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Coseismic Excitation of the Earth's Polar Motion

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Abstract. Apart from the "shaking" near the epicenter that is the earthquake, a seismic event creates a permanent field of dislocation in the entire Earth. This redistribution of mass changes (slightly) the Earth's inertia tensor; and the Earth's rotation will change in accordance with the conservation of angular momentum. Similar to this seismic excitation of Earth rotation variations, the same mass redistribution causes (slight) changes in the Earth's gravitational field expressible in terms of changes in the Stokes coefficients of its harmonic expansion. In this paper, we give a historical background of the subject and discuss the related physics. We then compute the geodynamic effects caused by earthquakes using Chao and Gross' (1987) formulas based on Gilbert's (1970) normal-mode summation scheme. The effects are computed using the centroid moment tensor (CMT) solutions for 15,814 major earthquakes from Jan., 1977, through Feb., 1999, as provided in the Harvard CMT catalog. The computational results update those of Chao and Gross (1987) and Chao *et al.* (1996), further strengthening their findings and conclusions: (i) the strong tendency for earthquakes to make the Earth rounder and more compact (however slightly) continues; (ii) so does the trend in the seismic "nudging" of the rotation pole toward the general direction of $\sim 140°$E, roughly opposite to that of the observed polar drift, but two orders of magnitude smaller in drift speed.

1. Introduction

The Earth's polar motion has been observed for over a hundred years, initially by astrometric and in modern times by space geodetic techniques. The polar motion *excitation* function derived from these observations shows a generally broad-band structure, but with certain prominent signals superimposed: a more-or-less secular drift (*e.g.*, Gambis, this issue) largely attributed to the present-day post-glacial rebound, a ~30-year Markowitz wobble whose origin remains mysterious (*e.g.*, Poma, this issue), and the very notable annual wobble of obvious meteorological origin (*e.g.*, Salstein, this issue).

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In addition, the observed polar motion has a strong Chandler wobble component with a time-varying amplitude comparable to that of the annual wobble. Although the Chandler wobble is a natural free mode, it still needs continual excitation to maintain its observed amplitude. Despite many studies, the Chandler wobble's excitation sources have remained elusive to date, although atmospheric angular momentum variations, perhaps together with oceanic variations, may prove to be largely responsible for its excitation.

Historically, another notable candidate excitation source for the Chandler wobble was seismic dislocation; a first proposal was made as early as Milne (1907), soon after the annual and Chandler wobbles were identified (Chandler, 1892; also Dick, this issue). Cecchini (1928) later noted some correlation between the large polar motion and the high seismicity during 1900–1908. Similar correlations have been alluded to in subsequent reports, such as Runcorn (1970), Pines and Shaham (1973), Press and Briggs (1975), Kanamori (1976).

However, to establish an unequivocal relationship between seismic excitation and the observed polar motion, one needs to be able to compute quantitatively how much an earthquake can excite polar motion by altering the Earth's inertia tensor. In their milestone geophysical monograph, Munk and MacDonald (1960) briefly treated the problem. They used a simplistic local block-dislocation model for an earthquake, and quickly dismissed the importance of earthquakes in polar motion excitation, even for the largest earthquakes.

Then came the great 1964 Alaskan earthquake, which provided new, fundamental insight into the displacement field of an earthquake: Based on a strainmeter record in Hawaii, Press (1965) announced that a static displacement was recorded at teleseismic distances several thousand kilometers from the epicenter. That prompted a series of investigations of seismic excitation of polar motion: Mansinha and Smylie (1967), Smylie and Mansinha (1968; 1971), Mansinha *et al.* (1970; 1979), Ben-Menahem and Israel (1970), Israel *et al.* (1973), Israel and Ben-Menahem (1975), Rice and Chinnery (1972), Dahlen (1971; 1973), O'Connell and Dziewonski (1976), Smith (1977). Unfortunately the search for signatures left by large earthquakes (*e.g.*, the great 1960 Chilean event and the 1964 Alaskan event) in polar motion was essentially inconclusive: the quality of the polar motion data at the time was insufficient for that purpose both in accuracy and temporal resolution.

