SEISMIC EXCITATION OF THE POLAR MOTION, 1977-1993

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1. Introduction

The Earth's rotation varies slightly with time. The 3-D Earth rotation variation can be conveniently separated into two components: (i) The 1-D variation in the spin rate, often expressed in terms of the length-of-day variation. (ii) The 2-D variation in the rotational axis orientation, generically called the nutation when viewed from the inertial reference frame and the polar motion as seen in the terrestrial reference frame.

There are two dynamic types of Earth rotation variations (Munk and MacDonald, 1960): (i) The "astronomical variations" due to the external luni-solar tidal torque that changes the Earth's angular momentum. Well-known examples include the tidal braking that causes the Earth's spin to slow down over geological times, and the astronomical precession and nutation caused by torques exerted on the Earth's equatorial bulge. (ii) The "geophysical variations" caused by large-scale mass movement of internal geophysical processes under the conservation of angular momentum. These include tidal deformation in the solid Earth and oceans, atmospheric fluctuations, hydrological variations, ocean currents, earthquake dislocations, post-glacial rebound, mantle tectonic movement, and core activities.

The present paper deals with the polar motion and earthquakes. Possible interactions between seismicity and Earth rotation have been under consideration since the discovery of the polar motion a hundred years ago (e. g., Lambeck, 1980). On one hand, the departure of the Earth's rotation from a uniform one would give rise to a time-varying stress field inside the Earth, which in turn may affect the triggering process of earthquakes. This effect with respect to the polar motion is in general an order of magnitude smaller than its tidal counter part, but having a longer time scale on the order of a year. Chao and Iz (1992) conducted a statistical test in an attempt to correlate the occurrence of earthquake with polar motion based on nearly 10000 earthquakes that occurred during 1977-1991; but only a weak correlation was found. A more definitive conclusion awaits further studies taking into account the tensorial nature of the seismic source mechanism and the stress field induced by polar motion.

On the other hand, the mass redistribution in the Earth as a result of an cart. hquake faulting changes the Earth's inertia tensor, and hence its rotation. Early simplistic earthquake faulting models considering only regional dislocations greatly underestimated the effect (Munk and MacDonald, 1960). After the 1964 Alaska earthquake it was recognized that the seismic dislocation, although decreasing with focal distance, remains non-zero even at teleseismic distances away from the fault (Press, 1965). When integrated globally (see equation 2 below) considering the size and source mechanism of the earthquake this displacement field can give finite effect (Mansinha and Smylie, 1967). Symlie and Mansinha (1971), Dahlen (1971; 1973), Rice and Chinnery (1972), and O'Connell and Dziewonski (1976) calculated the seismic excitation of polar motion based on realistic models. It was found that the great 1960 Chile and 1964 Alaska earthquakes should have produced significant changes in the polar motion (see below). Unfortunately the accuracy of the polar motion record at that time was insufficient. to yield any conclusive detection.

In the last two decades, the measurement of Earth rotation has been progressively improved by two orders of magnitude in both accuracy and temporal resolution (e. g., Eubanks, 1993, for a review). This is achieved through the advances in modern space geodetic techniques, primarily Very-Long-Baseline Interferometry (VLBI), Satellite Laser Ranging (SLR), and Global Positioning System (GPS). In fact, these techniques have measured the contemporary relative tectonic plate motions to within a few mm/year, and revealed co-seismic, near-field displacements caused by recent earthquakes. The SLR technique has also greatly benefited the determination of the Earth's global gravitational field and detected slight temporal variations in the low-degree field. In the area of Earth rotation measurement, the current accuracy is estimated to be within 1 milliarcsecond (1 mas $= 4.85 \times 10^{-9}$ radian, corresponding to about 3 cm of distance on the surface of the Earth), while the formal errors arc as low as 200 μas. The routine temporal resolution is typically a few days but can now reach one hour during intensive campaigns.

Taking advantage of these modern data as well as the seismic centroid moment tensor solutions made available in the Harvard CMT catalog (e. g., Dziewonski ct al., 1993), Souriau and Cazenave (1985) and Gross (1 986) calculated the seismic excitation of polar motion using Dahlen's (1973) formulation for post-1977 major earthquakes, It was concluded, however, that the polar motion excited by those earthquakes were too small to be detected even in the modern Earth rotation data. Chao and Gross (]987) developed complete formulae for calculating the earthquake-induced changes in Earth's rotation and low-degree gravitational field based on the normal mode theory of Gilbert (1970). They made calculations for the 2146 major earthquakes that occurred during 1977-1985. Chao and Gross (1994) and Chao ct al. (1994) extended the calculation to include 111015 major earthquakes for the period]977- 1993 to evaluate the corresponding rotational and gravitational energy changes.

