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Polar Motion Excited by Atmosphere and Ocean in Multi-frequency Bands

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Abstract This research aims to study the influences of the atmospheric and oceanic excitations on polar motion. Power spectrum density analyses show that the efficiencies of the atmospheric and oceanic excitations differ not only at different frequencies but also in the retrograde and prograde components, but the sum of atmospheric and oceanic excitations shows the best agreement with the observed excitation.

Keywords polar motion; atmospheric excitation; oceanic excitation

CLC number P223; P222

Introduction

Polar motion refers to the displacement of the Earth’s rotation axis in a frame tied to the Earth (or more precisely, the mantle). It contains two dominant components: one is the annual wobble (AW) with a ~12 month period and a ~100 mas (milli-arc-second) amplitude, and the other is the famous Chandler wobble (CW) with a ~14 month period and a ~160 mas amplitude.

The AW is now considered to be excited mainly by the seasonal fluctuations in the atmospheric and oceanic angular moments (denoted by AAM and OAM respectively). That is, the angular moment exchanges between the solid Earth and its superficial fluid layers (such as the atmosphere and ocean) give rise to this annual polar motion, since the angular moment of the solid Earth-atmosphere-ocean system should be conserved without external torques.

The AW excitation is mainly attributed to the atmospheric and oceanic effects due to the fact that the annual component dominates the AAM and OAM. However, the AAM and OAM also contain abundant fluctuations in other frequency bands, and these fluctuations will cause corresponding variations of the pole coordinate. It is the motivation of this study to evaluate the efficiencies of the atmospheric and oceanic excitations of polar motions at more frequencies besides the annual one.
1 Dynamic equation of polar motion

The governing equations of polar motion, namely the Liouville equation, can be written as \[1,2\]
\[
\frac{i}{\sigma_c} m + m = \psi 
\]
where \(\sigma_c\) is the Chandler frequency. To include its damping effect, the Chandler frequency can be expressed by a complex notation \[1,2\]
\[
\sigma_c = \sigma_0 (1 + \frac{i}{2Q})
\]
where \(\sigma_0 = \frac{1}{433}\) cycle per day and \(Q = 100\) is the quality factor of the CW. Let \(C\) and \(A\) be the principal moments of inertia, \(\Omega\) and \(\omega = \Omega(m_1 + m_2, 1 + m_3)\) be the mean and instantaneous rotation rate of the Earth, respectively, the excitation function \(\psi\) can be expressed as
\[
\psi = X_{\text{wobble}} \frac{\Omega^2 c - i \Omega \dot{c} + \Omega h - i \dot{h} + i L}{\Omega^2 (C - A)}
\]
with the transfer function \(X_{\text{wobble}}\) incorporating the effects of the rotational and loading deformation of the mantle, as well as the core-mantle decoupling. In Eqs.(1-3), the traditional complex notations are given by
\[
\begin{bmatrix}
m = m_1 + im_2 \\
c = c_{13} + ic_{23} \\
h = h_1 + ih_2 \\
\psi = \psi_r + i\psi_i
\end{bmatrix}
\]
where \(c\) and \(h\) are the product of inertia (caused by the mass redistribution within the Earth system), relative angular momentum (caused by the relative motion with respect to the mantle), respectively. As is well known, the solution to Eq.(1) is \[1,2\]
\[
m(t) = m + \frac{e^{i\sigma_c t}}{\sigma_c} \left[ m_0 - i\sigma_c \int_0^t \psi(\tau) e^{-i\sigma_c \tau} d\tau \right]
\]
The transfer function \(X_{\text{wobble}}\) takes the form
\[
X_{\text{wobble}} = (1 + k') \frac{k_s}{k_s - k} \frac{A}{A_m} = 1.0980
\]
when the geoprocess, perturbing the polar motion, loads the Earth, and takes another form
\[
X_{\text{wobble}} = \frac{k_s}{k_s - k} \frac{A}{A_m} = 1.5913
\]
when it does not. \[3\] In Eqs.(6-7), \(k_s\) and \(k\) are the secular and second-degree Love numbers, respectively, \(k'\) is the second-degree loading Love number, \(A_m\) is the equatorial principal moment of inertia of the mantle; \(k_s/(k_s - k)\) and \((1 - k')\) denote the effects of rotational and loading deformations respectively, and \(A/A_m\) represents the core-mantle decoupling, which is widely accepted as a reasonable assumption in modeling the Earth rotation for a period as long as, or shorter than, the Chandler period.\[1-3,5\]

In fact, \(m\) describes the motion of the rotational pole while Earth orientation observations actually provide the motion of the Celestial Intermediate Pole (CIP) rather than the rotational pole in accordance with the IAU2000 resolutions.\[11\] Thus, we need to find the correspondence between the above theory and the polar motion observations.