A revival of interest in the problem appeared during the latter half of the 1980s, largely because of advances in polar motion measurement techniques, but also owing to the availability of the Harvard centroid moment tensor (CMT) catalog of all major earthquakes (see below). Using Dahlen's (1973) formula on the thousands of earthquakes listed in the catalog, Souriau and Cazenave (1985) and Gross (1986) computed time series of seismic excitation of polar motion. They concluded that the earthquakes since 1977 were simply too small to produce any appreciable signature in polar motion, with the cumulative seismic excitation power being orders of magnitude smaller than that observed.

The next development was by Chao and Gross (1987) who again computed the seismic excitation of polar motion for all events listed in the CMT catalog, but using the normal-mode summation scheme of Gilbert (1970). Their method has since remained a most efficient way of computing the seismic excitation of not only polar motion, but also of other important geodynamic parameters such as gravitational field changes. Furthermore, Chao and Gross (1995) and Chao *et al.* (1995) extended the formulation to compute earthquake-induced changes in rotational energy and gravitational energy, respectively. These papers and later Chao *et al.* (1996) have updated, and in fact strengthened, the results of Chao and Gross (1987) who found many earthquake-induced phenomena having intriguing geodynamical implications (see Section 3).

The present paper will compute and discuss the seismic excitation of a host of geodynamic parameters as in Chao and Gross (1987), but much of the discussion will focus on that of polar motion, as in Chao *et al.* (1996). However, to avoid duplicating the latter (as the major findings and conclusions remain the same), we focus here on the historical development of the subject (given above) and on its physics, although often in a qualitative manner.

It should be stressed that the studies referenced above as well as the present one only pertain to the coseismic effects, that is, to the effect due to the elastic dislocation that happens within, say, an hour following the initial rupture of the fault. The inelastic pre- or post-seismic movements that are often associated with large earthquakes on timescales of months to years are outside our present scope. These effects typically augment the coseismic ones by a factor depending on the source mechanism and mantle rheology. The reader is referred to, *e.g.*, Dragoni *et al.* (1983), Sabadini *et al.* (1984), and Soldati and Spada (1999) for some numerical modeling of these inelastic effects.

In passing, we mention an effect in the opposite sense that has been proposed in the past (*e.g.*, Lambert, 1925) — the possible triggering of earthquakes by the centrifugal "pole tide" potential generated in the solid Earth by polar motion. This effect is physically analogous to tidal triggering of earthquakes; however, statistical studies have so far failed to conclusively detect the effect (*e.g.*, Chao and Liu, 2000).

2. Formulation and Data

The physics of the coseismic excitation of geodynamic effects is straightforward. An earthquake is an abrupt dislocation, or faulting, along a fault plane at the hypocenter located somewhere in the solid Earth. In addition to the seismic oscillations (the "earthquake") which eventually die away, this dislocation creates a static, global displacement field $u(r)$ in the Earth, as it would in any strained elastic body.

Here we should be more explicit as to the definition of $u(r)$. In a non-rotating Earth there is no ambiguity as long as one recognizes that the center of mass stays unchanged because earthquakes are internal processes which conserve the Earth's linear momentum. For a rotating Earth, any given mass point experiences a u relative to the unperturbed position r if not for the earthquake, reckoned in the rotating terrestrial reference frame centered at the center of mass of the Earth.

Various approaches (see references cited above) have been taken to compute the static displacement field given an earthquake source mechanism. Here we use Gilbert's (1970) normal-mode summation approach:

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$$u(r) = \sum_k \omega_k^{-2} u_k(r) M : E_k * (r_f), \tag{1}$$

where M and r_f are respectively the seismic moment tensor and the hypocenter location of the earthquake that generates the dislocation. Collectively called the "eigen-elements", ω_k, $u_k(r)$, and E_k, are the eigen-frequency, the eigen-function, and the elastic strain tensor, respectively, of the kth normal mode belonging to the elastic Earth. The symbol : is the second-degree tensor dot product; * denotes the complex conjugate.

