_3$) and 8.89 ($R_1$) at College. We also obtained a North-South strainmeter record at Pasadena, on which the Love wavetrains $G_1$ to $G_4$ are very prominent, with $M_m$ values ranging between 8.75 and 8.95.

No focal mechanism information is available for the event; the few ISS reports of first motion are predominantly dilatational, suggesting normal faulting. CMT solutions for shallow events in the vicinity of the epicenter show a complete spectrum of thrust, normal and strike-slip events. A normal faulting mechanism close to an available CMT ($\phi = 15°$; $\delta = 35°$; $\lambda = -83°$) yields $M_c$ ranging from 8.60 to 9.08, with most values above 8.80. While further study is warranted to resolve the focal mechanism, this clearly establishes this event as a very large earthquake indeed, with a probable moment around $7.5 \times 10^{28}$ dyn-cm.

• *10 November 1938; Alaska*

Records processed: UPP

$M_m$ values are 8.52 (Love) and 8.76 (Rayleigh). No focal mechanism information is available, but we assume that the event represents interplate underthrusting in the geometry of recent CMT solutions ($\phi = 236°$; $\delta = 16°$; $\lambda = 80°$), and obtain $M_c = 8.64$ (Love), 8.69 (Rayleigh), somewhat larger than proposed by KANAMORI (1977) ($2.8 \times 10^{28}$; $M_m = 8.45$) on the basis of aftershock area and BRUNE and ENGEN's (1969) 100-s spectral amplitudes.

• *24 May 1940; Peru*

Records processed: UPP, PAS

$M_m$ values are 7.92 (Love) and 8.28 (Rayleigh) at Uppsala, but only 7.73 to 7.87 at Pasadena. We can easily reconcile these values with a focal mechanism representing the expected overthrust of the Nazca plate by the South American continent: ($\phi = 330°$, $\delta = 11°$, $\lambda = 90°$ at 25 km depth), yielding $M_c = 8.23$–$8.37$ (PAS) and 8.20–8.33 (UPP). This mechanism is very comparable to the geometries of the nearby 1966 and 1974 shocks (ABE, 1972; DEWEY and SPENCE, 1979). The resulting moment is in good agreement with the value ($2.5 \times 10^{28}$ dyn-cm) proposed on the basis of the extent of rupture and aftershocks (KELLEHER, 1972; KANAMORI, 1977).

• *14 May 1942; Ecuador*

Records processed: PAS

We obtain $M_m = 8.00$, and assuming a focal mechanism similar to the CMT solution for the nearby 1979 event, a somewhat lower value of $M_c = 7.86$. The lower moment for this earthquake would be supported by the absence of tsunami reports in SOLOV'EV and GO (1984).

• *24 August 1942; Nazca, Peru*

Records processed: UPP, PAS

$M_m$ values are 8.15 (Love) and 8.38 (Rayleigh) at Uppsala, but only 7.55 (Rayleigh) at Pasadena, a clear result of a Rayleigh node in the azimuth of faulting for the expected underthrusting focal mechansim. In that geometry ($\phi = 325°$;

$\delta = 20°$; $\lambda = 87°$), $M_c$ values are 8.25 (Love) and 8.18 (Rayleigh) at Uppsala, and 7.97 at Pasadena. The corresponding value of the moment ($1.3 \times 10^{28}$ dyn-cm) is significantly smaller than KANAMORI's (1977) estimate ($2.7 \times 10^{28}$) from the aftershock area, although in line with the relatively small rupture zone (KELLEHER, 1972) and the absence of a teleseismic tsunami (SOLOV'EV and GO, 1984); it remains larger than BRUNE and ENGEN's (1969) estimate ($7.8 \times 10^{27}$ dyn-cm) based on the nodal stations COL and CHI. The extreme conventional magnitude reported ($M = 8.6$) illustrates the unreliability of such scales.