The present paper will study the seismic excitation of polar motion in the framework of Chao and Gross (1987) but making use of the updated result of Chao and Gross (]994). The corresponding changes in length-of-day will only be presented in passing, as they are in general two orders of magnitude smaller. ‘l’he physical reason for the latter is the following. The geophysical excitation acts against the ‘(inertia” of the Earth system. For length-of-day the inertia is the Earth’s axial moment of inertia C, whereas the inertia for polar motion is the difference between the axial and equatorial moments of inertia, $C - A$ *(see* equation 2), which is only 1/300 of C.

2. Dynamics and Calculation

The excitation of the polar motion is governed by the conservation of angular momentum. The equation of motion is customarily expressed as (Munk and MacDonald, 1960):

$$m + \frac{i}{\sigma}\frac{dm}{dt} = \Psi \tag{1}$$

where m is the complex-valued polar motion measured in radian, whose real part is the z component (along the Greenwich Meridian) and the imaginary part the y component (along the $90°E$ longitude), σ is the frequency of the free Chandler wobble with a nominal period of 435 days and a Q value of 100, and Ψ is the complex-valued excitation function.

Figure 1: The x and y components of the observed polar motion and its excitation function (Ψ, obtained by deconvolution) for 1976.4-1994.0, in units of milliarcseconds. The straight lines are the least-squares fit to the excitation function.

The polar motion m traces out a prograde, quasi-circular path on the order of 10 m in the vicinity of the North Pole. Figure 1 shows the “Space93” dataset at nominal daily intervals (Gross, 1994) derived from space geodetic measurements during 1’376.4--1994.0 in its x and y components. Besides a slow polar drift, the oscillation] consists mainly of the annual wobble and the Chandler wobble. It is continually excited (otherwise it would decay away in a matter of decades); and the excitation function Ψ can be obtained numerically according to equation (I) in a process of deconvolution. The Ψ thus obtained is also given in Figure 1, together with the least-squares fitted straight lines representing the polar drift (see below). The geophysical problem is to identify and understand the sources of this "observed" Ψ. It is now known that a major source is the variation of the atmospheric angular momentum (Chao, 1993). The problem is far from closed, and the earthquake dislocation remains a candidate excitation source.

The polar-motion excitation Ψ due to mass redistribution is given by

$$\Psi = 1.61(\Delta I_{xz} + i\Delta I_{yz})/(C - A) \tag{2}$$

where I denotes the inertia tensor, the factor 1.61 takes into account of the Earth's non-rigidity and the decoupling of the fluid core from the mantle in the excitation process. Note that Ψ should also include an additional term due to mass motion, but that term is negligible in the case of abrupt seismic sources (Chao, 1984).

With an abrupt step-function time history (compared to the much longer time scale of the polar motion), an earthquake faulting generates a co-seismic, step-function displacement field u (after the seismic waves have died away). Knowing the seismic moment tensor, u can be evaluated by the normal mode summation scheme of Gilbert (1970). The task, then, is to calculate the seismic Ψ according to equation (2), which consists of evaluating I by integrating u over the globe. The reader is referred to Chao and Gross (1987) for details of the formulation and calculation method. This normal mode *scheme* is found to be extremely efficient.

Calculation has been conducted for 11015 major earthquakes (with magnitude greater than 5.0) that occurred during 1977.0-1993.6, using the seismic centroid moment tensor solutions published in the Harvard CMT catalog (Dziewonski et al., 1993). Smaller earthquakes, although numerous, have completely negligible effects. The adopted normal mode eigenfrequencies and eigenfunctions belong to the spherically symmetric Earth model 1066B of Gilbert and Dziewonski (1975). The net effect is then the accumulation of individual step-function contributions: $\Psi(t) = \sum_{i=1}^{11015} \Psi_i H(t - t_i)$.

3. Results and Analysis

Figure 2 shows the calculated polar-motion excitation function $\Psi(t)$ by the earthquakes. The starting value is arbitrarily chosen to be zero. Comparing Figures 2 with 1, it is obvious that the magnitude of the seismic excitation is insignificant during 1977-1993: It is of two orders of magnitude too small to explain the observed polar-motion excitation, so that its presence in Figure 1 is completely overwhelmed by other excitations.