To compute the displacement $u(r)$ using Equation (1), two sets of quantities are required: (i) The eigen-elements of the Earth's normal modes, which have long been computed by seismologists; we adopt those computed for the spherically symmetric Earth Model 1066B of Gilbert and Dziewonski (1975) (see below). (ii) The seismic moment tensor of the earthquake, which seismologists routinely determine by seismic data inversion. In particular, the Harvard CMT catalog lists the moment tensor solutions of all major earthquakes above magnitude 5 (and sometimes magnitude 4) that have occurred worldwide since 1977 (*e.g.*, Dziewonski and Woodhouse, 1983), including origin times and hypocenter locations. The CMT solution contains the complete magnitude and source mechanism information that can be inverted from global teleseismic data (including the earthquake's strike, dip, and slip angles). A total of 15,814 events were available for our study, spanning Jan. 1977 through Feb. 1999.

The earthquake-induced mass displacement field in turn affects a host of geodynamic parameters. The relationship between the displacement field $u(r)$ and some geodynamic parameter F generally takes the form of a global integral:

$$\Delta F = \int_{Earth} u(r) \cdot K(r) dV, \tag{2}$$

where $K(r)$ is the kernel, or weighting function, specific to F (for complete details see Chao and Gross, 1987).

For example, the earthquake-induced mass redistribution alters the Earth's gravitational field according to Newton's gravitational law. The gravitational field is customarily expressed in terms of the Stokes coefficients of its spherical harmonic expansion. Hence, the gravitational change can naturally be expressed as changes in the Stokes coefficients. Equation (2) gives the earthquake-induced change of a particular Stokes coefficient when evaluated with a kernel function proportional to the spherical harmonic of the same degree and order as the Stokes coefficient in question. There are of course infinitely many Stokes coefficients; here we only compute those of lowest-degrees that are of interest (for example, the second degree zonal Stokes coefficient corresponding to the Earth's dynamic oblateness J_2).

Similarly, with proper kernels the integral (2) can be used to compute changes in the 6 elements of the Earth's inertia tensor (see Chao and Gross, 1987). Conservation of angular momentum then dictates that the Earth's rotation will change accordingly, analogous to a spinning skater changing his rotational speed by changing the position of his arms relative to his body. In particular, the polar motion excitation function χ can be computed using (2) with a kernel proportional to the degree-2 order-1 spherical harmonics (see Section 3 below for more discussions). χ in turn is related to the "reported" polar motion p (see below) by convolution (Munk and MacDonald, 1960):

$$p \equiv p_x + \imath p_y = exp(i\sigma t)[p_0 - \imath\sigma \int_t \chi(\tau)exp(\imath\sigma\tau)d\tau], \quad (3)$$

or equivalently through the deconvolution:

$$\chi \equiv \chi_x + \imath\chi_y = p - dp/dt/(i\sigma), \quad (4)$$

where $i = \sqrt{-1}$, σ is the complex-valued frequency of the Chandler wobble, p_0 is the initial condition, and the terrestrial coordinate system is such that the x-axis points to the Greenwich Meridian while the y-axis points to the 90°E Longitude. Here we adopt the observed value for σ estimated by Furuya and Chao (1996), corresponding to a Chandler period of 433.7 days and a Chandler Q of 50.

Incidentally, many geophysical processes also produce "geocenter motion," which is the relative motion between the center of mass of the solid Earth and that of the whole Earth system (including its fluid envelopes). However, earthquakes do not cause geocenter motion because, as stated above, earthquakes are internal to the solid Earth and hence do not change the center of mass of the solid Earth nor that of the Earth system.