• *06 April 1943; Illapel, Chile*

Records processed: UPP, PAS

$M_m$ values are 7.95 (Love) and 8.02 (Rayleigh) at Uppsala, 7.77 (Love) at Pasadena. This event may have been somewhat deeper than usual, since it was felt as far away as Buenos Aires. A mechanism of pure underthrusting at a depth of 30 km ($\phi = 10°$; $\delta = 15°$; $\lambda = 90°$) gives $M_c = 8.15$ (Love) and 8.35 (Rayleigh) at Uppsala, 7.86 at Pasadena. Although these estimates are somewhat scattered, the Uppsala figures suggest a moment ($M_0 = 1.8 \times 10^{28}$), slightly less than estimated from the aftershock area (KANAMORI, 1977). The fact that the tsunami was well recorded in Japan but small in Hawaii (1 cm in Honolulu) is probably an artifact of focusing induced by lateral heterogeneity in basin bathymetry (WOODS and OKAL, 1987).

• *01 April 1946; Aleutian Islands*

Records processed: FLO, PAS, DBN, CHR, WES, UPP

The 1946 Aleutian Island earthquake, which generated the largest Pacific-wide tsunami of the 20th century, but featured only a conventional magnitude $M_s = 7.4$, remains a challenge to seismologists. BRUNE and ENGEN (1969) first remarked that 100-s spectral amplitudes were much larger than expected from the $M_s$ value. Their figure of up to 10, possibly 15, cm-s would correspond approximately to 1 to 2 times (possibly 3 to 4 times for a favorable focal mechanism) $10^{28}$ dyn-cm. KANAMORI (1972a) proposed a comparable value of $3.7 \times 10^{28}$ dyn-cm. This is still too low to account for the observed catastrophic Pacific-wide tsunami. We analyzed as many records as possible from this event (some of them graciously provided by Professors R. B. Herrmann and D. A. Wiens), in order to detect whether the ultra-long period characteristics of the source could explain the enhanced tsunami excitation, or if rupture into sedimentary material (OKAL, 1988), or underwater slumping (as suggested by KANAMORI, 1985), had to be invoked. Unfortunately, most of the available records, obtained on electromagnetic instruments whose response falls off like $\omega^3$ at very long periods, are inadequate for this purpose. We

Table A-1

*$M_m$ measurements for the Aleutian earthquake of 01 April 1946*

| Station | Code | Wavetrain | Instrument | Distance (°) | $M_m$ | Period (s) |
|---|---|---|---|---|---|---|
| De Bilt | DBN | $R_1$ | Golitsyn | 74.90 | 8.33 | 192 |
| Uppsala | UPP | $R_1$ | Wiechert | 67.57 | 8.21 | 171 |
| Uppsala | UPP | $G_1$ | Wiechert | 67.57 | 7.63 | 102 |
| Pasadena | PAS | $G_1$ | Benioff 1-90 | 37.01 | 7.75 | 128 |
| Weston | WES | $G_1$ | Benioff 1-60 | 58.81 | 8.44 | 256 |
| Weston | WES | $G_2$ | Benioff 1-60 | 301.19 | 8.09 | 142 |
| Weston | WES | $G_3$ | Benioff 1-60 | 418.81 | 8.14 | 160 |
| Christchurch | CHR | $G_1$ | Golitsyn | 98.29 | 8.19 | 110 |
| Florissant | FLO | $G_1$ | Wood-Anderson | 50.62 | 8.88 | 256 |

report in Table A-1 the individual values obtained for $M_m$, and the periods at which they were retained. There is some evidence for a general growth of the seismic moment at long periods, but this behavior is only tentative. We did not attempt to compute $M_c$ values.

• *20 December 1946; Nankaido*

Records processed: UPP, PAS

$M_m$ values are 8.63 ($G_2$) and 8.69 ($G_3$) at Pasadena, and only 8.08 (Love) and 8.49 (Rayleigh) at Uppsala. KANAMORI (1972b) gives a well-constrained first-motion focal mechanism and estimates $M_0$ at $1.5 \times 10^{28}$ dyn-cm, based on comparisons with the 1923 Kanto and 1933 Sanriku events. However, as discussed above, both of these earthquakes may be significantly larger than previously estimated, so that this figure may be underestimated. By reducing the depth of the event, it would be possible, in principle, to let the mechanism vary slightly. However, the Pasadena and Uppsala figures cannot be reconciled without changing the strike of the fault. We note that KANAMORI (1972b) mentions that his solution does not account for the full geodetic slip. We suggest that the Nankaido earthquake may have been accompanied by a slow component with a possibly rotated focal mechanism. The total moment would probably be in the vicinity of $3 \times 10^{28}$ dyn-cm.

