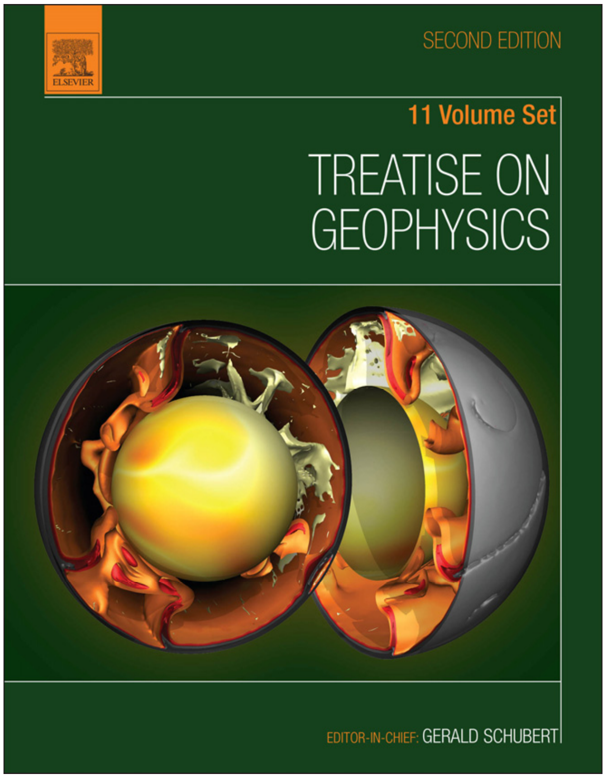

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3.09 Earth Rotation Variations – Long Period

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3.09.1 Introduction

The Earth is a dynamic system – it has a fluid, mobile atmosphere and oceans; a continually changing global distribution of ice, snow, and water; a fluid core that is undergoing some type of hydromagnetic motion; a mantle both thermally convecting and rebounding from the glacial loading of the last ice age; and mobile tectonic plates. In addition, external forces due to the gravitational attraction of the Sun, Moon, and planets also act upon the Earth. These internal dynamical processes and external gravitational forces exert torques on the solid Earth or displace its mass, thereby causing the Earth's rotation to change.

Changes in the rotation of the solid Earth are studied by applying the principle of conservation of angular momentum to the Earth system. Under this principle, the rotation of the solid Earth changes as a result of (1) applied external torques, (2) internal mass redistribution, and (3) the transfer of angular momentum between the solid Earth and the fluid regions with which it is in contact; concomitant torques are due to hydrodynamic or magnetohydrodynamic stresses acting at the fluid/solid Earth interfaces.

Here, changes in the Earth's rotation that occur on timescales greater than a day are discussed. Using the principle of conservation of angular momentum, the equations governing small variations both in the rate of rotation and in the position of the rotation vector with respect to the Earth's crust are first derived. These equations are then rewritten in terms of the Earth rotation parameters that are actually reported by Earth rotation measurement services. The techniques that are used to monitor the Earth's rotation by the measurement services are then reviewed, a description of the variations that are observed by these techniques is given, and the possible causes of the observed variations are discussed.

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3.09.2 Theory of Earth Rotation Variations at Long Periods

3.09.2.1 Instantaneous Rotation Vector

In a rotating reference frame that has been attached in some manner to the solid body of the Earth, the Euler equation of motion that relates changes in the angular momentum $\mathbf{L}(t)$ of the Earth to the external torques $\boldsymbol{\tau}(t)$ acting on it is (Eubanks, 1993; Lambeck, 1980, 1988; Moritz and Mueller, 1988; Munk and MacDonald, 1960; Seitz and Schuh, 2010)

$$\frac{\partial \mathbf{L}(t)}{\partial t} + \boldsymbol{\omega}(t) \times \mathbf{L}(t) = \boldsymbol{\tau}(t) \qquad [1]$$

where, strictly speaking, $\boldsymbol{\omega}(t)$ is the angular velocity of the rotating frame with respect to inertial space. But since the rotating frame has been attached to the solid body of the Earth, it is also interpreted as being the angular velocity of the Earth with respect to inertial space. In general, the angular momentum $\mathbf{L}(t)$ can be written as the sum of two terms: (1) that part $\mathbf{h}(t)$ due to motion relative to the rotating reference frame and (2) that part due to the changing inertia tensor $\mathbf{I}(t)$ of the Earth, which is changing because the distribution of the Earth's mass is changing:

$$\mathbf{L}(t) = \mathbf{h}(t) + \mathbf{I}(t) \cdot \boldsymbol{\omega}(t) \qquad [2]$$

Combining eqns [1] and [2] yields the Liouville equation

$$\frac{\partial}{\partial t}[\mathbf{h}(t) + \mathbf{I}(t) \cdot \boldsymbol{\omega}(t)] + \boldsymbol{\omega}(t) \times [\mathbf{h}(t) + \mathbf{I}(t) \cdot \boldsymbol{\omega}(t)] = \boldsymbol{\tau}(t) \qquad [3]$$

The external torques acting on the Earth due to the gravitational attraction of the Sun, Moon, and planets cause the Earth to nutate and precess. Since the nutations and precession of the Earth are discussed in Chapter 3.10 of this volume, the external torques $\boldsymbol{\tau}(t)$ in eqn [3] will be set to zero. Note, however, that tidal effects on the Earth's rotation, which are also caused by the gravitational attraction of the Sun, Moon, and planets, are discussed here in Sections 3.09.4.1.3 and 3.09.4.2.3.

The Earth's rotation deviates only slightly from a state of uniform rotation, the deviation being a few parts in 10^8 in speed, corresponding to changes of a few milliseconds (ms) in the length of day, and about a part in 10^6 in the position of the rotation axis with respect to the crust of the Earth, corresponding to a variation of several hundred milliarcseconds (mas) in polar motion. Such small deviations in rotation can be studied by linearizing eqn [3]. Let the Earth initially be uniformly rotating at the constant rate Ω about the z-coordinate axis of the body-fixed reference frame and orient the frame within the Earth in such a manner that the inertia tensor of the Earth is diagonal in this frame:

$$\boldsymbol{\omega}_0 = \Omega \hat{\mathbf{z}} \qquad [4]$$

$$\mathbf{I}_0 = \begin{pmatrix} A & 0 & 0 \\ 0 & B & 0 \\ 0 & 0 & C \end{pmatrix} \qquad [5]$$

where the hat denotes a vector of unit length, Ω is the mean angular velocity of the Earth, and A, B, and C are the mean principal moments of inertia of the Earth ordered such that $A < B < C$. In this initial state, in which all time-dependent quantities vanish, the Earth is rotating at a constant rate about its figure axis, there are no mass displacements, and there is no relative angular momentum. So, for example, the atmosphere, oceans, and core are at rest with respect to the solid Earth and merely corotate with it.

Now let this initial state be perturbed by the appearance of mass displacements and relative angular momentum. In general, since the crust and mantle of the Earth can deform, they can undergo motion relative to the rotating reference frame and hence can contribute to the relative angular momentum. However, let the body-fixed reference frame in the perturbed state be oriented in such a manner that the relative angular momentum due to the motion of the crust and mantle vanishes. In this frame, which is known as the Tisserand mean-mantle frame of the Earth (Tisserand, 1891), the motion of the atmosphere, oceans, and core has a relative angular momentum, but the motion of the crust and mantle does not.

In the Tisserand mean-mantle frame, the perturbed instantaneous rotation vector and inertia tensor of the Earth can be written without loss of generality as

$$\boldsymbol{\omega}(t) = \boldsymbol{\omega}_o + \Delta\boldsymbol{\omega}(t) = \Omega \hat{\mathbf{z}} + \Omega \left[m_x(t)\hat{\mathbf{x}} + m_y(t)\hat{\mathbf{y}} + m_z(t)\hat{\mathbf{z}} \right] \qquad [6]$$

$$\begin{aligned} \mathbf{I}(t) &= \mathbf{I}_o + \Delta\mathbf{I}(t) \\ &= \begin{pmatrix} A & 0 & 0 \\ 0 & B & 0 \\ 0 & 0 & C \end{pmatrix} + \begin{pmatrix} \Delta I_{xx}(t) & \Delta I_{xy}(t) & \Delta I_{xz}(t) \\ \Delta I_{xy}(t) & \Delta I_{yy}(t) & \Delta I_{yz}(t) \\ \Delta I_{xz}(t) & \Delta I_{yz}(t) & \Delta I_{zz}(t) \end{pmatrix} \end{aligned} \qquad [7]$$

where the terms with the subscript o denote the initial values given by eqns [4] and [5], the $\Omega m_i(t)$ are the elements of the time-dependent perturbation $\Delta\boldsymbol{\omega}(t)$ to the rotation vector, and the $\Delta I_{ij}(t)$ are the elements of the time-dependent perturbation $\Delta\mathbf{I}(t)$ to the inertia tensor.