3. Physical and Numerical Considerations

The coseismic displacement amplitude is largest near the rupturing fault and decreases rapidly away from the fault. However, the amount of mass being displaced grows as the cube of the distance from the fault. This is the reason why a global integration is necessary to obtain quantitative results, and why the above-mentioned simplistic localized estimate by Munk and MacDonald (1960) proved to be inaccurate.

In the spherically symmetric Earth approximation (which is sufficient for our present application), there are two types of normal modes: spheroidal and toroidal. Only spheroidal modes contribute to our calculations because toroidal modes do not involve mass density variations.

The temporal history of a single earthquake faulting, and hence its coseismic effect, is modeled here as a step function $H(t)$. This is justifiable because of the short timescales of interest here — measurement intervals and the natural resonance period for the polar motion (*i.e.* the Chandler period of 14 months) are much longer than the coseismic timescale which is no longer than an hour. The cumulative (temporal) effect of multiple earthquakes is simply obtained by a superposition of the individual step functions accounting for differences in origin time t_j and magnitude (with the proper positive or negative sign):

$$\Delta F(t) = \sum_j \Delta F_j H(t - t_j) \quad (5)$$

where the summation is over all earthquakes $j = 1, 2, 3, ..., N$. In our case, N is the 15,814 earthquakes listed in the Harvard CMT catalog.

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Earthquakes of smaller size (smaller than the magnitude threshold of the CMT catalog) can be safely neglected in our study. Although much more numerous, their geodynamic effects decrease exponentially with decreasing seismic magnitude much faster than their numbers increase.

The normal-mode summation scheme has certain fundamental physical and computational advantages. The existence and reality of the normal mode eigen-elements have been observed and tested empirically. More importantly in the present context, the eigen-elements already account for not only the elastic and gravitational forces, but also the appropriate boundary conditions at the physical boundaries in the Earth, so none of these complications need be taken into explicit consideration in the formulation. Numerically the computation is very efficient: the mode summation converges to the final results rather rapidly — typically to within a few percent after summing just several of the lowest-degree (spheroidal overtone) modes; the number of modes required is of course progressively greater for higher-degree Stokes coefficients. In any event, we use all the spheroidal overtone modes that are available to us, namely those having eigen-periods longer than 45 s (typically 60 modes).

The polar motion excitation function χ that we compute (see Section 2) is often referred to as the effective equatorial angular momentum (*cf.* Barnes *et al.*, 1983). χ consists of two terms: the "mass" term and the "motion" term (Munk and MacDonald, 1960). What we compute is actually the mass term. The motion term in the present case comes from the relative motion of the mass during the faulting, and has been shown to be negligible by Chao (1984) (the ultimate physical reason is the near-spherical configuration of the Earth). For gravitational changes (and geocenter motions for that matter) only mass redistribution matters.

The χ function is different from the geophysical excitation referred to as the Ψ function by Munk and MacDonald (1960). The difference is that while χ contains the mass and motion terms (but see the above paragraph), Ψ contains two additional terms involving the time rate-of-change of these two terms. Gross (1992) has shown that it is χ that is directly related by Equations (3) and (4) to the polar motion measurements p reported by astrometric and space geodesy techniques.

Seismic excitation of changes in the Earth's rotational speed is generally small because the inertia to overcome in changing the length-of-day (LOD) is the mantle's axial moment of inertia C_m (that is, for LOD changes the kernel in equation 2 is inversely proportional to C_m). On the other hand, the excitation of the polar motion is relatively "easier" as the inertia to overcome is, and the kernel for polar motion excitation is inversely proportional to, the difference between the axial and equatorial moments of inertia, $C_m - A_m$, which is about 300 times smaller than C_m or A_m.