• *24 January 1948; Panay, Philippines*

Records processed: PAS

At $M_{PAS} = 8.3$, this event has the largest reported magnitude for shallow historical earthquakes in the Philippines. The only phase available for processing is

$G_1$ at Pasadena, resulting in $M_m = 7.55$. No focal mechanism is available for this event, but it may represent eastwards subduction of the South China Sea under the western margin of the archipelago. Such an underthrusting mechanism would result in a moment of about $4 \times 10^{27}$ dyn-cm. While this figure remains an estimate, it is clear that the event is only of moderate size.

• *22 August 1949; Queen Charlotte Islands*

Records processed: UPP

We do not use Love waves, expected to be nodal at Uppsala. $M_m$ reaches 8.76 for Rayleigh waves, and in the predictable strike-slip geometry ($\phi = 331°$; $\delta = 90°$; $\lambda = 180°$), $M_c$ falls to 8.37. This still gives the event a moment nearly double BEN-MENAHEM's (1978) estimate ($M_0 = 1.15 \times 10^{28}$ dyn-cm), based on a single-frequency analysis at Pasadena. The Pasadena record was not available for study. At any rate, this event, together with the 1906 San Francisco earthquake, and the 1942 Prince Edward Fracture Zone shock (OKAL and STEIN, 1987), is one of the largest transform fault earthquakes documented.

• *15 August 1950; Assam*

Records processed: UPP, PAS

Only the NS component (mostly Love polarization) at Uppsala could be processed, resulting in $M_m = 8.96$. At Pasadena, the Benioff 1-90 records of the first passages are missing and the available strainmeter records clipped. Fortunately, 1-90 vertical records are available for $R_3$, $R_4$, and $R_5$, yielding very consistent values of $M_m = 9.03$, 8.97, and 9.02 respectively.

Some controversy still exists on the focal mechanism of this event. BEN-MENAHEM *et al.* (1974) have proposed a mostly strike-slip solution ($\phi = 334°$; $\delta = 60°$; $\lambda = 176°$), while CHEN and MOLNAR (1977) have noticed that an overthrusting mechanism ($\phi = 82°$; $\delta = 78°$; $\lambda = 90°$) provides an equally adequate fit to the polarity of $P$ waves (see their Figure 4). One problem with BEN- MENAHEM *et al.*'s (1974) solution is that it draws the null axis through a dense group of compressional readings at Japanese stations; while managing to skillfully place all of them in compressional quadrants, the authors disregard the fact that stations close to the null axis should all be very emergent, since their $P$-wave amplitudes decay like $\sin^2 \psi$, where $\psi$ is the angular distance to the axis; more importantly, the null axis is also a node of $SV$ and $SH$ radiation, and it becomes very difficult to justify the sharp $SH$ wave at Tokyo shown on their Figure 3, which they themselves describe as "conspicuous." On this basis, we prefer the CHEN and MOLNAR (1977) mechanism, for which the $M_c$ values are grouped between 9.07 and 9.23, in good agreement with these authors' estimate of $M_0 = 9.5 \times 10^{28}$ dyn-cm ($M_m^P = 8.97$).

While further work to solve the focal mechanism discrepancy would be welcome, there is no doubt that the Assam earthquake is one of the largest ever recorded.

• *02 November 1950; Banda Sea*

Records processed: PAS

This earthquake is given a large conventional magnitude ($M_{PAS} = 8.1$), second only, among Banda Sea shallow shocks, to the 1938 earthquake (see above). $M_m$ values are 7.79 (Love) and 7.51 (Rayleigh) at Pasadena. Because of the complex geometry of subduction in the Banda Sea, no attempt was made to compute $M_c$.