For the purpose of illustration, we single out in Table 1 the results for the following ~~seven~~ eight largest earthquakes in recent decades (with seismic moment M_0 exceeding 10^{21} N m):

Event I:	May 22, 1960,	Chile
Event II:	March 28, 1964,	Alaska, USA
Event III:	August 19, 1977,	Sumba, Indonesia
Event IV:	March 3, 1985,	Chile
Event V:	September 19, 1985,	Mexico
Event VI:	May 23, 1989,	Macquarie Ridge
Event VII:	June 9, 1994,	Bolivia
Event VIII :	October 4, 1994	Kuril Island

The source mechanism of Events I and II, which occurred before the span of the Harvard CMT catalog, are taken from Kanamori and Cipar (1974) and Kanamori (1970), respectively. Events VII and VIII in 1994 ~~is~~ are also outside our studied period. ~~It is a deep-focused event and has~~ They have the largest seismic moments since Event III in 1977.

Table 1. Magnitude and direction of polar-motion excitation by seven great earthquakes, and the corresponding length-of-day changes.

Event	I	II	III	IV	V	VI	VII	VIII
(M_0, 10^{21} N m)	(270)	(75)	(3.6)	(1.0)	(1.1)	(1.4)	(3.0)	(3.9)
$\|\Psi\|$ (mas)	22.6	7.5	0.21	0.18	0.084	0.114	0.376	0.256
arg(Ψ) (°E)	115	198	160	110	277	+323	123	129
ΔLOD (μs)	−8.4	6.8	0.33	−0.10	−0.089	−0.059	0.234	−0.054

Figure 3 compares the power spectra of the observed and the seismic excitations for the same period 1977.0-1993.6. The spectra are computed using Thomson's (1980) multitaper technique after removal of the mean values. The different. spectral characteristic is evident. Tile frequency dependence f^{-n} of the spectrum of the seismic excitation is found to be $n = 2.0$, similar to a Brownian motion process. The power of the seismic Ψ is in general 40-60 dB lower than that of the observed. In particular, the power difference at the Chandler frequency band around 0.83 cycle per year is about 45 dB. We have also calculated and examined possible correlation between the two excitation functions in the Chandler band. Only a weak coherence at moderate confidence level was found

Despite the small magnitude, the seismic excitations ψ collectively are interesting in their own right. Following Chao and Gross (1987), we shall now conduct statistical analyses to reveal their peculiar behavior.

Figure 2: The x and y components of the seismic excitation of polar motion . The straight lines are the least-squares fit to the excitation function.

Figure 3: The (multitapered) power spectrum (in dB, arbitrary unit) of the observed polar motion excitation compared with that induced by earthquakes during 1977.0-1993.6. Mean values have been removed.

Figure 4 shows tile angular histogram ("rose diagram") of the arguments of the 11015 Ψ's in thirty-six 10° increments. Apart from a concentration around $15°E$, an abnormally large number of Ψ's cluster around $140°E$. The distribution pattern is remarkably similar to the 2146 Ψ's analyzed by Chao and Gross (1987), indicating that this pattern is robust with respect to time and number of earthquake samples. In fact., the statistical tendency is stronger with the much greater number of samples: the normalized χ^2 found here is as high as 14,1, compared to 3.81 found in Chao and Gross (1987) and much higher than the 1% significant level of 1.64 for a random distribution (at 35 degrees of freedom). This asserts the extremely non-random nature of the distribution of Figure 4.

Figure 4: The angular histogram ("rose diagram") for the direction of the seismic excitation of polar motion by major earthquakes for 1977.0-1993.6, in 36 angular bins with respect to the terrestrial coordinate system.

The preference of earthquakes in nudging the rotation pole towards $\sim 140°E$ is also evident in Figure 2. The straight lines are the least-squares fit to the curves. The slopes of the lines give the velocity of the polar drift induced by earthquakes. This polar drift velocity vector (0.019 mas, 131° E), as plotted in Figure 5, indeed points to that general direction. The magnitude of the vector is here magnified by 100 times, in order to be shown against other vectors: The one labeled "O&D" (4.5 rnas, $148°E$) is that similarly obtained using O'Connell and Dziewonski's (1976) calculation for 32 great earthquakes during 1900-1964. It is known (Kanamori, 1976) that O'Connell and Dziewonski nave greatly overestimated the size of their studied earthquakes. However, the direction of the resultant polar drift is determined by the source mechanisms according to plate tectonics, and is hence realistic. Thus it appears that the earthquake-induced polar drift has continued its journey toward $\sim 140°E$ at least since 1900 when fairly reliable seismic records had become available.