The equation that relates small changes in the Earth's rotation to the mass displacements and relative angular momenta that are causing the rotation to change can be derived by substituting eqns [6] and [7] into eqn [3] and then linearizing the resulting expression by assuming that the $h_i(t) \ll \Omega C$, the $m_i(t) \ll 1$, and the $\Delta I_{ij}(t) \ll C$. By keeping terms to first order in these small quantities, the equatorial and axial components of eqn [3] can be written as

$$\frac{1}{\sigma_r}\frac{\partial m_x(t)}{\partial t} + \left[\frac{B(C-B)}{A(C-A)}\right]^{1/2} m_y(t) = -\left(\frac{B}{A}\right)^{1/2}\left[\frac{1}{\Omega}\frac{\partial \phi_{r,x}(t)}{\partial t} - \phi_{r,y}(t)\right] \qquad [8]$$

$$\frac{1}{\sigma_r}\frac{\partial m_y(t)}{\partial t} - \left[\frac{A(C-A)}{B(C-B)}\right]^{1/2} m_x(t) = -\left(\frac{A}{B}\right)^{1/2}\left[\frac{1}{\Omega}\frac{\partial \phi_{r,y}(t)}{\partial t} + \phi_{r,x}(t)\right] \qquad [9]$$

$$\frac{1}{\Omega}\frac{\partial m_z(t)}{\partial t} = -\frac{1}{\Omega}\frac{\partial \phi_{r,z}(t)}{\partial t} \qquad [10]$$

where the external torques have been set to zero,

$$\sigma_r^2 = \left(\frac{C-A}{A}\right)\left(\frac{C-B}{B}\right)\Omega^2 \qquad [11]$$

and the $\phi_{r,i}(t)$, known as excitation functions, are

$$\phi_{r,x}(t) = \frac{h_x(t) + \Omega \Delta I_{xz}(t)}{\Omega\sqrt{(C-A)(C-B)}} \qquad [12]$$

$$\phi_{r,y}(t) = \frac{h_y(t) + \Omega\Delta I_{yz}(t)}{\Omega\sqrt{(C-A)(C-B)}} \quad [13]$$

$$\phi_{r,z}(t) = \frac{1}{C\Omega}[h_z(t) + \Omega\Delta I_{zz}(t)] \quad [14]$$

Equations [8] and [9] are coupled, first-order differential equations that describe the motion of the rotation pole in the rotating, body-fixed reference frame as it responds to the applied excitation. In the absence of excitation, the solution of these equations, which describes the natural or free motion of the rotation pole, can be written as

$$m_x(t) = m\cos(\sigma_r t + \alpha) \quad [15]$$

$$m_y(t) = \left[\frac{A(C-A)}{B(C-B)}\right]^{1/2} m\sin(\sigma_r t + \alpha) \quad [16]$$

where m is the amplitude of the motion along the x-axis and α is the phase of the motion. The natural motion described by eqns [15] and [16] is prograde undamped elliptical motion of frequency σ_r. Using the values in **Table 1** for A, B, and C of the whole Earth, the period of the natural frequency, given by $2\pi/\sigma_r$, is found to be 304.46 sidereal days.

Table 1 Geodetic parameters of the Earth

Parameter	*Value*	*Source*
G	6.67259×10^{-11} m^3 (kg s^2)$^{-1}$	(a)
M_{atm}	5.1441×10^{18} kg	(b)
M_{ocn}	1.4×10^{21} kg	(c)
Whole Earth (*observed*)		
Ω	7.292115×10^{-5} rad s^{-1}	(a)
M	5.9737×10^{24} kg	(a)
C	8.0365×10^{37} kg m^2	(a)
B	8.0103×10^{37} kg m^2	(a)
A	8.0101×10^{37} kg m^2	(a)
$C-A$	2.6398×10^{35} kg m^2	(a)
$C-B$	2.6221×10^{35} kg m^2	(a)
$B-A$	1.765×10^{33} kg m^2	(a)
Whole Earth (*modeled*)		
a	6371.0 km	(d)
n_0	0.15505	(e)
k_2	0.298	(f)
$\Delta k_{\text{ocn,w}}$	0.047715	(g)
$\Delta k_{\text{ocn,s}}$	0.043228	(g)
k_r	0.997191	(g)
k'_2	-0.305	(f)
$\Delta k'_{\text{an}}$	$-0.011 + i\,0.003$	(f)
α_3	0.792	(h)
Crust and mantle (*PREM*)		
ε_a	3.334×10^{-3}	(d)
M_m	4.0337×10^{24} kg	(d)
C_m	7.1236×10^{37} kg m^2	(d)
A_m	7.0999×10^{37} kg m^2	(d)
Core (*PREM*)		
ε_c	2.546×10^{-3}	(d)
M_c	1.9395×10^{24} kg	(d)
C_c	9.1401×10^{36} kg m^2	(d)
A_c	9.1168×10^{36} kg m^2	(d)

PREM, Preliminary Reference Earth Model (Dziewonski and Anderson, 1981).
(a) Groten (2004), (b) Trenberth and Guillemot (1994), (c) Yoder (1995), (d) Mathews et al. (1991), (e) Dahlen (1976), (f) Wahr (2005), (g) this work, and (h) Wahr (1983).

Euler (1765) first predicted that the Earth should freely wobble as it rotates and that the period of this free wobble, assuming that the Earth is rigid, would be about 10 months. However, it was not until 1891 that the free wobble of the Earth was first detected in astronomical observations by Seth Carlo Chandler, Jr. (Chandler, 1891), albeit at a period of 14 months. The free wobble of the Earth is now known as the Chandler wobble in his honor.

Equations [8]–[14] describe changes in the rotation of a rigid body of arbitrary shape that is subject to small perturbing excitation. By recognizing that $(B-A)/A = 2.2 \times 10^{-5} \ll 1$ for the Earth (see **Table 1**), so that dynamically the Earth is nearly axisymmetric, eqns [8] and [9] can be simplified by replacing A and B in them with the average $A' = (A+B)/2$ of the equatorial principal moments of inertia of the Earth:

$$\frac{1}{\sigma_{ra}}\frac{\partial m_x(t)}{\partial t} + m_y(t) = \phi_{ra,y}(t) - \frac{1}{\Omega}\frac{\partial \phi_{ra,x}(t)}{\partial t} \quad [17]$$

$$\frac{1}{\sigma_{ra}}\frac{\partial m_y(t)}{\partial t} - m_x(t) = -\phi_{ra,x}(t) - \frac{1}{\Omega}\frac{\partial \phi_{ra,y}(t)}{\partial t} \quad [18]$$

where the excitation functions $\phi_{ra,i}(t)$ of a rigid axisymmetric body are

$$\phi_{ra,x}(t) = \frac{h_x(t) + \Omega\Delta I_{xz}(t)}{(C-A')\Omega} \quad [19]$$

$$\phi_{ra,y}(t) = \frac{h_y(t) + \Omega\Delta I_{yz}(t)}{(C-A')\Omega} \quad [20]$$

In the absence of excitation, the free motion of a rigid axisymmetric body is prograde undamped circular motion of natural frequency

$$\sigma_{ra} = \left(\frac{C-A'}{A'}\right)\Omega \quad [21]$$

Note that eqns [10] and [14], which describe changes in the rate of rotation of a rigid body, are the same whether the body is dynamically axisymmetric or triaxial.

Equations [10], [14], and [17]–[21] describe changes in the rotation of a rigid axisymmetric body that is subject to small perturbing excitation. But the Earth is not rigid – it has an atmosphere and oceans, a fluid core, and a solid crust and mantle that can deform in response not only to the applied excitation but also to changes in rotation that are caused by the excitation. In general, changes in rotation can be expected to cause changes both in the Earth's inertia tensor and in relative angular momentum. However, by definition of the Tisserand mean-mantle frame, there are no changes in relative angular momentum caused by the motion of the crust and mantle. Furthermore, if it is assumed that the oceans stay in equilibrium as the rotation of the solid Earth changes so that no oceanic currents are generated by the changes in rotation, then there are also no changes in relative angular momentum due to the motion of the oceans. And the effects of the atmosphere can be ignored here because of its relatively small mass (see **Table 1**). Thus, only the core will contribute to changes in relative angular momentum caused by changes in rotation.

Smith and Dahlen (1981) used the results of Hough (1895) to show that the change $\delta h_i(\sigma)$ in relative angular momentum due to core motion caused by a rigid rotation of an axisymmetric crust and mantle is

$$\begin{pmatrix} \delta h_x(\sigma) \\ \delta h_y(\sigma) \\ \delta h_z(\sigma) \end{pmatrix} = \begin{pmatrix} E & iE' & 0 \\ -iE' & E & 0 \\ 0 & 0 & \tilde{E} \end{pmatrix} \begin{pmatrix} m_x(\sigma) \\ m_y(\sigma) \\ m_z(\sigma) \end{pmatrix} \quad [22]$$

where to first order in the ellipticity ε_c of the surface of the core and at frequencies $\sigma \ll \Omega$

$$E = (\sigma^2/\Omega)A_c \quad [23]$$

$$E' = -\sigma(1-\varepsilon_c)A_c \quad [24]$$

$$\tilde{E} = -\Omega C_c \quad [25]$$

where A_c and C_c are the equatorial and axial principal moments of inertia of the core, respectively, and eqn [25] for $\tilde{E}$ has been inferred here by realizing that the core cannot respond to axial changes in the rotation of the mantle if the core–mantle boundary is axisymmetric and if there is no coupling between the core and the mantle (Merriam, 1980; Wahr et al., 1981; Yoder et al., 1981). Note that because of the assumption of dynamical axisymmetry, the equatorial components of eqn [22] are uncoupled from the axial, so that no spin–wobble coupling is introduced by the response of the core to changes in rotation.