• *02 December 1950; Vanuatu*

Records processed: PAS

We analyze this earthquake, whose conventional magnitude ($M_{PAS} = 8.1$) is one of the largest reported in Vanuatu. $M_m$ values range from 7.71 to 7.98. Assuming the earthquake represents interplate underthrusting ($\phi = 350°$; $\delta = 28°$; $\lambda = 90°$), we obtain $M_c$ values ranging from 7.72 to 7.97. This clearly confirms that this earthquake, although large by Vanuatu standards, is not in the league of the great earthquakes occurring at subduction zones with stronger coupling as discussed for example by UYEDA and KANAMORI (1979).

• *18 November 1951; Tibet*

Records processed: PAS

We analyzed only a vertical Benioff 1-90 record of $R_1$ at Pasadena, yielding $M_m = 7.54$; $M_c = 7.43$, in good agreement with the value from CHEN and MOLNAR (1977) ($M_0 = 1.9 \times 10^{27}$ dyn-cm), whose focal geometry we used in computing $M_c$.

• *04 March 1952; Tokachi-Oki*

Records processed: UPP, PAS

This earthquake presents a challenge in that the Pasadena and Uppsala records yield very different values of $M_m$: 7.93 ($R_2$) at Pasadena vs. 9.10 ($R_1$) at Uppsala. The latter figure is probably excessive since no second passages could be observed.

ABE (1975) proposed a moment of $1.6 \times 10^{28}$ dyn-cm based on his estimate of the rupture area. He later revised this figure upwards to $2.5 \times 10^{28}$ dyn-cm, based on a tsunami magnitude $M_t = 8.4$. BRUNE and ENGEN's (1969) 100-s spectral amplitudes convert to moments as large as $3.8 \times 10^{28}$ dyn-cm. Furthermore, ABE (1975) considers the earthquake as overthrusting of the Pacific plate under Japan, while DENHAM (1977) proposes a normal faulting solution ($\phi = 132°$; $\delta = 58°$; $\lambda = -25°$) based on the large number of reported dilatations in North America. In Denham's geometry, Pasadena is nodal for Rayleigh waves, and $M_c$ reaches 8.72 at PAS, 8.93 at UPP, suggesting a moment as high as $6 \times 10^{28}$ dyn-cm. This figure may be too large, since the tsunami did not exceed 50 cm at teleseismic distances (SOLOV'EV and GO, 1984). ABE's (1975) mechanism, on the other hand, cannot reconcile $M_c$ at the two stations (7.98 at PAS; 8.97 at UPP). Further work on this event, in particular the determination of its focal mechanism, is highly desirable.

• *04 November 1952; Kamchatka*

Records processed: UPP, PAS

Results from both usable passages at UPP are $M_m = 9.30$ ($R_1$) and 9.26 ($R_2$); $M_c = 9.52$ ($R_1$) and 9.41 ($R_2$), in excellent agreement with KANAMORI's (1976) value of 9.54, obtained from a seismological investigation of Pasadena records.

However, our results from Pasadena are somewhat lower than Kanamori's ($M_m$ from 9.13 to 9.39; $M_c$ from 9.03 to 9.41). This is surprising since the Pasadena records are precisely those used in KANAMORI (1976). We interpret this discrepancy as resulting from the use of a slightly different earth model (Kanamori was using Model 5.08M as opposed to the more recent PREM model used in the $M_m$ algorithm). A slight adjustment of the focal mechanism would reconcile the UPP and PAS results around a value of $2.3 \times 10^{29}$ dyn-cm ($M_m = 9.36$), which still makes this event one of the largest ever recorded.

• *25 November 1953; Japan*

Records processed: TIN

The only record available for processing was a Benioff 1-90 (EW) at Tinemaha, California. The resulting $M_m$ (7.84) is in excellent agreement with M. Ando's estimate, quoted by ABE and KANAMORI (1980) ($M_0 = 8.9 \times 10^{27}$ or $M_m^p = 7.95$). In this region located at the intersection of three plates, large earthquakes can have many geometries; therefore, no attempt was made to compute an $M_c$ value.