The other two vectors in Figure 5 are from polar motion observations; they agree well with each other. The one labeled "Spat.e93" (3.84 mas, $-68°E$) is simply obtained from the fitted lines in Figure 1. The one labeled "Pole93" (3.22 mas, $-81°$ E) is similarly obtained from a polar motion dataset for 1900-1993, primarily based on the International Latitude Service data since 1900. It is seen that the observed polar

Figure 5: The calculated polar drift velocitity induced by major earthquakes for 1977.0-1993.6, magnified 100 times, in comparison with O'Connell and Dziewonski's estimate (labeled "O&II") and the observed polar drift (labeled "Space93" and "Pole93").

drift directions are roughly opposite to those induced by earthquakes.

4. Conclusions

What we compute is the co-seismic effects. No consideration is given to possible pre- and post-seismic viscoelastic movement of the fault. Such movement, can augment the co-seismic effect by several times over a period of months to years, depending on the source rheology (e.g. Dragoni et al., 1983).

The direction of the seismic excitation depends on the focal location and source mechanism. It is found that, at least since 1900, the seismic excitations collectively have a statistically strong tendency to nudge the pole towards -- $140^{\circ}E$, away from the actual polar drift direction. This non-random behavior is similar to other earthquake-induced changes in length-of-day and low-degree gravitational field, which indicate a strong tendency for the earthquakes to make the Earth rounder and more compact (Chao and Gross, 1987). Specific questions arises: Why do earthquakes seem to "recognize" the existence of the rotation axis? Is there any as yet unseen, dynamic connections between the pole position and the occurrence and source mechanism of earthquakes? Or are they manifestations of some behind-the-scene geophysical processes? The dynamic reason for such peculiar behavior of earthquakes is not clear, and at the present time the above questions remain mere speculations.

The magnitude of individual earthquake effect, on the other hand, depends largely on the seismic moment of the event. The rule of thumb is that a seismic moment of 10^{22} N m would roughly produce 1 mas in polar-motion excitation and 1μs in length-of-day change; the actual values depend on the focal location and seismic source mechanism (cf. Table 1). The polar motion excitations produced by the largest earthquakes during 1977--1993 were still an order of magnitude too small to be detected in polar motion data even with today's accuracy (1 mas) and temporal resolution (a few days). However, it was calculated that the great, 1960 Chile event should have produced a discontinuity as large as 23 mas in the polar motion excitation function (but only 8 μs in LOD) So an earthquake of only one tenth the size of that event, if happened today, could be comfort ably detect.ed in polar mot ion observations. Furt her improvements in observation should allow detection of smaller (and hence potentially more numerous) earthquake signatures in the future.