Dahlen (1976) studied the passive influence of the oceans on the Earth's rotation, including the changes δI_{ij} in the Earth's inertia tensor that are caused by changes in the rotation of the Earth. In the absence of oceans and assuming that the Earth responds to the centripetal potential associated with changes in rotation in exactly the same manner that a nonrotating Earth would respond to a static potential of the same amplitude and type, Dahlen (1976) found

$$\begin{pmatrix} \delta I_{xz} \\ \delta I_{yz} \\ \delta I_{zz} \end{pmatrix} = \frac{a^5\Omega^2}{3G} \begin{pmatrix} k_2 & 0 & 0 \\ 0 & k_2 & 0 \\ 0 & 0 & n_o + \frac{4}{3}k_2 \end{pmatrix} \begin{pmatrix} m_x \\ m_y \\ m_z \end{pmatrix} \quad [26]$$

where k_2 is the second-degree body tide Love number of the whole Earth (not just of the mantle (see Smith and Dahlen, 1981, p. 239; although for a different opinion, see Dickman, 2005), n_o arises from the change in the mean moment of inertia of the Earth caused by the term in the centripetal potential that gives rise to a purely radial deformation of the Earth, G is the Newtonian gravitational constant, and a is the mean radius of the Earth (i.e., the radius of a sphere having the same volume as the Earth).

When equilibrium oceans are present, Dahlen (1976) found that eqn [26] for the changes in the inertia tensor caused by changes in rotation is modified to

$$\begin{pmatrix} \delta I_{xz} \\ \delta I_{yz} \\ \delta I_{zz} \end{pmatrix} = \begin{pmatrix} D+\delta D & \delta D_{12} & \delta D_{13} \\ \delta D_{12} & D-\delta D & \delta D_{23} \\ \delta D_{13} & \delta D_{23} & \tilde{D} \end{pmatrix} \begin{pmatrix} m_x \\ m_y \\ m_z \end{pmatrix} \quad [27]$$

where

$$D = (k_2 + \Delta k_{\mathrm{ocn,w}})\frac{a^5\Omega^2}{3G} \quad [28]$$

$$\tilde{D} = \left[n_o + \frac{4}{3}(k_2 + \Delta k_{\mathrm{ocn,s}})\right]\frac{a^5\Omega^2}{3G} \quad [29]$$

where the influence of equilibrium oceans has been written in terms of an 'oceanic Love number' Δk_{ocn}, which modifies the second-degree body tide Love number k_2. Dahlen (1976) found that because of the nonuniform distribution of the oceans, the oceanic Love number is different for each component. However, the average of the equatorial components has been taken here to define a mean oceanic Love number $\Delta k_{\mathrm{ocn,w}}$ for the wobble. From eqn [27], it is seen that the nonuniform distribution of the oceans has also coupled the equatorial components to the axial via the off-diagonal elements δD_{13} and δD_{23}. However, the numerical results of Dahlen (1976) for the Earth model 1066A (Gilbert and Dziewonski, 1975) show that this coupling is very weak, with $\delta D_{13}/D = 2.17 \times 10^{-3}$ and $-\delta D_{23}/D = 0.55 \times 10^{-3}$. The coupling between the equatorial components is also very weak, with $-\delta D_{12}/D = 3.15 \times 10^{-3}$.

Using eqns [22]–[25] to account for the relative angular momentum of the core caused by changes in rotation, using eqns [27]–[29] to account for both the rotational deformation of the Earth and the passive response of equilibrium oceans to changes in rotation, and keeping terms to first order in small quantities, the linearized Liouville equation becomes

$$\frac{1}{\sigma_{\mathrm{cw}}}\frac{\partial m_x(t)}{\partial t} + m_y(t) = \chi_y(t) - \frac{1}{\Omega}\frac{\partial \chi_x(t)}{\partial t} \quad [30]$$

$$\frac{1}{\sigma_{\mathrm{cw}}}\frac{\partial m_y(t)}{\partial t} - m_x(t) = -\chi_x(t) - \frac{1}{\Omega}\frac{\partial \chi_y(t)}{\partial t} \quad [31]$$

$$\frac{1}{\Omega}\frac{\partial m_z(t)}{\partial t} = -\frac{1}{\Omega}\frac{\partial \chi_z(t)}{\partial t} \quad [32]$$

where the theoretical frequency of the Chandler wobble is

$$\sigma_{\mathrm{cw}} = \left(\frac{C - A' - D}{A'_m + \varepsilon_c A_c + D}\right)\Omega \quad [33]$$

where $A'_m = A' - A_c$ is the equatorial principal moment of inertia of the crust and mantle, and the excitation functions $\chi_i(t)$ are

$$\chi_x(t) = \frac{h_x(t) + \Omega(1 + k'_2)\Delta I_{xz}(t)}{(C - A' - D)\Omega} \quad [34]$$

$$\chi_y(t) = \frac{h_y(t) + \Omega(1 + k'_2)\Delta I_{yz}(t)}{(C - A' - D)\Omega} \quad [35]$$

$$\chi_z(t) = k_r \frac{h_z(t) + \Omega(1 + \alpha_3 k'_2)\Delta I_{zz}(t)}{C_m\Omega} \quad [36]$$

where C_m is the axial principal moment of inertia of the crust and mantle and k_r is a factor, whose value is near unity (see **Table 1**), that accounts for the effects of rotational deformation on the axial component:

$$k_r = \left\{1 + \left[n_o + \frac{4}{3}(k_2 + \Delta k_{\mathrm{ocn,s}})\right]\frac{a^5\Omega^2}{3G}\frac{1}{C_m}\right\}^{-1} \quad [37]$$

The deformation of the Earth associated with surficial excitation processes that load the solid Earth has been taken into account in eqns [34]–[36] by including the second-degree load Love number k'_2 where, because of core decoupling, the load Love number in the axial component is modified by a factor of α_3 (Dickman, 2003; Merriam, 1980; Nam and Dickman, 1990; Wahr, 1983). Expressions for the excitation functions for processes that do not load the solid Earth can be recovered from eqns [34]–[36] by setting the load Love number k'_2 to zero.

Equations [30]–[36] describe changes in the rotation of an elastic axisymmetric body having a fluid core and equilibrium oceans that is subject to small perturbing excitation. Equation [33] for the theoretical Chandler frequency of such a body was first derived by Smith and Dahlen (1981). Applying this result to the Earth, they found that the elasticity of the solid Earth lengthens the period of the Chandler wobble from the rigid Earth value by 143.0 sidereal days, deformation of the oceans lengthens it a further 29.8 sidereal days, and the presence of a fluid core decreases it by 50.5 sidereal days. Using the values in **Table 1** for the Earth, the theoretical period of the Chandler wobble is found to be 426.8 sidereal days, or about 7.4 sidereal days shorter than the observed period of 434.2 ± 1.1 (1σ) sidereal days (Wilson and Vicente, 1990). This discrepancy between the theoretical and observed periods of the Chandler wobble is probably mainly due to the effects of mantle anelasticity, since departures of the oceans from equilibrium as large as 1% increase the Chandler period by only 0.3 sidereal days (Smith and Dahlen, 1981).

Mantle anelasticity modifies the body tide Love number k_2 and hence, via D, the frequency of the Chandler wobble and the equatorial excitation functions. It also modifies the load Love number k'_2. In the absence of accurate models of mantle anelasticity at the frequencies of interest here, namely, at frequencies $\sigma < \Omega$, a hybrid approach is taken to include its effects. The observed complex-valued frequency σ_o of the Chandler wobble is substituted for its theoretical value in eqns [30] and [31]. It is also substituted for its theoretical value in the excitation functions after rewriting them in terms of the theoretical value by using eqn [33] to eliminate D. The results are

$$\frac{1}{\sigma_o}\frac{\partial m_x(t)}{\partial t} + m_y(t) = \chi_y(t) - \frac{1}{\Omega}\frac{\partial \chi_x(t)}{\partial t} \quad [38]$$

$$\frac{1}{\sigma_o}\frac{\partial m_y(t)}{\partial t} - m_x(t) = -\chi_x(t) - \frac{1}{\Omega}\frac{\partial \chi_y(t)}{\partial t} \quad [39]$$

$$m_z(t) \quad = \quad -\chi_z(t) \quad [40]$$

where the excitation functions become

$$\chi_x(t) = \frac{h_x(t) + \Omega\left[1 + \left(k'_2 + \Delta k'_{\mathrm{an}}\right)\right]\Delta I_{xz}(t)}{\left[C - A' + A'_m + \varepsilon_c A_c\right]\sigma_o} \quad [41]$$

$$\chi_y(t) = \frac{h_y(t) + \Omega\left[1 + \left(k'_2 + \Delta k'_{\mathrm{an}}\right)\right]\Delta I_{yz}(t)}{\left[C - A' + A'_m + \varepsilon_c A_c\right]\sigma_o} \quad [42]$$

$$\chi_z(t) = k_r \frac{h_z(t) + \Omega\left[1 + \alpha_3\left(k'_2 + \Delta k'_{an}\right)\right]\Delta I_{zz}(t)}{C_m \Omega} \quad [43]$$

where $\Delta k'_{\mathrm{an}}$ accounts for the effects of mantle anelasticity on the load Love number. Equations [38]–[43] are the final expressions for the changes $m_i(t)$ in the rotation of the Earth caused by small excitation $\chi_i(t)$. Numerically, using 434.2 sidereal days (Wilson and Vicente, 1990) for the observed period of the Chandler wobble and the values in **Table 1** for the other constants, the real parts of the excitation functions can be written as

$$\chi_x(t) = \frac{1.608\left[h_x(t) + 0.684\,\Omega \Delta I_{xz}(t)\right]}{(C - A')\Omega} \quad [44]$$

$$\chi_y(t) = \frac{1.608\left[h_y(t) + 0.684\,\Omega \Delta I_{yz}(t)\right]}{(C - A')\Omega} \quad [45]$$

$$\chi_z(t) = \frac{0.997}{C_m \Omega}\left[h_z(t) + 0.750\,\Omega \Delta I_{zz}(t)\right] \quad [46]$$

These results agree with those of Wahr (1982, 1983, 2005) to within 2%, with most of the disagreement being due to differences in the values of the numerical constants.