Representing the geophysical causes that sustain polar motion through time, the excitation function χ obtained by deconvolving the observed polar motion p (see equation 4) is generally broad-band, with a few prominent signals mentioned in Section 1 superimposed. The χ function is not, and need not be, strong in power at the Chandler frequency. The reason why the observed Chandler wobble is particularly large is simply because the Chandler wobble is a natural resonance of the Earth's rotation: Any geophysical excitation that contains power in the

Chandler frequency band excites the Chandler wobble causing it to be greatly magnified.

For polar motion, the earthquake-induced excitation function χ is assumed to be a step function, or "jump" in time (Equation 5). However, being the convolution of the excitation function with the Chandler resonance (Equation 4), the signal in the observed p is no longer a "jump," but rather a more obscure "kink," or change in the direction of motion.

4. Conclusions

We recapitulate the major findings of geophysical significance by Chao and Gross (1987) in the present context:

1. Individual geodynamic signatures in either the Earth's rotation or gravitational field due to even the largest earthquakes that occurred during our studied period were below present-day detection thresholds. These earthquake-induced signatures are in general two orders of magnitude smaller than the observed fluctuations, which are known to be primarily caused by mass transports of other geophysical processes.

2. The collective effects of all earthquakes greater than magnitude 5 in the last two decades have an extremely strong statistical tendency with time. The parameters that show the strongest non-randomness are the dynamic oblateness J_2, the total moment of inertia (the trace of the inertia tensor), the length of day, the sum of the two equatorial principal moments of inertia, and the difference J_{22} between the two equatorial principal moments of inertia. Their time series all exhibit a strong decrease with time, indicating the tendency of earthquakes to make the Earth rounder and more compact. No such tendency is evident for higher harmonics of the gravitational field changes caused by earthquakes (*e.g.* J_3, J_4, J_5).

3. A similar strong tendency is seen in the polar motion excitation: earthquakes cumulatively are trying to "nudge" the Earth rotation pole towards $\sim$ 140°E, roughly opposite to that of the observed polar drift. However, the speed of this earthquake-induced polar drift in the last two decades is two orders of magnitude smaller than that observed.

Figure 1 shows the time series of the specific geodynamic parameters mentioned in paragraph (2) above. The decreasing trend is rather obvious; their extremely strong statistical tendencies were verified by conducting χ^2 and Wilcoxon tests as in Chao and Gross (1987). Now based on more than 7 times as many earthquakes, our present results are similar to, and in fact strengthen, the results of Chao and Gross (1987).

Figure 2 shows the cumulative χ time series for the two components (equation 4) of the coseismic excitation of polar motion for the studied period 1977.0 – 1999.2. The thick (jagged) curve in Figure 3 shows the corresponding polar drift plotted on the Earth's surface (near the geographical North Pole). The excitation curve tracks the center of the circular curve, which is the corresponding earthquake-induced polar motion p obtained by evaluating the convolution

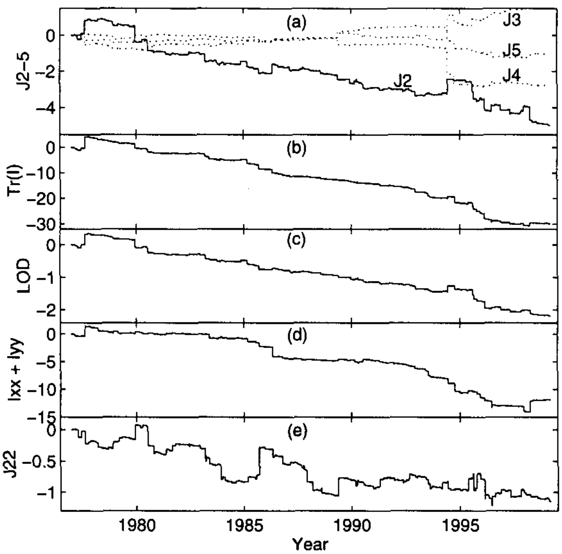

Figure 1. Series of computed cumulative changes induced by the 15,814 major earthquakes that occurred during 1977.0–1999.2 for: (a) dynamic oblateness J_2 (and J_3, J_4, J_5 as the dashed lines); (b) the total moment of inertia; (c) the length of day; (d) the sum of the two equatorial principal moments of inertia; and (e) the difference J_{22} between the two equatorial principal moments of inertia. The time series all exhibit a strong decrease with time, indicating the tendency of earthquakes to make the Earth rounder and more compact.