Since the terrestrial reference frame (TRF) is fixed to the Earth, the rotation of the Earth is equivalent to the rotation of the TRF. The intermediate reference frame (IRF, with its \(z\) axis pointing to the CIP) is the intermediate transition between the TRF and the celestial reference frame (CRF). The TRF and IRF are linked by three transformation matrices, namely the spinning one \(S(t)\), the polar motion ones \(X(t)\) and \(Y(t)\), and we have \[12\]
\[
r_r(t) = S(t)X(t)Y(t)r_t(t) = A^T(t)r_r(t)
\]
where the superscript \(T\) denotes transposition, \(r_r(t)\) and \(r_t(t)\) are coordinates of the field point in the TRF and IRF, respectively. Gross (2002)\[12\] showed that
\[
W(t) = \frac{dA(t)}{dt} A^T(t)
\]
where
\[
W(t) = \begin{bmatrix} 0 & \omega_x & -\omega_y \\ -\omega_z & 0 & \omega_x \\ \omega_y & -\omega_x & 0 \end{bmatrix}
\]
According to McCarthy & Petit (2003),\[11\] the transformation matrices are
\[
S(t) = \begin{bmatrix} \cos H & -\sin H & 0 \\ \sin H & \cos H & 0 \\ 0 & 0 & 1 \end{bmatrix}
\]
\[
X(t) = \begin{bmatrix} \cos p_x & 0 & -\sin p_x \\ 0 & 1 & 0 \\ \sin p_x & 0 & \cos p_x \end{bmatrix}
\]
\[
Y(t) = \begin{bmatrix}
1 & 0 & 0 \\
0 & \cos p_y & \sin p_y \\
0 & -\sin p_y & \cos p_y
\end{bmatrix}
\]

where \( p = p_x + ip_y = x - iy \). Substituting \( A(t) = S(t) \)
\( X(t)Y(t) \) and Eq.(10) into Eq.(9) and keeping the
first order, one gets
\[
\omega_x = \Omega p_x - \frac{dp_x}{dt}
\]
\[
\omega_y = -\Omega p_y - \frac{dp_y}{dt}
\]
\[
\omega_z = \Omega[1 + ip_z]
\]

Noting \( (\omega_x, \omega_y) = \Omega(m_1, m_2) \), the motions of the
CIP \( (p = x - iy) \) and the rotational pole \( (m = m_1 + im_2) \)
are related with each other by \(^{[12]}\)
\[
m = p - \frac{i}{\Omega} \hat{p}
\]

Adopting the angular momentum function (AMF) introduced by Barnes et al. (1983): \(^{[7]}\)
\[
\chi = \frac{\Omega c + h}{\Omega(C - A)} \equiv \chi_p + \chi_w
\]

the excitation function can be rewritten as
\[
\psi = X_{\text{wobble}}(\chi - i \frac{\chi}{\Omega})
\]

Substituting Eqs.(13-15) into Eq.(1), one gets
\[
\frac{i}{\sigma_c} \hat{p} + p = \chi_{\text{eff}}
\]

where
\[
\chi_{\text{eff}} = X_{\text{wobble}}(\chi) = X_{\text{wobble}} \frac{\Omega c + h}{\Omega(C - A)}
\]
is the effective AMF, and is often called the excitation
function for the CIP.

Eq.(16) is just the Euler-Liouville equations that
represent for the polar motion of CIP, while Eq.(1) is
for the motion of the rotational pole. Based on Eq.(16)
and geodetic observations (such as VLBI and GPS) of
the temporal pole coordinate \((x, y)\), one can get the
observed (or geodetic) AMF
\[
\chi_{\text{obs}} = \frac{i}{\sigma_c} \hat{p} + p = \frac{i}{\sigma_c}(\dot{x} - i\dot{y}) + (x - iy)
\]

Eq.(17) describes the theoretical AMF based on
g eo processes accompanied with mass redistributions
and motions, while Eq.(18) is the observed AMF
which illustrates how strong the total excitations are
required to explain the observed polar motion. The
excitation efficiency of the selected geoprocess can
be evaluated by comparing the results based on
Eqs.(17) and (18).