The approach used here to derive the linearized Liouville equations that can be used to study small changes in the Earth's rotation follows the approach of Smith and Dahlen (1981) and Wahr (1982, 1983, 2005). Other approaches have been given by Barnes et al. (1983), Eubanks (1993) (also see Aoyama and Naito, 2000), Dickman (1993, 2003), and Jochmann (2009). Dickman (2003) compared some of these different approaches and discussed the implications of nonzero coupling between the core and mantle; Wahr (2005) discussed the implications of mantle anelasticity. A frequency-dependent transfer function approach has been used by Chen and Shen (2010) to develop a theory of the Earth's rotation that accounts for triaxiality of the mantle and core, anelasticity of the mantle, and dissipation in the oceans. The influence of triaxiality on oceanless elastic bodies with a fluid core, with application to the rotation of Mars, has been studied by Yoder and Standish (1997) and Van Hoolst and Dehant (2002).

3.09.2.2 Celestial Intermediate Pole

Small changes in the Earth's rotation caused by small changes in relative angular momentum or small changes in the Earth's inertia tensor can be studied using eqns [3]–[43] where $\Omega m_i(t)$ are the elements of the change $\Delta\boldsymbol{\omega}(t)$ to the Earth rotation vector, so that $m_x(t)$, $m_y(t)$, and $1 + m_z(t)$ are the direction cosines of the rotation vector with respect to the coordinate axes of the rotating, body-fixed terrestrial reference frame. Alternatively, $m_x(t)$ and $m_y(t)$ can be interpreted as being the angular offsets of the rotation vector from the $\hat{\mathbf{z}}$-axis of the rotating reference frame in the $\hat{\mathbf{x}}$ and $\hat{\mathbf{y}}$ directions. That is, $m_x(t)$ and $m_y(t)$ specify the location of the rotation pole within the rotating, body-fixed terrestrial reference frame, where the rotation pole is that point defined by the intersection of the rotation axis with the surface of the Earth near the North Pole. But Earth rotation measurement services do not report the location of the rotation pole within the rotating, body-fixed terrestrial reference frame. Instead, they report the location of the celestial intermediate pole (CIP).

Just three time-dependent angles, the Euler angles, are required to directly transform the coordinates of some station from the terrestrial frame to the celestial frame. These angles are time-dependent, of course, because the Earth, and hence the body-fixed terrestrial reference frame, is rotating. But, by tradition, an intermediate reference frame is used with the result that five angles are required to completely transform station coordinates from the terrestrial to the celestial reference frames (e.g., Sovers et al., 1998):

$$\mathbf{r}_c(t) = \mathbf{PNSXY}\,\mathbf{r}_t(t) \quad [47]$$

where $\mathbf{r}_t(t)$ are the, in general time-dependent, coordinates of the station in the rotating, body-fixed terrestrial frame; $\mathbf{r}_c(t)$ are the coordinates of the station in the celestial frame; and **P**, **N**, **S**, **X**, and **Y** are the classical transformation matrices with P accounting for the precession of the Earth, **N** accounting for nutation, **S** accounting for spin, and **X** and **Y** accounting for the x- and y-components of polar motion. By first applying **X** and **Y**, the terrestrial coordinates of the station are transformed to an intermediate frame whose reference pole is the CIP; **S** represents a spin through a large angle about the $\hat{z}$-axis of the intermediate frame; **P** and **N** finally transform the intermediate frame to the celestial frame. This approach of using an intermediate frame and five angles (two polar motion parameters, two nutation parameters, and a spin parameter) to transform station coordinates between the terrestrial and celestial reference frames has been traditionally followed in order to separate polar motion from precession and nutation. This separation is done in such a manner that the precessional and nutational motion of the Earth is long period when observed in the celestial reference frame and polar motion is long period when observed in the terrestrial reference frame.

Earth rotation measurement services report the parameters that are needed to carry out the transformation given by eqn [47], namely, the polar motion parameters $p_x(t)$ and $p_y(t)$ that are required in **X** and **Y** and that give the location of the CIP in the rotating, body-fixed terrestrial reference frame, the nutation parameters $\delta\psi(t)$ and $\delta\varepsilon(t)$ that are required in **N** and that are corrections in longitude and obliquity to the adopted nutation model that are needed to give the location of the CIP in the celestial reference frame, and a spin parameter UT1(t) that is required in **S** and that represents the angle through which the Earth has rotated. The precession transformation matrix **P** depends on the lunisolar and planetary precession constants.

The relationship between the polar motion parameters $p_x(t)$ and $p_y(t)$ that are reported by Earth rotation measurement services and the elements $\Omega m_x(t)$ and $\Omega m_y(t)$ of the Earth rotation vector that are needed in eqns [38] and [39] can be derived by considering the properties of transformation matrices (Goldstein, 1950). The transformation of station coordinates between two frames having a common origin implies a rotation. Applying the transformation matrix to the position vector of some station to get its coordinates in a new frame is equivalent to a rotation of the coordinate axes. If the initial reference frame is the terrestrial reference frame of the Earth and the final frame is the celestial reference frame and because the terrestrial reference frame has been fixed to the body of the Earth, then the equivalent rotation of the coordinate axes is simply the rotation of the Earth. Because an intermediate frame has been used to separate polar motion from precession and nutation, in order to derive the relationship between the polar motion parameters $p_x(t)$ and $p_y(t)$ and the elements $\Omega m_x(t)$ and $\Omega m_y(t)$ of the Earth rotation vector, it is sufficient to consider the transformation matrix $\mathbf{A}^{\mathrm{T}}=\mathbf{SXY}$ that transforms station coordinates between the terrestrial and intermediate reference frames:

$$\mathbf{r}_i(t)=\mathbf{A}^{\mathrm{T}}(t)\mathbf{r}_t(t) \qquad [48]$$

where the superscript T denotes the transpose and $\mathbf{r}_i(t)$ are the coordinates of the station in the intermediate frame. The elements ω_i of the rotation vector that is associated with this transformation matrix, that is, of the rotation vector of the Earth, are the three independent elements of the antisymmetric matrix $\mathbf{W}(t)$ (Gross, 1992; Kinoshita et al., 1979):

$$\mathbf{W}(t)=\frac{\mathrm{d}\mathbf{A}(t)}{\mathrm{d}t}\mathbf{A}^{\mathrm{T}}(t)=\begin{pmatrix}0 & \omega_z & -\omega_y\\ -\omega_z & 0 & \omega_x\\ \omega_y & -\omega_x & 0\end{pmatrix} \qquad [49]$$

From Sovers et al. (1998), the classical **X**, **Y**, and **S** transformation matrices are

$$\mathbf{X}(t)=\begin{pmatrix}\cos p_x(t) & 0 & -\sin p_x(t)\\ 0 & 1 & 0\\ \sin p_x(t) & 0 & \cos p_x(t)\end{pmatrix} \qquad [50]$$

$$\mathbf{Y}(t)=\begin{pmatrix}1 & 0 & 0\\ 0 & \cos p_y(t) & \sin p_y(t)\\ 0 & -\sin p_y(t) & \cos p_y(t)\end{pmatrix} \qquad [51]$$

$$\mathbf{S}(t)=\begin{pmatrix}\cos H(t) & -\sin H(t) & 0\\ \sin H(t) & \cos H(t) & 0\\ 0 & 0 & 1\end{pmatrix} \qquad [52]$$

where by tradition the positive direction of $p_y(t)$ is taken to be toward 90°W longitude and H is the hour angle of the true equinox of date that is related to UT1 and the hour angle of the mean equinox of date by the equation of the equinoxes. By forming $\mathbf{A}^{\mathrm{T}}=\mathbf{SXY}$, using eqn [49], and keeping terms to first order in small quantities, the desired elements of the rotation vector of the terrestrial frame with respect to the intermediate frame is obtained (Brzezinski, 1992; Brzezinski and Capitaine, 1993; Gross, 1992):

$$\omega_x(t)=\Omega p_x(t)-\frac{\mathrm{d}p_y(t)}{\mathrm{d}t} \qquad [53]$$

$$\omega_y(t)=-\Omega p_y(t)-\frac{\mathrm{d}p_x(t)}{\mathrm{d}t} \qquad [54]$$

$$\omega_z(t)=[1+p_z(t)]\Omega \qquad [55]$$

where the time rate of change of H has been set equal to $(1+p_z)\ \Omega$ where $p_z(t)=m_z(t)$ represents small departures from uniform spin at the mean sidereal rotation rate Ω of the Earth.