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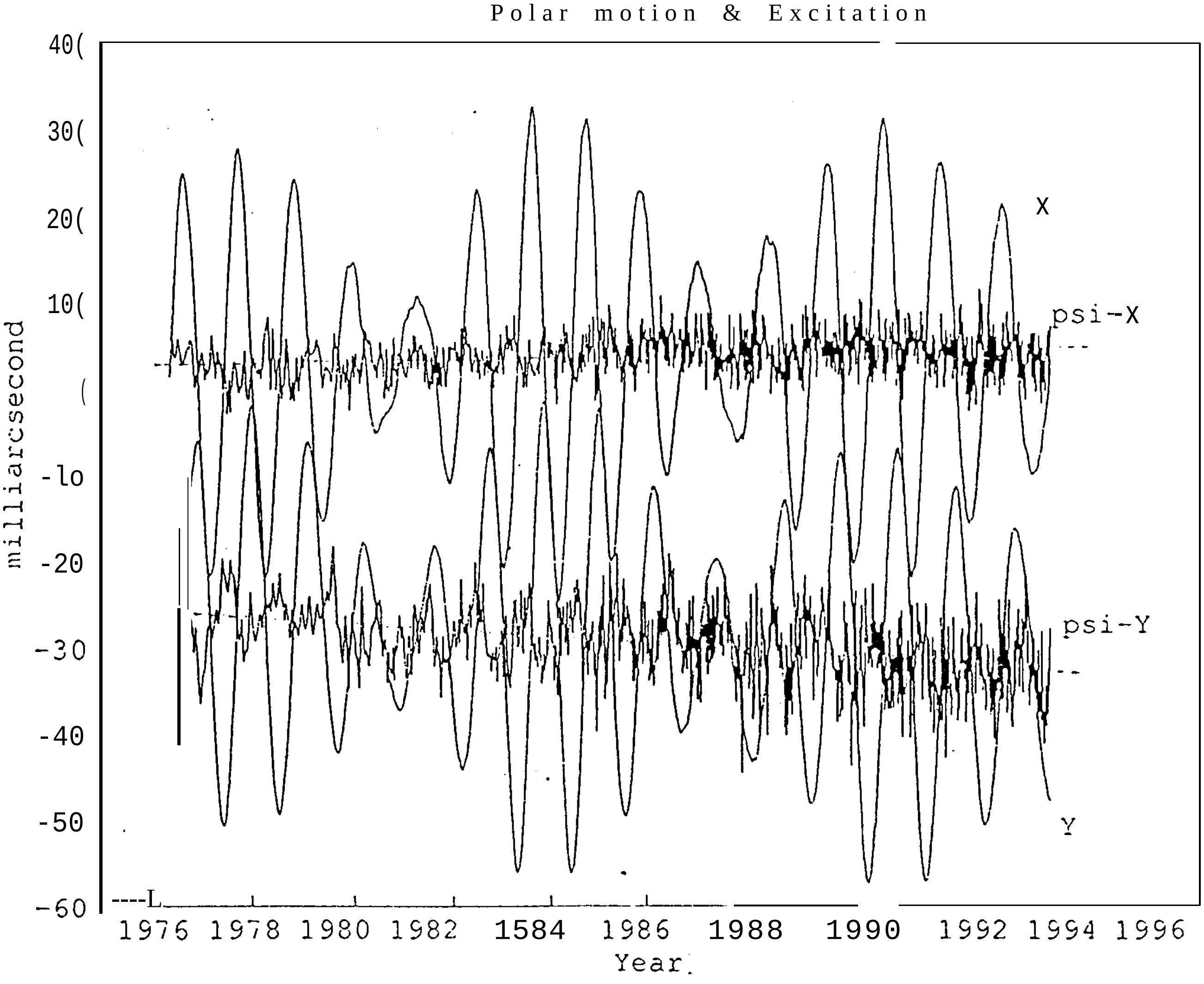

Polar motion & Excitation
40(
30(
20(
10(
(
-lo
-20
-30
-40
-50
-60
milliarcsecond
X
psi-X
psi-Y
Y
1976 1978 1980 1982 1584 1986 1988 1990 1992 1994 1996
Year

Fig. 1.