• *09 March 1957 and 07 May 1986; Aleutian Is.*

Records processed: 1957: UPP, PAS, ABU, PER; 1986: UPP

The great 1957 Aleutian event remains the subject of much controversy. KANAMORI (1977) estimated its moment at $5.85 \times 10^{29}$ dyn-cm ($M_m^p = 9.77$) on the basis of the extent of its aftershocks. On the other hand RUFF *et al.* (1985) could resolve only one third of this proposed moment. While the focal mechanism of this event has not been constrained, LANE and BOYD (1990) have suggested an extremely slow rupture extending over more than 700 km, and lasting 200 s. Finally, the earthquake did create a Pacific-wide catastrophic tsunami. In this study, we compare the Uppsala records to those of the 1986 event, which occurred at a similar location, but was clearly smaller, as demonstrated by its benign tsunami.

Few well-calibrated records were available for study. Among the PAS records, only $G_2$ and $R_3$ can be studied on the 70-s strainmeter. The values obtained ($M_m = 8.11$ and 8.28, respectively) are suprisingly low. Assuming the same subduction mechanism as for the 1986 event, but a greater depth in view of the generally larger size of the earthquake, PAS is nodal for both waves, yielding $M_c = 8.49$ and 8.97, respectively. These numbers are still considerably lower than RUFF *et al.*'s (1985) estimate, let alone KANAMORI's (1977).

Only the North-South component was usable at Uppsala. $R_2$ could not be identified, suggesting an event below $10^{29}$ dyn-cm. The corresponding $M_m$ is 8.62, with $M_c$ around 8.45. An additional record was obtained at Perth, courtesy of Dr. Thomas Boyd. Results are $M_m = 8.32$ ($R_3$) and 8.16 ($R_4$); $M_c = 8.50$ and 8.36, respectively. Again, these figures are much lower than reported, but the exact calibration of the Press-Ewing instruments used at Perth at the time is in doubt (I.L. Cifuentes, pers. commun., 1991). A vertical Wiechert record at Abuyama was also used, its response interpreted thanks to UMEDA and ITO's (1987) detailed description of the microfilming project at that station. $M_m$ grows at very long periods, to a figure of 8.85 at 204.8 s. $M_c$, at 9.27 would approach RUFF *et al.*'s (1985) estimate, but the signal-to-noise ratio becomes questionable given the relatively short period of that particular instrument ($T_p = 4.7$ s).

A comparison of the records of the 1957 and 1986 events on the same instrument at Uppsala (Figure A-3) supports the conclusion that the 1957 earthquake does not seem significantly larger than the 1986 one. However, its tsunami was clearly much stronger. In view of the study by LANE and BOYD (1990), it is possible that the event was very slow, with the bulk of the moment release taking place beyond 200 s, outside the range of efficiency of the instruments available at the time. Indeed, if we attempt to push the measurement of $M_m$ beyond 300 s, a value of 9.54 is obtained at 341 s, but this value becomes very close to the noise level. Further work on that event would be highly desirable, but it would require additional data recorded on instruments with well-documented characteristics.

For the 1986 event, the apparent disparity between the Love and Rayleigh $M_m$ at Uppsala (7.83 and 8.64) is an artifact of the focal geometry. The $M_c$ values (8.09 and 8.40) agree better with the CMT one (8.02), although the Rayleigh value does remain too large.