In complex notation, with $\mathbf{m}(t)=m_x(t)+im_y(t)$ and $\mathbf{p}(t)=p_x(t)-ip_y(t)$, where the negative sign accounts for $p_y(t)$ being positive toward 90°W longitude, eqns [53] and [54] can be written as

$$\mathbf{m}(t)=\mathbf{p}(t)-\frac{i}{\Omega}\frac{\mathrm{d}\mathbf{p}(t)}{\mathrm{d}t} \qquad [56]$$

For frequencies of motion $\sigma\ll\Omega$, the second term on the right-hand side of eqn [56] becomes much smaller than the first and the motion of the rotation pole becomes the same as the motion of the CIP. But for frequencies of motion $|\sigma|\approx\Omega$, the motions of the rotation and CIPs are very different. This difference becomes important when studying rapid motions such as those caused by the diurnal and semidiurnal ocean tides (see **Section 3.09.4.2.3**). For example, the amplitude of the prograde diurnal tide-induced motion of the rotation pole is about twice as large as that of the CIP.

Using eqns [53] and [54] for the relationship between the reported polar motion parameters $p_x(t)$ and $p_y(t)$ and the elements $\omega_x(t)$ and $\omega_y(t)$ of the Earth rotation vector, eqns [38] and [39] can be written in complex notation as (Brzezinski, 1992; Gross, 1992)

$$\mathbf{p}(t)+\frac{i}{\sigma_{\mathrm{o}}}\frac{\mathrm{d}\mathbf{p}(t)}{\mathrm{d}t}=\boldsymbol{\chi}(t) \qquad [57]$$

where $\boldsymbol{\chi}(t)=\boldsymbol{\chi}_x(t)+i\chi_y(t)$. The axial component, eqn [40], is usually written in terms of changes $\Delta\Lambda(t)$ of the length of day as

$$\frac{\Delta\Lambda(t)}{\Lambda_{\mathrm{o}}}=\chi_z(t) \qquad [58]$$

where Λ_{o} is the nominal length of day (LOD) of 86 400 s. Equations [57] and [58] are the final expressions, written in terms of the parameters actually reported by Earth rotation measurement services, for the changes in the rotation of the Earth caused by small excitation $\chi_i(t)$, where the excitation functions are given by eqns [41]–[43].

Since five angles are traditionally used to transform station coordinates between the terrestrial and celestial reference frames when only three are required, the five traditional Earth Orientation Parameters (EOPs) are not independent of each other. Because the frequency σ_c of some motion as observed in the celestial reference frame is related to the frequency σ_t of that same motion as observed in the terrestrial reference frame by

$$\sigma_c=\sigma_t+\Omega \qquad [59]$$

then motion having a retrograde nearly diurnal frequency in the terrestrial reference frame ($\sigma_t\approx-\Omega$) will be of low frequency (long period) when observed in the celestial reference frame. That is, retrograde nearly diurnal polar motions are equivalent to nutations. In particular, the two polar motion parameters are related to the two nutation parameters by (Brzezinski, 1992; Brzezinski and Capitaine, 1993)

$$\mathbf{p}(t)=-\mathbf{n}(t)\mathrm{e}^{-\mathrm{i}\Omega t} \qquad [60]$$

where Greenwich Mean Sidereal Time (GMST) has been approximated by Ωt in the exponent and $\mathbf{n}(t)=\delta\psi(t)\sin\varepsilon_o+i\,\delta\varepsilon(t)$ with ε_{o} being the mean obliquity of the ecliptic.

A degeneracy also exists between the EOPs and different realizations of the terrestrial reference frame. Since by eqn [49] the Earth rotation vector can be determined from the transformation matrix that transforms station coordinates between the terrestrial and celestial reference frames, if the realization of the terrestrial reference frame changes, then the elements of the rotation vector will change. In particular, a positive change in the x-component of polar motion is equivalent to a left-handed (clockwise) rotation of the terrestrial reference frame about the $\hat{\mathbf{y}}$-axis; a positive change in the y-component of polar motion, remembering that $p_y(t)$ is defined to be positive toward 90°W longitude, is equivalent to a left-handed rotation of the terrestrial reference frame about the $\hat{\mathbf{x}}$-axis; and a positive change in UT1 is equivalent to a right-handed (counter-clockwise) rotation of the terrestrial reference frame about the $\hat{\mathbf{z}}$-axis.

The polar motion parameters $p_x(t)$ and $p_y(t)$ give the location in the terrestrial frame of the reference pole of the intermediate frame, whatever that intermediate frame may be. The 1980 International Astronomical Union (IAU) theory of nutation adopted the celestial ephemeris pole (CEP) as the reference pole of the intermediate frame (Seidelmann, 1982), defining it to be a pole that exhibits no nearly diurnal motions either in the body-fixed terrestrial frame or in the celestial frame. The CEP was chosen (Seidelmann, 1982) to be the B-axis of Wahr (1981), which is the axis of figure for the Tisserand mean outer surface of the Earth, where the averaging procedure is such that the resulting B-axis does not move in response to body tides. Since observing stations are attached to the outer surface of the Earth, their measurements are sensitive to the motion of the Earth's surface in space. Wahr (1981) thus generalized the concept of the Tisserand mean-mantle frame to that of the Tisserand mean surface frame, with his B-axis being the reference axis of the Tisserand mean surface frame that moves in space with the mean motion of the observing stations.

In 2000, the IAU adopted the CIP as the reference pole of the intermediate frame (Petit and Luzum, 2010, **Chapter 1.06**). The definition of the CIP extends that of the CEP by clarifying the definition of polar motion and precession–nutation. The CEP was defined in such a manner that it exhibits no nearly diurnal motion in either the terrestrial or celestial reference frames. That is, precession–nutation were considered to be the motion of the CEP as viewed in the celestial reference frame with the frequency of motion ranging between −0.5 and +0.5 cycles per sidereal day (cpsd), and polar motion was considered to be the motion of the CEP as viewed in the terrestrial reference frame with the frequency of motion in that frame ranging between −0.5 and +0.5 cpsd (Capitaine, 2000). Since frequencies of motion in the two frames are related by eqn [59], in the celestial reference frame, polar motion was the motion of the CEP with frequencies ranging between +0.5 and +1.5 cpsd. Thus, the motion of the CEP in the celestial frame was defined for frequencies between −0.5 and +1.5 cpsd, with the division between polar motion and precession–nutation being at a frequency of +0.5 cpsd. The motion of the CEP outside this celestial frequency band was undefined. Similarly, the motion of the CEP in the terrestrial reference frame was defined for frequencies between −1.5 and +0.5 cpsd with the division between polar motion and precession–nutation being at a frequency of −0.5 cpsd. The motion of the CEP outside this terrestrial frequency band was also undefined (see **Figure 1**).

Since 1980, when the CEP was adopted as the reference pole of the intermediate frame, models of polar motion with frequencies outside the terrestrial frequency band within which the motion of the CEP was defined became available. These polar motions were due to the effects of diurnal and semidiurnal ocean tides. Models of nutations having frequencies outside the celestial frequency band within which the CEP was defined also became available, as did space-geodetic measurements having subdaily temporal resolution. With these improvements in models and measurements came the need to extend the definition of the CEP to all possible frequencies of motion, not only those between −0.5 and +1.5 cpsd in the celestial frame (−1.5 and +0.5 cpsd in the terrestrial frame). At the time that the definition of the intermediate pole was extended, its name was changed to the CIP. When defining the CIP, the concept used in defining the CEP of a pole having no nearly diurnal motions in either the terrestrial or celestial reference frames had to be abandoned because ocean tides can cause polar motions having frequencies near +1 cpsd as viewed in the terrestrial reference frame.

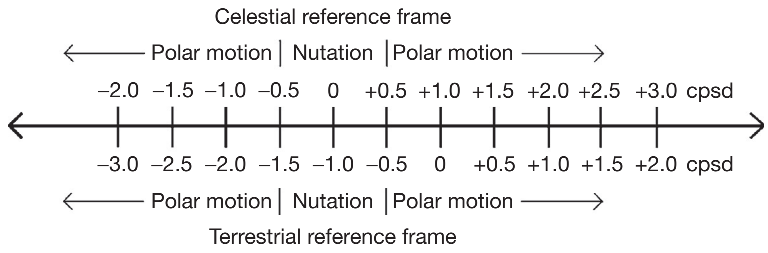

Figure 1 Schematic illustration of the relationship between the frequency of some motion as viewed in the celestial (top half of figure) and terrestrial (bottom half of figure) reference frames (see eqn [59]). By convention, nutation is the motion of the CIP having frequencies in the range between −0.5 and +0.5 cpsd as viewed in the celestial frame. As viewed in the terrestrial frame, this same nutational motion of the Earth has frequencies in the range between −1.5 and −0.5 cpsd. Motion at all other frequencies is considered to be polar motion.