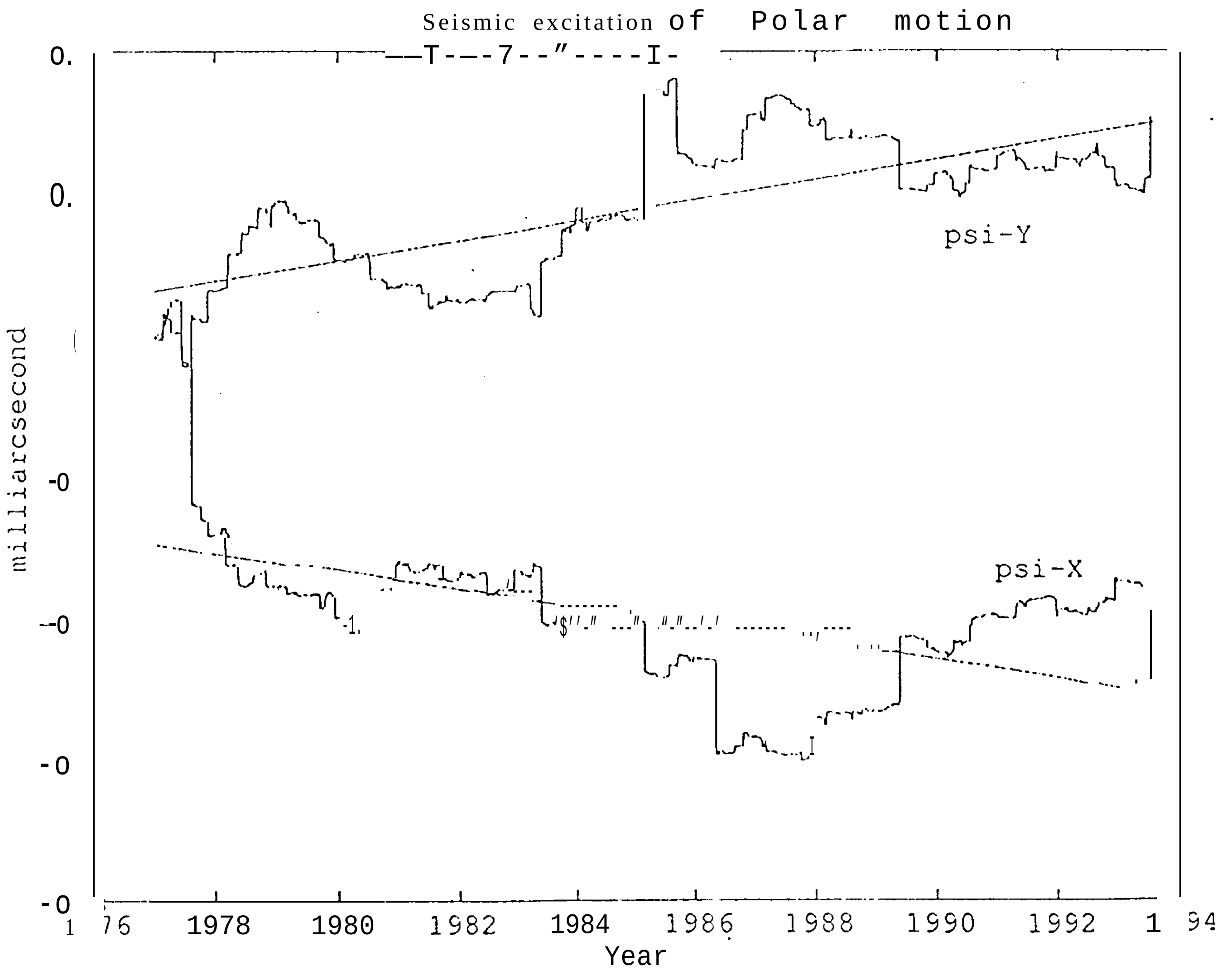