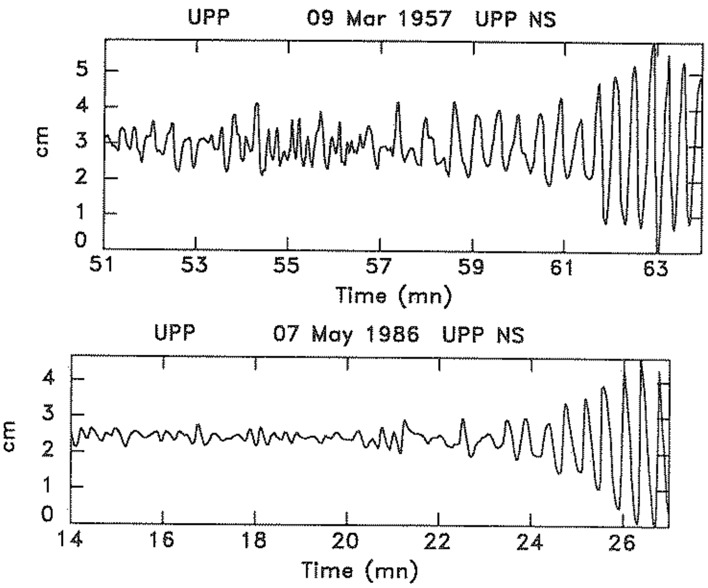


Figure A-3
Same as Figure A-2 for the 09 March 1957 and 07 May 1986 earthquakes in the Aleutians.

• *04 December 1957; Gobi-Altai, Mongolia*

Records processed: PAS, UPP

This earthquake was studied in detail by OKAL (1976), who gave a moment of $M_0 = 1.8 \times 10^{28}$ dyn-cm ($M_m^P = 8.26$). Thanks to the activation of the Press-Ewing 30-90 s, a wealth of data is available at Pasadena, up to $R_7$. $M_m$ values range from 8.39 to 8.52 with $M_c$ from 8.03 to 8.23. The fact that these values are slightly below OKAL's (1976) estimate of 8.26, is again probably due to the use of PREM in our present study, rather than 5.08 M in the older one.

Uppsala is expected to be nodal for Love waves in OKAL's (1976) geometry, but still yields $M_m = 8.02$ for both Rayleigh and Love records. A slight adjustment of the rake angle would yield $M_c$ values of 8.05 for Rayleigh and 8.4 for Love, in reasonable agreement with the published value (8.26).

• *10 July 1958; Alaska Panhandle*

Records processed: UPP, PAS

Uppsala is nodal for Love waves and Pasadena for Rayleigh waves. The Rayleigh wave at Uppsala yields $M_m = 8.46$, while the Pasadena values are much smaller: $M_m = 8.08$ for $G_2$, 7.99 for $G_4$. The focal mechanism of the event was given by STAUDER (1960) as $\phi = 338°$; $\delta = 72°$; $\lambda = 172°$. For this mechanism, $M_c$ values are 8.10 at Uppsala, 7.67 ($G_2$) and 7.57 ($G_4$) at Pasadena, all significantly smaller than proposed by KANAMORI (1977) on the basis of the aftershock area ($2.9 \times 10^{28}$ dyn-cm; $M_m^p = 8.46$). However, BEN-MENAHEM (1978) gives a "potency" of 16.7 km$^3$, equivalent, for this crustal earthquake, to a moment of $5.3 \times 10^{27}$ dyn-cm ($M_m^p = 7.72$). The latter figure would also agree with $M_0 = \mu \cdot S \cdot \Delta u$, as inferred from the extent of rupture (250 km); the presumed width of the fault (24 km); and an *average* slip of 3 m (TOCHER, 1960; BEN-MENAHEM and TOKSÖZ, 1963). While $M_c$ at Uppsala remains large, we think that the event is certainly much smaller than listed by KANAMORI (1977).

• *06 November 1958; Kurile Islands*

Records processed: UPP, PAS

$M_m$ values at Uppsala are 8.73 ($R_1$), 8.78 ($R_2$) and 8.37 ($G_2$). Pasadena is expected to be nodal for Rayleigh waves, so we used only $R_2$, and a set of Love waves, recorded up to $G_5$ on the 70-s strainmeter. The corresponding $M_m$ values are in general lower than at Uppsala (7.92–8.34).

The mechanism of the event is given by FUKAO and FURUMOTO (1979) as $\phi = 45°$; $\delta = 60°$; $\lambda = 90°$ at a depth of 60 km. This yields values of $M_c$ at Uppsala (8.43, 8.52, 8.78) generally consistent with the published values ($M_m^p = 8.64$), but values computed from the 70-s strainmeter remain slightly deficient (ranging from 8.02 to 8.60).