equation (3) and letting $p_0 = 0$. The entire earthquake-induced polar motion in Figure 3 is only about 1 milliarcsecond (mas) in amplitude. It would be completely buried in the observed polar motion which is 2–3 orders of magnitude larger.

Thus, it is clear that our updated result for χ greatly strengthens the earlier finding summarized in paragraph (3) above. The average earthquake-induced polar drift during the studied period amounts to about 0.07 mas/year (approximately 2 mm/year) in the direction of $\sim$ 140°E. The observed average polar drift during the past 2 decades is about 3 mas/year (approximately 10 cm/year) towards the direction of $\sim$ −80°E, roughly opposite to that of the earthquake-induced drift.

These findings are intriguing in terms of long-term geophysical processes occurring in the Earth's mantle. Why do earthquakes strive to make the Earth more compact and rounder in all directions? Why do earthquakes prefer a certain direction in which to nudge the Earth's rotation pole? Spada (1997) has provided a kinematic explanation for the latter based on the geographical distribution of the earthquakes and their dominant source mechanism types. However, many dynamic questions await to be answered. In any case, earthquakes

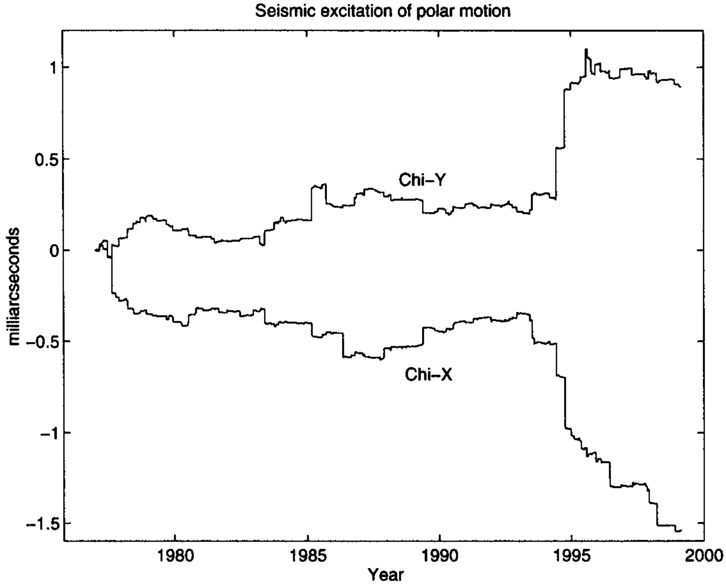

Figure 2. Same as Figure 1, but for the two components of the polar motion excitation.

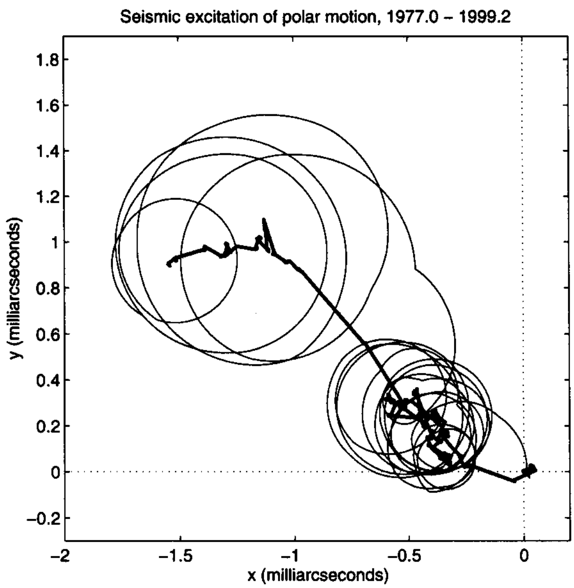

Figure 3. The thick (jagged) curve is the series of coseismic excitation of polar motion of Figure 2, or the equivalent polar drift, plotted on the Earth's surface, where 1 milliarcsecond corresponds to about 3 cm at the Earth's surface. The thin circular curve is the corresponding polar motion p generated by these excitations.