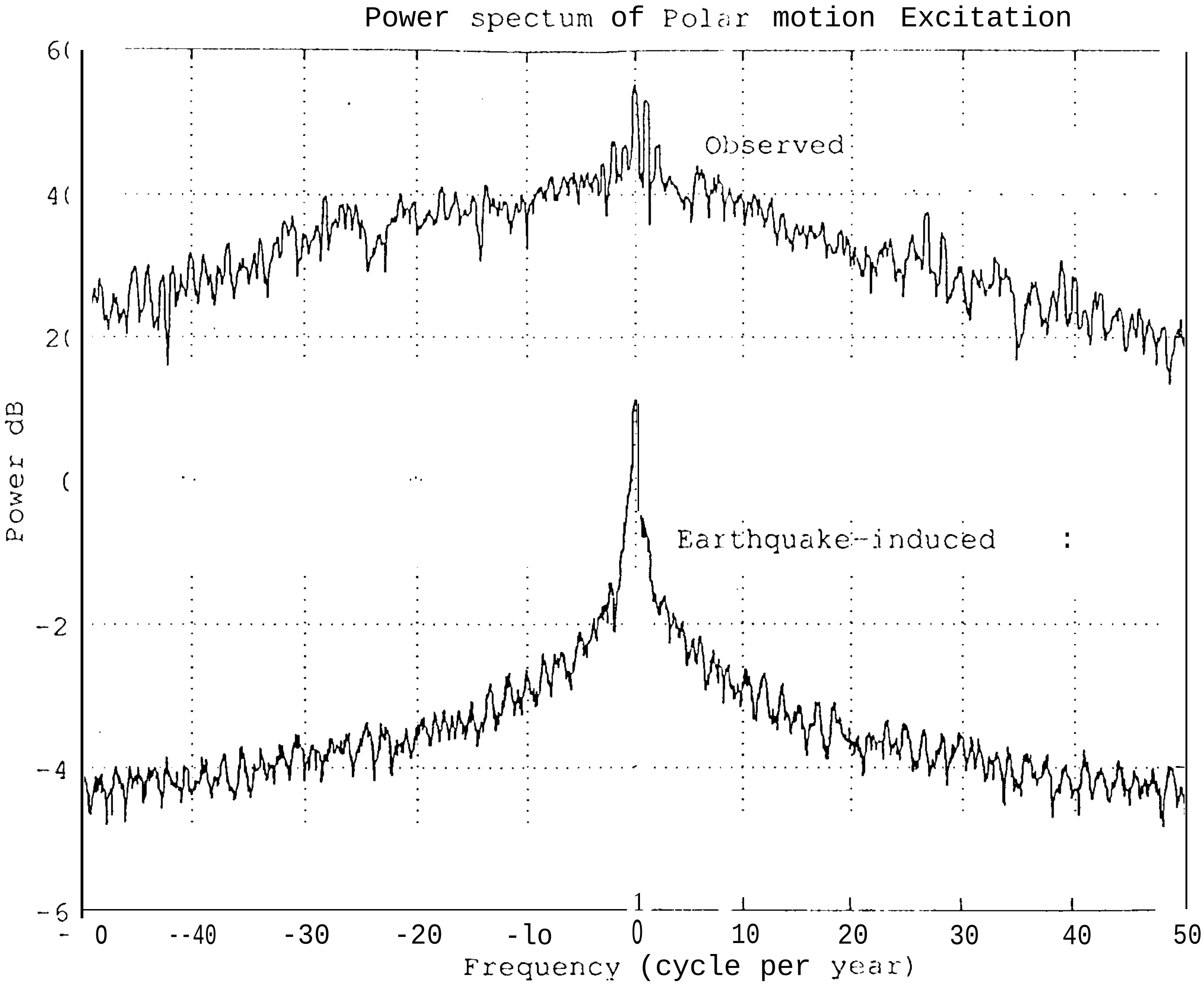