• *04 May 1959; Kamchatka*

Records processed: PAS, RVR, UPP

$M_m$ values are 8.35 (UPP, Rayleigh), 8.09 (PAS, Rayleigh), 7.49 (PAS, Love) and 7.83 (RVR, Rayleigh). These values can be reconciled in the normal faulting geometry of DENHAM (1977) ($\phi = 105°$; $\delta = 48°$; $\lambda = 335°$), yielding $M_c = 8.08$, 8.09, 7.97 and 7.77, respectively. In that geometry, UPP is nodal for Love waves. We estimate the earthquake's moment at $9.5 \times 10^{27}$ dyn-cm, significantly less than suggested by KANAMORI (1977) from its aftershock area ($2.6 \times 10^{28}$ dyn-cm).

• *22 May 1960; Chile*

The study of this earthquake is crucial to the success of the mantle magnitude $M_m$: as discussed in our original paper (OKAL and TALANDIER, 1989), one of the goals of the whole endeavor is to alleviate the saturation which plagues any magnitude (and especially $M_s$) measured at a constant period. It is therefore important to test the $M_m$ algorithm on the largest event ever recorded. With this in

Table A-2

*$M_m$ and $M_c$ measurements for the great Chilean earthquake of 22 May 1960*

| Station | Code | Wavetrain | Distance (°) | $M_m$ | Period (s) | $M_c$ |
|---|---|---|---|---|---|---|
| Palisades | PAL | $R_4$ | 640.68 | 9.86 | 233 | 10.19 |
| Palisades | PAL | $R_6$ | 1000.68 | 9.79 | 233 | 10.13 |
| Palisades | PAL | $R_8$ | 1360.68 | 10.16 | 122 | 10.68 |
| Seven Falls | SFA | $G_4$ | 634.54 | 10.29 | 142 | 10.77 |
| Resolute | RES | $G_6$ | 966.07 | 10.13 | 142 | 10.45 |
| Resolute | RES | $G_8$ | 1326.07 | 10.24 | 213 | 10.57 |
| Resolute | RES | $G_{10}$ | 1686.07 | 10.17 | 183 | 10.50 |
| Resolute (Z) | RES | $R_{10}$ | 1686.07 | 10.18 | 171 | 10.55 |
| Resolute (NS) | RES | $\mathrm{R}_{10}$ | 1686.07 | 10.57 | 213 | 10.88 |
| Uppsala | UPP | $R_2$ | 235.75 | 9.67 | 171 | 9.91 |
| Uppsala | UPP | $R_4$ | 595.75 | 9.74 | 228 | 9.93 |
| Uppsala | UPP | $G_2$ | 235.75 | 9.64 | 256 | 9.81 |
| Pasadena | PAS | $G_2$ | 276.53 | 9.77 | 284 | 10.02 |
| Pasadena | PAS | $G_4$ | 636.53 | 9.78 | 284 | 10.03 |
| Pasadena | PAS | $R_2$ | 276.53 | 9.99 | 233 | 10.16 |
| Pasadena | PAS | $G_2$ | 636.53 | 10.05 | 284 | 10.20 |
| La Folinière | FLN | $R_6$ | 971.62 | 10.31 | 223 | 10.54 |
| La Folinière | FLN | $G_6$ | 971.62 | 10.33 | 205 | 10.50 |

mind, we collected a number of records from the Chilean event, and processed them for $M_m$. Results are given in Table A-2. The average values of $M_m$ is 10.04, and of $M_c$ 10.32, corresponding to the often quoted moment of $2-3 \times 10^{30}$ dyn-cm, as the "main shock" part of a total moment release reaching up to $5.5 \times 10^{30}$ dyn-cm (KANAMORI and CIPAR, 1974; KANAMORI and ANDERSON, 1975; CIFUENTES and SILVER, 1989). It is doubtful, however, that the remainder of the moment, released over a total time of up to 25 mn, could be identified on the basis of a single wavetrain. As discussed in the main text, and elsewhere (OKAL and TALANDIER, 1991), the gigantic character of the event is perfectly retrieved by the $M_m$ computations, thereby justifying the whole approach.