The CIP is still chosen to be the axis of figure for the Tisserand mean outer surface of the Earth, as it was for the CEP. And the CIP is defined in such a manner that precession and nutation are still considered to be the motion of the CIP as viewed in the celestial reference frame with the frequency of motion ranging between −0.5 and +0.5 cpsd (Capitaine, 2000). But now polar motion is considered to be the motion of the CIP in the celestial frame at all other frequencies. That is, polar motion is considered to be the motion of the CIP in the terrestrial frame at all frequencies except those between −1.5 and −0.5 cpsd (see **Figure 1**). This has the effect of including in nutation those ocean tidal terms having retrograde nearly diurnal frequencies as viewed in the terrestrial reference frame and of including in polar motion those nutation terms having frequencies less than −0.5 cpsd or greater than +0.5 cpsd as viewed in the celestial reference frame (see **Table 2**).

3.09.3 Earth Rotation Measurement Techniques

Changes in the Earth's rate of rotation become apparent when comparing time kept by the rotating Earth, known as Universal Time (UT), to uniform timescales based either upon atomic clocks or upon the motion of the Sun and other celestial bodies. Prior to the development of atomic clocks, the most accurate measurements of changes in the Earth's rate of rotation were obtained by timing the occultations of stars by the Moon. With the advent of atomic clocks in 1955, a uniform atomic timescale became available that could be used as a reference when measuring the transit times of stars as they pass through the local meridian. Changes in the Earth's rate of rotation could then be determined more accurately from optical astrometric measurements of star transits than they could from measurements of lunar occultations. And prior to the development of space-geodetic techniques, optical astrometric measurements of changes in the apparent latitudes of observing stations yielded the most accurate estimates of polar motion. The space-geodetic techniques of very long-baseline interferometry (VLBI), global navigation satellite systems (GNSSs) like the Global Positioning System (GPS), and satellite laser ranging and lunar laser ranging (SLR and LLR, respectively) are now the most accurate techniques available for measuring changes both in the Earth's rate of rotation and in polar motion.

3.09.3.1 Lunar Occultation

The most recent rereduction of lunar occultation measurements for UT and LOD changes is that of Jordi et al. (1994) who analyzed about 53 000 observations of lunar occultations spanning 1830.0–1955.5. They used a reference frame defined by the FK5 star catalog, the LE200 lunar ephemeris, and corrections for the limb profile of the Moon. The UT series they obtained consists of values and 1σ uncertainties for the difference between Terrestrial Time (TT) and UT form 1 (TT–UT1) spanning 1830.0–1955.5 at 4-month intervals. TT is a dynamical timescale that can be related to the International Atomic Time (TAI) by adding 32.184 s to TAI (McCarthy and Guinot, 2012; McCarthy and Seidelmann, 2009; Seidelmann et al., 1992).

Jordi et al. (1994) extended their (TT–UT1) series to 1992 by using values of (UT1–TAI) obtained from the Bureau International de l'Heure (BIH) and the International Earth Rotation and Reference Systems Service (IERS). They then derived an LOD series spanning 1830–1987 at 4-month intervals by finite differencing and smoothing the extended UT1 series. Gross (2001) combined the lunar occultation measurements of Jordi et al. (1994) with optical astrometric and space-geodetic measurements to produce a smoothed LOD series spanning 1832.5–1997.5 at yearly intervals. Other UT1 and LOD series based upon lunar occultation measurements are those of Morrison (1979), Stephenson and Morrison (1984, 1985), McCarthy and Babcock (1986), and Liao and Greiner-Mai (1999).

3.09.3.2 Optical Astrometry

The International Latitude Service (ILS) was established by the International Association of Geodesy (IAG) in 1895 (e.g., Höpfner, 2000a) for the purpose of monitoring the wobbling motion of the Earth that had been detected by Seth Carlo Chandler, Jr., in 1891. As the Earth wobbles, the apparent latitude of an astronomical observing station will vary. To measure this variation of latitude and infer the underlying polar motion that is causing it, the ILS established six observing stations that were well distributed in longitude and that were all located at nearly the same latitude of 39°8′N. A seventh station, Kitab, was added in 1930 to replace the station at Tschardjui that ceased operations in 1919 due to a nearby river changing its course and adversely affecting the

Table 2 Coefficients of those nutation terms that are included in polar motion by the definition of the CIP

	Fundamental argument							$p_x(t)$ *(μas)*		$p_y(t)$ *(μas)*	
Degree of potential	γ	l	l′	F	D	Ω	*Period (solar days)*	*sin*	*cos*	*sin*	*cos*
2	1	1	0	0	0	0	0.96244	0.76	−0.43	0.43	0.76
2	1	0	0	0	0	−1	0.99712	1.93	−1.11	1.11	1.93
2	1	0	0	0	0	0	0.99727	14.27	−8.19	8.19	14.27
2	1	0	0	−2	2	−2	1.00275	−4.76	2.73	−2.73	−4.76
2	1	−1	0	0	0	0	1.03472	0.84	−0.48	0.48	0.84
2	1	0	0	−2	0	−2	1.07581	−11.36	6.52	−6.52	−11.36
2	1	0	0	−2	0	−1	1.07598	−2.14	1.23	−1.23	−2.14
2	1	1	0	−2	−2	−2	1.22346	−0.44	0.25	−0.25	−0.44
2	1	−1	0	−2	0	−2	1.11951	−2.31	1.32	−1.32	−2.31
2	1	−1	0	−2	0	−1	1.11970	−0.44	0.25	−0.25	−0.44
3	0	1	0	1	0	1	13.719	1.28	0.16	−0.16	1.28
3	0	0	0	1	0	0	27.212	2.62	0.32	−0.32	2.62
3	0	0	0	1	0	1	27.322	16.64	2.04	−2.04	16.64
3	0	0	0	1	0	2	27.432	−0.87	−0.11	0.11	−0.87
3	0	1	0	1	−2	1	193.560	2.10	0.27	−0.27	2.10
3	0	0	0	1	−1	1	365.242	1.31	0.20	−0.20	1.31
3	0	1	1	−1	0	−1	411.807	1.05	0.27	−0.27	1.05
3	0	1	1	−1	0	0	438.360	−0.63	0.12	−0.12	−0.63
3	0	−1	0	1	0	0	2190.35	−2.78	−0.31	0.31	−2.78
3	0	−1	0	1	0	1	3231.50	−16.16	−1.83	1.83	−16.16
3	0	−1	0	1	0	2	6159.14	0.78	0.09	−0.09	0.78
3	0	1	0	−1	0	−2	−6159.14	−0.68	−0.09	0.09	−0.68
3	0	1	0	−1	0	−1	−3231.50	12.32	1.59	−1.59	12.32
3	0	1	0	−1	0	0	−2190.35	1.86	0.24	−0.24	1.86
3	0	−1	0	−1	2	−1	−193.560	0.81	0.10	−0.10	0.81
3	0	0	0	−1	0	−2	−27.432	−0.82	−0.10	0.10	−0.82
3	0	0	0	−1	0	−1	−27.322	15.75	1.93	−1.93	15.75
3	0	0	0	−1	0	0	−27.212	2.48	0.30	−0.30	2.48
3	0	−1	0	−1	0	−1	−13.719	1.39	0.17	−0.17	1.39
Rate of secular polar motion (μas/year) due to the zero frequency tide											
4	0	0	0	0	0	0			−3.80		−4.31

Terms with amplitudes less than 0.5 microarcseconds (μas) are not tabulated. γ is GMST reckoned from the lower culmination of the vernal equinox (GMST + π). *l*, *l*′, *F*, *D*, and Ω are the Delaunay arguments, expressions for which are given in Simon et al. (1994). The period, given in solar days, is the approximate period of the term as viewed in the terrestrial reference frame. Terms having positive (negative) periods indicate prograde (retrograde) circular motion. Summing prograde and retrograde circular motions having the same period yields elliptical motion. The nearly diurnal terms, like the nearly diurnal and nearly semidiurnal ocean tidal terms, are not included in the polar motion parameters reported by Earth rotation measurement services. However, the secular rate and the long-period terms, like the long-period ocean tidal terms, are included in the reported polar motion parameters.

Source: Mathews PM and Bretagnon P (2003) Polar motions equivalent to high frequency nutations for a nonrigid Earth with anelastic mantle. *Astronomy and Astrophysics* 400: 1113–1128, with permission from Springer. Also, see Petit and Luzum (2010, Table 5.1a.)

seeing conditions at Tschardjui. Locating all the ILS stations at nearly the same latitude allowed common star pairs to be observed by the same Horrebow–Talcott method (Munk and MacDonald, 1960, **Chapter 1.09**), thereby allowing the polar motion to be determined from the latitude observations free of first-order errors in the reference star catalog.

The use of different star catalogs, standards, and data reduction procedures during the history of the ILS observing program can introduce discontinuities in the polar motion series derived from the ILS observations. In order to produce a homogeneous polar motion series unaffected by these sources of error, Yumi and Yokoyama (1980) rereduced 772 395 latitude observations taken at the seven ILS observing stations using the Melchior and Dejaiffe (1969) star catalog and the 1964 IAU System of Astronomical Constants. The resulting polar motion series, known as the homogeneous ILS series, spans from October 1899 to December 1978 at monthly intervals.