represent episodic motions of the plates at the plate boundaries. Their source mechanisms are determined by plate tectonics which in turn is a surface manifestation of mantle convection which acts on geological timescales of millions of years. It seems reasonable to conclude that our findings, although specific to the short two decades of our study, should be indicative of the long-term statistical trend on geological timescales. It would be interesting to examine the long-term implications of these geodynamic changes in terms of the Earth's rotation and gravitational field.

Finally, let us examine the detectability of the coseismic excitation of polar motion for an individual earthquake in relation to the present-day polar-motion measurement thresholds. Certain factors conspire to make such a detection difficult. First of all, the earthquake has to be very large in order to leave a signature above the noise level. As pointed out by Chao *et al.* (1996), an earthquake of scalar seismic moment $M = 10^{22} Nm$ would cause a polar motion excitation on the order of 1 mas. The 1960 Chilean event, at $M = 27 \times 10^{22} Nm$, is by-far the largest earthquake ever recorded. For length-of-day (LOD), it would only have produced an 8 microsecond coseismic decrease in LOD, an effect hardly discernible even with today's best measurements. However, it should have left a coseismic kink in the polar motion caused by the 23 mas of polar motion excitation, barely discernible in measurements taken back in 1960 but certainly very noticeable if it were to happen today (see below). The second largest earthquake on record, the Alaska event of 1964 with $M = 7.5 \times 10^{22} Nm$, should have produced a coseismic increase in J_2 by 5.3×10^{-11}, which would take post-glacial rebound two years to "iron out", but which is still an order of magnitude smaller than the observed short-term fluctuations in J_2 known to be largely due to atmospheric mass transport.

The most accurate polar motion measurements that can be achieved today during intensive VLBI (Very Long Baseline Interferometry) measurement campaigns have formal errors of about 0.2 mas (Clark *et al.*, 1998). In Figure 3, the largest kink in the pole path (see Section 3) is due to the 1994 Kuril Island event, which had a moment of $M = 0.3 \times 10^{22} Nm$ (corresponding to a moment magnitude of 8.3) (Chao *et al.*, 1996). The centers of the circular polar motion before and after the kink delineates the excitation vector due to this event, which with an amplitude of about 0.4 mas is barely above the noise level.

Secondly, the duration of the coseismic excitation is rather short, no longer than an hour (although the near-field post-seismic deformation can continue for some time). To capture the event the polar-motion measurement must be on-going at the time and the data must have sufficient temporal resolution, for example at hourly intervals. These conditions can be met by the VLBI CORE (Continuous Observation of the Rotation of the Earth) project now under implementation in phases (Clark *et al.*, 1998). Furthermore, the GPS (Global Positioning System) technique is now routinely yielding sub-daily polar motion measurements (Rothacher *et al.*, 1997), with precision approaching that of VLBI.

A third factor acting against a clear detection is a more fundamental one pertaining to geophysics. The observed polar motion excitation (or any other geodynamic effect for that matter) results from the combination of all geophysical excitation sources, many of which are orders of magnitude larger than, and hence easily obscure, the coseismic ones under study here (*e.g.*, Chao, 1994). These sources, however, often have rather different temporal characteristics than do earthquakes. Thus, under the most favorable conditions, an earthquake several times more energetic than the Kuril Island event (let alone one comparable to the above-mentioned Chilean and Alaskan events of the 1960s) can be expected to leave detectable signatures. Such a detection would be a significant, and long overdue, geophysical event.

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