Seismic excitation of Polar motion
milliarcsecond
Year
psi-Y
psi-X
1978
1980
1982
1984
1986
1988
1990
1992

Fig. 2



Power spectum of Polar motion Excitation
Observed
Earthquake-induced
Power dB
6(
4(
2(
(
-2
-4
-6
- 0
--40
-30
-20
-lo
0
10
20
30
40
50
Frequency (cycle per year)

Fig. 3

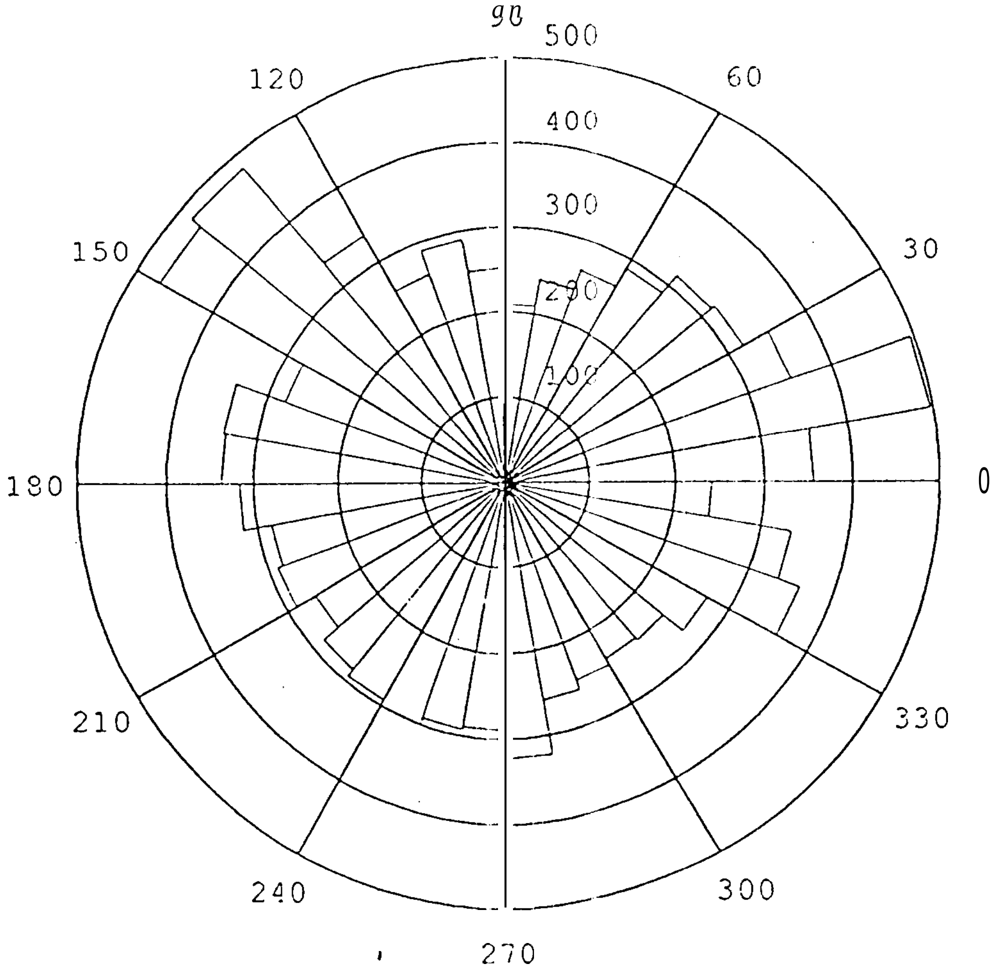


Fig. 4

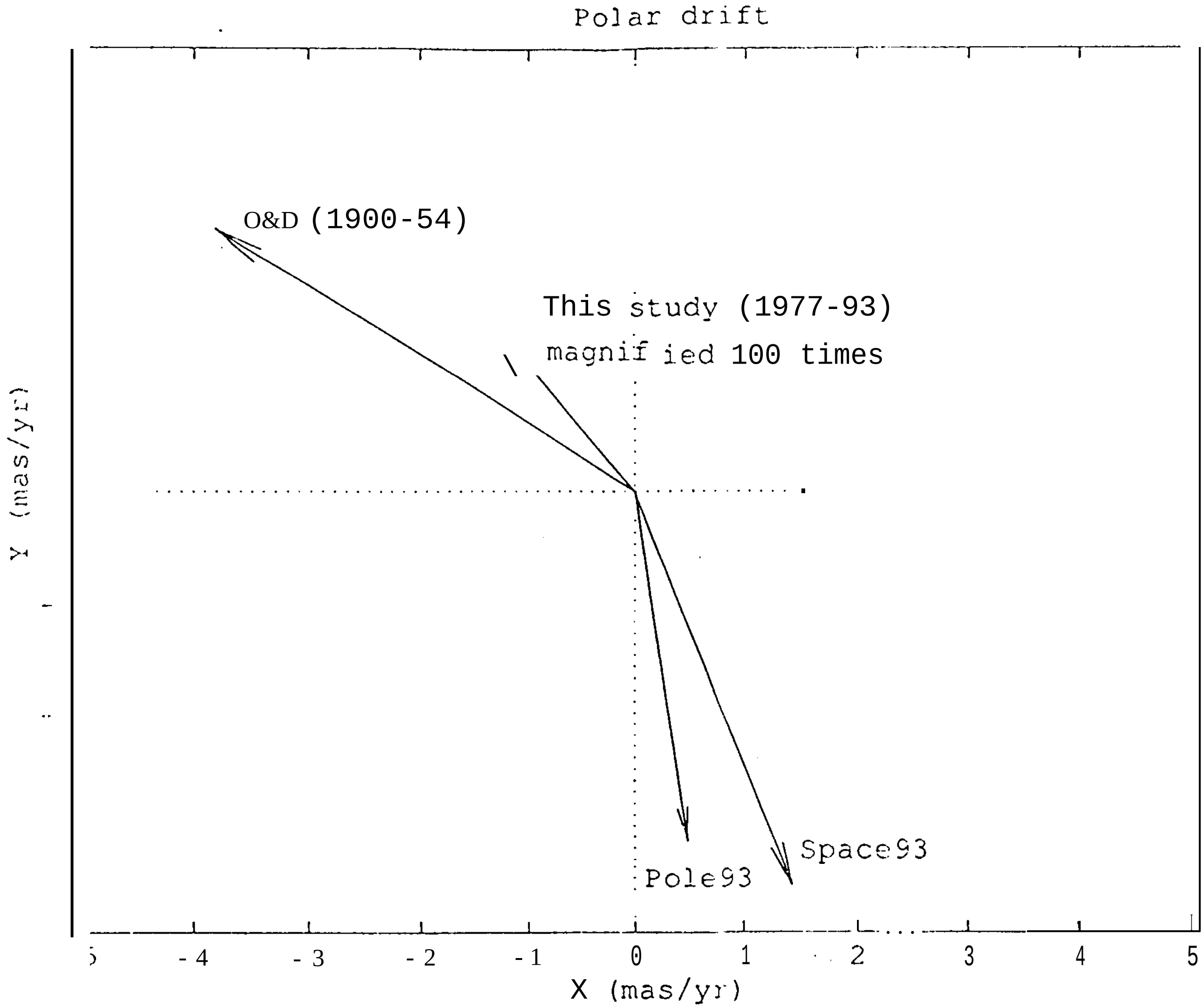

Polar drift
O&D (1900-54)
This study (1977-93)
magnified 100 times
Space93
Pole93
Y (mas/yr)
X (mas/yr)
-4
-3
-2
-1
0
1
2
3
4
5

Fig. 5