During the twentieth century, numerous other optical astrometric measurements of latitude and longitude were taken at other stations and by other methods besides those of the ILS. At the BIH, Li (1985) and Li and Feissel (1986) rereduced 240 140 optical astrometric measurements of longitude and 259 159 measurements of latitude taken at 136 observing stations. In order to produce an Earth orientation series independent of the ILS series, no measurements taken at the ILS stations were used. The resulting UT1 and polar motion series, which is known as the BIH series, spans from 5 January 1962 to 31 December 1981 at 5-day intervals.

The High-Precision Parallax Collecting Satellite (Hipparcos) star catalog, being constructed from observations taken above the Earth's atmosphere, is substantially more accurate than earlier ground-based catalogs such as the Melchior and Dejaiffe (1969) catalog used by Yumi and Yokoyama (1980). The improved accuracy of the Hipparcos star catalog motivated

Vondrák (1991, 1999) and Vondrák et al. (1992, 1995, 1997, 1998) to once again rereduce optical astrometric measurements for Earth orientation parameters. All available optical astrometric measurements, numbering 4 315 628 from 48 instruments including those taken at the ILS stations, were collected and corrected for instrumental errors and such systematic effects as plate tectonic motion, ocean loading, and tidal variations. The corrected measurements were then used to estimate nutation, polar motion, and UT. The resulting Earth orientation series, known as the Hipparcos series, consists of values and uncertainties for polar motion and nutation spanning 1899.7–1992.0 at quasi-5-day intervals, with UT1 estimates starting in 1956 shortly after atomic clocks first became available.

3.09.3.3 Space Geodesy

An integral part of geodesy has always been the definition and realization of a terrestrial, body-fixed reference frame, a celestial, space-fixed reference frame, and the determination of the EOPs (precession, nutation, spin, and polar motion) that link these two reference frames together. But with the advent of space geodesy – with the placement of laser retroreflectors on the Moon by Apollo astronauts and Soviet landers, the launch of the Laser Geodynamics Satellite (LAGEOS), the development of VLBI, and the development of GNSSs like GPS – a quantum leap has been taken in our ability to realize the terrestrial and celestial reference frames and to determine the Earth Orientation Parameters.

The only space-geodetic measurement technique capable of independently determining all of the EOPs is multibaseline VLBI (see Chapter 3.11). All of the other techniques either need to apply external constraints to the determined EOPs or can determine only subsets of the EOPs, only linear combinations of the EOPs, or only their time rates of change.

3.09.3.3.1 Very long-baseline interferometry

Radio interferometry is routinely used to make highly accurate measurements of UT1 and polar motion with observing sessions lasting from about an hour to a day. The VLBI technique measures the difference in the arrival time of a radio signal at two or more radio telescopes that are simultaneously observing the same distant extragalactic radio source (Lambeck, 1988, Chapter 1.08; Robertson, 1991; Sovers et al., 1998). This technique is therefore sensitive to processes that change the relative position of the radio telescopes with respect to the source, such as a change in the orientation of the Earth in space or a change in the position of the telescopes due to, for example, tidal displacements or tectonic motions. If just two telescopes are observing the same source, then only two components of the Earth's orientation can be determined. A rotation of the Earth about an axis parallel to the baseline connecting the two radio telescopes does not change the relative position of the telescopes with respect to the source, and hence, this component of the Earth's orientation is not determinable from VLBI observations taken on that single baseline. Multibaseline VLBI observations with satisfactory geometry can determine all of the components of the Earth's orientation including their time rates of change.

The International VLBI Service for Geodesy and Astrometry (IVS; Schlüter et al., 2002; Schlüter and Behrend, 2007), a service of both the IAG and the IAU, was established on 11 February 1999 to support research in geodesy, geophysics, and astrometry. As part of its activities, it coordinates the acquisition and reduction of VLBI observations for the purpose, in part, of monitoring changes in the Earth's rotation and defining and maintaining the international terrestrial and celestial reference frames. VLBI data products, including EOPs determined from both single and multibaseline observations, are available through the IVS website (see Table 3).

Table 3 Sources of Earth orientation data

Data type	*URL*
Very long-baseline interferometry	
IVS	http://ivscc.gsfc.nasa.gov/
Global navigation satellite system	
IGS	http://igscb.jpl.nasa.gov/
Satellite and lunar laser ranging	
ILRS	http://ilrs.gsfc.nasa.gov/
DORIS	
IDS	http://ids.cls.fr/
Intertechnique combinations	
IERS	http://www.iers.org/
KEOF	http://keof.jpl.nasa.gov/

URL, uniform resource locator; IVS, International VLBI Service for Geodesy and Astrometry; VLBI, very long-baseline interferometry; IGS, International GNSS Service; GNSS, global navigation satellite system; ILRS, International Laser Ranging Service; DORIS, Doppler Orbitography and Radiopositioning Integrated by Satellite; IDS, International DORIS Service; IERS, International Earth Rotation and Reference Systems Service; KEOF, Kalman Earth Orientation Filter.

3.09.3.3.2 Global navigation satellite system

GNSSs consist of two major elements: (1) a space-based element consisting of a constellation of transmitting satellites and (2) a ground-based element consisting of a network of receivers. In the United States' GPS, the satellites, including spares, are at altitudes of 20 200 km in orbits with periods of 11 h 58 min located in six orbital planes each inclined at 55° to the Earth's equator with four or more satellites in each plane. In the Global Navigation Satellite System (GLONASS) of Russia, the satellites are at altitudes of 19 100 km in circular orbits with periods of 11 h 15 min located in three orbital planes each inclined at 64.8° to the Earth's equator with eight satellites in each plane. In the Galileo system of Europe, which is not yet fully operational, the satellites will be at altitudes of 23 222 km in circular orbits with periods of 14 h 22 min located in three orbital planes each inclined at 56° to the Earth's equator with nine satellites and one spare in each plane.

In GNSSs, the navigation signals are broadcast by the satellites at more than one frequency, thereby enabling first-order corrections to be made for ionospheric refraction effects. The ground-based multichannel receivers detect the navigation signals being broadcast by those satellites that are above the horizon (up to the number of channels in the receiver). In principle, trilateration can then be used to determine the position of each receiver and, by extension, the orientation of the network of receivers as a whole. In practice, in order to achieve higher accuracy, more sophisticated analysis techniques are employed to determine the EOPs and other quantities such as orbital parameters of the satellites, positions of the stations, and atmospheric parameters such as the zenith path delay (Beutler et al., 1996; Blewitt, 1993, 2007; Bock and Leppard, 1990; Hofmann-Wellenhof et al., 2008; Leick, 2003).

Only polar motion and its time rate of change can be independently determined from GNSS measurements. UT1 cannot be separated from the orbital elements of the satellites and hence cannot be determined from GNSS data. The time rate of change of UT1, which is related to the length of day, can be determined from GNSS measurements. But because of the corrupting influence of orbit error, VLBI measurements are usually used to constrain the GNSS-derived LOD estimates.

The International GNSS Service (IGS; Beutler et al., 1999, 2009; Dow et al., 2005, 2009), a service of the IAG, was established on 1 January 1994 under its former name of the International GPS Service to support Earth science research. As part of its activities, it coordinates the acquisition and reduction of GNSS observations for the purpose, in part, of maintaining the International Terrestrial Reference Frame (ITRF) and monitoring changes in the Earth's rotation and geocenter motion. GNSS data products, including EOPs, are available through the IGS website (see **Table 3**).

3.09.3.3.3 Satellite and lunar laser ranging

In the technique of SLR, the round-trip flight times of laser light pulses are accurately measured as they are emitted from a laser system located at some ground-based observing station, travel through the Earth's atmosphere to some artificial satellite orbiting the Earth, are reflected by retroreflectors carried onboard that satellite, and return to the same observing station from which they were emitted (Combrinck, 2010; Exertier et al., 2006; Lambeck, 1988, **Chapter 1.07**). This time-of-flight range measurement is converted into a distance measurement by using the speed of light and correcting for a variety of known or modeled effects such as atmospheric path delay and satellite center-of-mass offset. Although a number of satellites carry retroreflectors for tracking and navigation purposes, the LAGEOS I and II satellites were specifically designed and launched to study geodetic properties of the Earth including its rotation and are the satellites most commonly used to determine the EOPs. Including range measurements to the Etalon I and II satellites has been found to strengthen the solution for the EOPs, so these satellites are now often included when determining the EOPs.

The EOPs are recovered from the basic range measurements in the course of determining the satellite's orbit. The basic range measurement is sensitive to any geophysical process that changes the distance between the satellite and the observing station, such as displacements of the satellite due to perturbations of the Earth's gravitational field, motions of the observing station due to tidal displacements or plate tectonics, or a change in the orientation of the Earth (which changes the location of the observing station with respect to the satellite). These and other geophysical processes must be modeled when fitting the satellite's orbit to the range measurements as obtained at a number of globally distributed tracking stations. Adjustments to the a priori models used for these effects can then be obtained during the orbit determination procedure, thereby enabling, for example, the determination of station positions and EOPs (Smith et al., 1985, 1990, 1994; Tapley et al., 1985, 1993). However, because variations in UT1 cannot be separated from variations in the orbital node of the satellite, which are caused by the effects of unmodeled forces acting on the satellite, it is not possible to independently determine UT1 from SLR measurements. But independent estimates of the time rate of change of UT1, or equivalently of LOD, can be determined from SLR measurements, as can polar motion and its time rate of change.