2 Atmospheric and oceanic excitations

The atmosphere, forced primarily by the diurnal
and seasonal cycles, is the primary excitation source
for the Earth rotation on intraseasonal and seasonal
timescales. \(^{[3,7,8]}\) The atmospheric excitation function
includes two portions: the “wind” terms due to the
atmospheric motion relative to the crust plus mantle
and the “pressure” terms due to the variations of at-
mospheric mass distribution, evident through surface
pressure changes. The pressure terms, relevant to \( c \),
will load the Earth while the wind terms, relevant to \( h \),
will not. Thus, the effective AMF can be written as
\[
\chi_{\text{eff}} = 1.0980\chi_p + 1.5913\chi_w
\]
\[
= 1.0980 \frac{c}{C - A} + 1.5913 \frac{h}{\Omega(C - A)}
\]
where the coefficients 1.0980 and 1.5913 have been
discussed in the last section.

The atmospheric and oceanic AMFs are determined
based on the global AAM \(^{[13-15]}\) as calculated from
NCEP/NCAR (National Centers for Environmental
Prediction/National Center for Atmospheric Research)
re-analyses archived on pressure surfaces, and the
ECCO (Estimating the Circulation and Climate of the
Ocean) values of global OAM \(^{[8]}\) as computed from
the products of an ocean model based on the MIT-GCM
(Massachusetts Institute of Technology-General Cir-
culation Model).

The AAM data, sampled to four times daily from 1
Jan. 1948 to 31 Dec. 2008, are provided by the Global
Geophysical Fluids Data (GGFD) Center of Interna-
tional Earth Rotation and Reference Systems Service
(IERS). We adopt the IB-model for the pressure term,
namely the pressure is associated with oceans react-
ing as ‘Inverted Barometer’ (IB) in front of the pres-
sure variations, which is a realistic approximation for
variations larger than 10 days. \(^{[13-15]}\) Thus, the IB
pressure (or, mass) terms and wind (or, motion) terms
of the AAM data are used to calculate the atmos-

pheric angular momentum function. In the AAM data, the wind terms are computed by integrating winds from the Earth surface to 10 hPa, the top of the atmospheric model. The IB correction involves applying the mean atmospheric surface pressure over the whole world ocean to every point over the world ocean.

The OAM data, also provided by the GGFD, contain mass and motion terms with daily-interval from 1 Jan. 1993 to 31 Dec. 2008. The OAM data is based on the Boussinesq assumption with a horizontal resolution of 1/3 degree at the equator to 1 degree at high latitudes by 1 degree longitude, and a vertical resolution of 46 levels ranging in thickness from 10 m at the surface to 400 m at depth. Since the Boussinesq assumption conserves volume rather than mass, the OAM mass terms have been corrected for volume changes due to steric effects.

The atmospheric-oceanic angular moment functions and excitation functions derived from these AAM and OAM data should be compared with the observations to check their efficiencies. Thus, we also need the Earth Orientation Parameter (EOP) data to calculate the observed (or geodetic) excitations.

The EOP data, routinely provided by the IERS, report the temporal position of the reference pole CIP in accordance with the IAU2000 resolutions. Here we adopt the latest product EOP 05 C04 series, which provides the temporal coordinate \((x, y)\) of the CIP, with the \(x\)-axis pointing to the mean Greenwich meridian plane and the \(y\)-axis pointing to 90°W (west).

The AAM, OAM and EOP data are available at http://ftp.aer.com/pub/anon_collaborations/sba/, http://euler.jpl.nasa.gov/sbo/sbo_data.html and ftp://hpiers.obspm.fr/eop-pc/eop/eopc04_05/, respectively. More information about these data can be accessed at the above mentioned websites. All the adopted data are ranging from 1 Jan. 1993 to 31 Dec. 2008 since the OAM data are available only during that period though the time spans of EOP and AAM series are much longer.

A plot of the AAM data can be referred to the study of Chen et al. (2010) while that for the OAM is illustrated by Fig.1. The fluctuations of the OAM are smaller than those of the AAM, thus the oceanic contributions to polar motion are secondary to the atmospheric ones.