• *20 November 1960; Peru*

Records processed: DBN, WES, RES, PAS, BKS

For this "tsunami earthquake" which PELAYO and WIENS (1990) have shown to have a slow character, we processed a large number of records, kindly provided by Professor D. A. Wiens. Because of an unfavorable focal mechanism, most American stations are nodal for both Rayleigh and Love waves, and the $M_m$ values are deficient: $M_m = 6.85$ ($R_2$) and 6.98 ($G_2$) at Pasadena, 6.72 ($R_2$) at Resolute, 6.78 ($R_2$) at Berkeley. Only $G_2$ at Berkeley (7.21) and $G_1$ at De Bilt (7.43) approach the value published by PELAYO and WIENS (1990) ($M_m^p = 7.53$). On the other hand values of $M_c$ range from 7.16 (Berkeley $R_2$) to 7.91 (De Bilt $G_1$); furthermore, a detailed look at the spectral amplitudes confirm that $M_m$ and $M_c$ values grow substantially with period, the (maximum) value retained by the program being always at the longer period end of the spectrum. PELAYO and WIENS' (1990) analysis is therefore fully upheld in our approach.

• *13 October 1963; Kuriles*

Records processed: UPP

$M_m$ values for this event are 8.31 (Love) and 8.80 (Rayleigh). $M_c$ values using KANAMORI's (1970a) focal solution remain different, 8.41 (Love) and 8.63 (Rayleigh). Adjusting the slip angle to 80° would reconcile them around 8.68, in good agreement with the published values ($M_m = 8.83$; KANAMORI, 1977).

• *28 March 1964; Alaska*

Records processed: UPP

Only EW records (Love polarization) at Uppsala could be processed. The earthquake is clearly gigantic, since $G_2$ and $G_4$ wavetrains are well recorded. The resulting values are $M_m = 9.57$ and 9.45, respectively. Using KANAMORI's (1970b) mechanism, $M_c$ values are somewhat deficient, reaching only 9.59 and 9.48, respectively. This is probably due to the poor performance of the Wiechert instrument at ultra-long periods. Nevertheless, these figures are among the highest measured at UPP, confirming that $M_m$ has the potential to recognize truly gigantic events. OKAL and TALANDIER (1991) have shown that the moment of the event can also be correctly retrieved from individual WWSSN stations.

• *11 August 1969; Kuriles*

Records processed: UPP

Rotated records at Uppsala yield $M_m = 8.31$ (Love) and 8.62 (Rayleigh). In the geometry of ABE (1973) ($\phi = 220°$, $\delta = 16°$; $\lambda = 90°$), the $M_c$ values (in both cases, 8.50) are in good agreement with his estimate of $M_0$ ($2.2 \times 10^{28}$ dyn-cm or $M_m^p = 8.34$).

• *19 August 1977; Indonesia*

Records processed: UPP

Uppsala is expected to be nodal for Love waves. Rayleigh waves yield a very strong $M_m = 9.21$, which in the CMT geometry corresponds to $M_c = 8.82$, significantly larger than the published value (8.56).

• *03 March 1985; Chile*

UPP records yield $M_m = 7.82$ (Rayleigh) and 7.96 (Love). In the CMT geometry, we obtain $M_c = 7.90$ and 7.88, respectively, in good agreement with the published value ($M_m^p = 8.01$).

• *19 September 1985; Mexico*

Records processed: UPP

Uppsala is expected to be nodal for Love waves; $M_m$ for Rayleigh waves is 8.50, and in the CMT geometry, $M_c = 8.26$, somewhat larger than, but still agreeing well with, the published value ($M_0 = 1.1 \times 10^{28}$ dyn-cm).

• *23 May 1989; Macquarie Ridge*

Records processed: UPP

UPP is expected to be nodal for Rayleigh waves; only Love waves are studied, with $M_m$ very strong (8.76), but $M_c$ (8.29) more in agreement with the CMT value ($M_m^p = 8.15$).

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