In order to evaluate the efficiencies of the atmospheric and oceanic excitations, the multi-taper method (MTM) is adopted to calculate the power spectrum densities (PSD) of \(\chi^{\text{obs}}\), \(\chi^{\text{eff}}\), \(\chi^{\text{O}}\) and \(\chi^{\text{AO}}\) (see Fig.2), where \(\chi^{\text{obs}}, \chi^{\text{eff}}, \chi^{\text{O}}\) and \(\chi^{\text{AO}}\) denote the geodetic (observed) excitation (GE), the effective atmospheric excitation (AE), the oceanic excitation (OE) and the atmospheric excitation plus oceanic excitation (AE + OE), respectively. The retrograde and prograde components of polar motion are denoted by negative and positive frequencies, respectively. The MTM uses a sequence of orthogonal tapers, specified from the discrete prolate spheroidal

![Fig.1](image)

\(x\) and \(y\) components of OAM data (the means and linear trends are removed)
Fig. 2  Power spectra density (PSD) comparisons among $\chi^{\text{obs}}$ (Observed), $\chi^{\text{at}}_A$ (AE) and $\chi^{\text{at}}_{AO}$ (AE + OE).

sequences, to estimate the modified periodograms and then combines them to estimate the PSD. The MTM can reduce signal energy leakage and provide better resolution and estimate-accuracy than traditional spectrum estimate methods (e.g., the periodogram method). Only the low frequency part of the PSD is presented in Fig. 2 since the IB-model is not adequate at high frequency bands.\[13\text{-}15\]

In Fig. 2, either $\chi^{\text{obs}}$, or $\chi^{\text{at}}_A$, or $\chi^{\text{at}}_{AO}$ contains the annual components (1 cycle per year, cpy), semiannual components (2 cpy) and terannual components (3 cpy). AE is more efficient than OE in the annual and semiannual frequency bands in both retrograde and prograde components, as well as in the retrograde terannual component. Especially for frequencies higher than 3 cpy, AE is in general more significant than OE. However, OE becomes more significant than AE in the prograde terannual component as well as both the retrograde and prograde interannual (with period longer than 1 year but less than 2 years) components. These effects are summarized in Table 1.

Taking the second row (i.e., the annual frequency) of Table 1 as an example, AE and OE are both significant in exciting the prograde annual wobble, but the former is more efficient than the latter; only AE is significant in exciting the retrograde annual wobble.

Overall, $\chi^{\text{at}}_{AO}$ represents better coincidence with $\chi^{\text{obs}}$ than $\chi^{\text{at}}_A$ and $\chi^{\text{at}}_{AO}$, namely, the sum of atmospheric and oceanic excitations shows the best agreement with the observed excitation. The annual and semiannual components of $\chi^{\text{at}}_{AO}$ are close to those of $\chi^{\text{obs}}$, but the oceanic effects tend to decrease the two components of $\chi^{\text{at}}_{AO}$ since AE and OE are out of phase in these frequency bands.\[8\] However, the phase difference between AE and OE in the terannual frequency band seems sufficiently small and consequently their total effect AE + OE is enhanced.

### Table 1  Comparisons of the atmospheric and oceanic excitations

| Frequency   | Prograde | Retrograde |
|-------------|----------|------------|
| Interannual | OE       | OE         |
| Annual      | AE, OE   | AE         |
| Semiannual  | AE, OE   | AE, OE     |
| Terannual   | AE, OE   | AE         |
| $>3$ cpy    | AE       | AE         |

3  Discussion and conclusion

This paper studies the atmospheric and oceanic excitations of polar motions at multiple frequencies, namely the terannual, semiannual, annual and the interannual excitations. The efficiencies of the atmospheric and oceanic excitations differ not only at different frequencies but also in the retrograde and prograde components (see Table 1). In general, AE is more efficient in the higher frequency band (with periods of 1 year and less) while OE is more efficient in the lower frequency band; AE is usually efficient at both the retrograde and prograde wobbles while OE tends to be more efficient at the prograde ones.

Some results of this study are only preliminary
achievements since the time span of the OAM data is relatively too short to give a sound conclusion on the atmospheric and oceanic excitations of long period wobbles (e.g., the interannual and decadal wobbles). Thus, further studies are needed when the time spans of relevant data are sufficiently long.

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