The technique of LLR is similar to that of SLR except that the laser retroreflector is located on the Moon instead of on an artificial satellite (Dickey et al., 1994a; Exertier et al., 2006; Lambeck, 1988, **Chapter 1.09**; Mulholland, 1980; Shelus, 2001; Williams et al., 1993, 2009). LLR is technically more challenging than SLR because of the need to detect the much weaker signal that is returned from the Moon. Larger, more powerful laser systems with more sophisticated signal detection systems need to be employed in LLR; consequently, there are far fewer stations that range to the Moon than range to artificial satellites. In fact, there are currently only three stations that are active in ranging to the Moon: the Observatoire de la Côte d'Azur in the south of France, the McDonald Observatory in Texas, and the Apache Point Observatory in New Mexico.

The EOPs are typically determined from LLR data by analyzing the residuals at each station after the lunar orbit and other parameters such as station and reflector locations have been fit to the range measurements (Dickey et al., 1985; Langley et al., 1981a; Stolz et al., 1976). From this single-station technique, two linear combinations of UT1 and the polar motion parameters $p_x(t)$ and $p_y(t)$ can be determined, namely, UT0 and the variation of latitude $\Delta\phi_i(t)$, at that station:

$$\Delta\phi_i(t) = p_x(t)\cos\lambda_i - p_y(t)\sin\lambda_i \qquad [61]$$

$$\mathbf{UT0}(\mathrm{t}) - \mathbf{TAI}(\mathrm{t}) = \mathbf{UT1}(\mathrm{t}) - \mathbf{TAI}(\mathrm{t}) + \left[p_x(t)\sin\lambda_\mathrm{i} + p_y(t)\cos\lambda_\mathrm{i}\right]\tan\phi_\mathrm{i} \qquad [62]$$

where ϕ_i and λ_i are the nominal latitude and longitude of station i. A rotation of the Earth about an axis connecting the station with the origin of the terrestrial reference frame does not change the distance between the station and the Moon, and hence, this component of the Earth's orientation cannot be determined from single-station LLR observations.

The International Laser Ranging Service (ILRS; Gurtner et al., 2005; Pearlman et al., 2002, 2005) is a service of the IAG that was established on 22 September 1998 to support research in geodesy, geophysics, and lunar science. As part of its activities, the ILRS coordinates the acquisition and reduction of SLR and LLR observations for the purpose, in part, of maintaining the ITRF and monitoring the Earth's rotation and geocenter motion. SLR and LLR data products, including EOPs, are available through the ILRS website (see **Table 3**).

3.09.3.3.4 Doppler Orbitography and Radiopositioning Integrated by Satellite

Like GNSSs, the French Doppler Orbitography and Radiopositioning Integrated by Satellite (DORIS) system also consists of space-based and ground-based elements (Tavernier et al., 2005, 2006; Willis et al., 2006, 2010). But, in the DORIS system, the transmitting beacons are located on the ground and the receivers are located on the satellites. Currently, there are about 60 globally distributed beacons broadcasting to receivers onboard six satellites (Jason-1, OSTM/Jason-2, CryoSat-2, HY-2A, SPOT-4, and SPOT-5). In the past, the SPOT-2, SPOT-3, TOPEX/Poseidon, and Envisat satellites carried DORIS receivers. And current plans call for the future SARAL, Sentinel-3A, Sentinel-3B, and Jason-3 satellites to carry DORIS receivers.

In the DORIS system, the Doppler shift of two transmitted frequencies is accurately measured. The use of two frequencies allows corrections to be made for ionospheric effects. Processing these Doppler measurements, usually as range-rate (Tapley et al., 2004a), allows the orbit of the satellite to be determined along with other quantities such as station positions and EOPs. As with other satellite techniques, UT1 cannot be determined from DORIS measurements, but its time rate of change can be determined, as can polar motion and its rate of change (Gambis, 2006; Willis et al., 2006).

The International DORIS Service (IDS; Tavernier et al., 2005, 2006; Willis et al., 2010) is a service of the IAG that was established on 1 July 2003 to support research in geodesy and geophysics. As part of its activities, the IDS coordinates the acquisition and reduction of DORIS observations for the purpose, in part, of maintaining the international terrestrial reference frame and monitoring the Earth's rotation and geocenter motion. DORIS data products, including EOPs, are available through the IDS website (see **Table 3**).

3.09.3.4 Ring Laser Gyroscope

Ring laser gyroscopes are a promising emerging technology for determining the Earth's rotation. In a ring laser gyroscope, two laser beams propagate in opposite directions around a ring. Since the ring laser gyroscope is rotating with the Earth, the effective path length of the beam that is corotating with the Earth is slightly longer than the path that is counterrotating with it. Because the effective path lengths of the two beams differ, their frequencies differ, so they interfere with each other to produce a beat pattern (Stedman, 1997). The beat frequency $\delta f(t)$, which can be measured and is known as the Sagnac frequency, is proportional to the instantaneous angular velocity $\boldsymbol{\omega}(t)$ of the Earth:

$$\delta f(t) = \frac{4A}{\lambda P}\mathbf{n}\cdot\boldsymbol{\omega}(t) \qquad [63]$$

where A is the area of the ring, P is the optical path length, λ is the wavelength of the laser beam in the absence of rotation, and $\mathbf{n}$ is the unit vector normal to the plane of the ring. By making the instrument rigid and the laser stable so that A, P, and λ are constants, the component of the Earth's instantaneous rotation vector that is parallel to the normal of the plane of the ring can be determined from measurements of the Sagnac frequency. All three components of the Earth rotation vector are determinable from three mutually orthogonal co-located ring laser gyroscopes or from a globally distributed network of gyroscopes.

Ring laser gyroscopes measure the absolute rotation of the Earth in the sense that, in principle, just a single measurement of the Sagnac frequency is required to determine $\mathbf{n}\cdot\boldsymbol{\omega}$. All of the other techniques discussed in the preceding text are relative sensors because they infer the Earth's rotation from the change in the orientation of the Earth that takes place between at least two measurements that are separated in time. Thus, ring laser gyroscopes are fundamentally different from the space-geodetic techniques that are currently used to regularly monitor the Earth's rotation. Measurements taken by ring laser gyroscopes can therefore be expected to contribute to our understanding of the Earth's changing rotation, on subdaily to daily timescales (Mendes Cerveira et al., 2009; Nilsson et al., 2012; Schreiber et al., 2004, 2009) and on longer timescales (Schreiber et al., 2011).

3.09.3.5 Intertechnique Combinations

EOPs can be determined from measurements taken by each of the techniques discussed in the preceding text. But each technique has its own unique strengths and weaknesses in this regard. Not only is each technique sensitive to a different subset and/or linear combination of the EOPs, but also the averaging time for their determination is different, as is the interval between observations, the precision with which they can be determined, and the duration of the resulting EOP series. By combining the individual series determined by each technique, a series of the Earth's orientation can be obtained that is based upon independent measurements and that spans the greatest possible time interval. Such a combined Earth orientation series is useful for a number of purposes, including a variety of scientific studies and as an a priori series for use in data reduction procedures. However, care must be taken when generating such a combined series to account for differences in the underlying reference frames within which each individual series is determined (which can lead to differences in the bias and rate of the Earth orientation series). Care must also be taken to properly assign weights to the observations prior to combination.

The IERS, a service of both the IAG and the IAU, was established on 1 January 1988 under its former name of the International Earth Rotation Service to serve the astronomical, geophysical, and geodetic communities by, in part, providing the International Celestial Reference Frame (ICRF), the ITRF, and the EOPs that are needed to transform station coordinates between these two frames (Dick and Richter, 2004; Feissel and Gambis, 1993; Vondrák and Richter, 2004). As part of its activities, the IERS combines and predicts EOPs determined by the space-geodetic techniques using a weighted-average approach (Bizouard and Gambis, 2009; Gambis, 2004; Gambis and Luzum, 2011; Johnson et al., 2005; Luzum et al., 2001). The EOPs produced by the IERS are available through its website (see **Table 3**).

Since the 1980s, a Kalman filter, the Kalman Earth Orientation Filter (KEOF), has been used at the Jet Propulsion Laboratory (JPL) to combine and predict EOPs in support of interplanetary spacecraft tracking and navigation (Freedman et al., 1994a; Gross et al., 1998). A Kalman filter has many properties that make it an attractive method of combining Earth orientation series. It allows the full accuracy of the measurements to be used, whether they are degenerate or full rank, are irregularly or regularly spaced in time, or are corrupted by systematic or other errors that can be described by stochastic models. And by using a stochastic model for the process, a Kalman filter can objectively account for the growth in the uncertainty of the EOPs between measurements. The combined and predicted Earth orientation series produced at JPL using a Kalman filter are available through the KEOF website (see Table 3).

3.09.4 Observed and Modeled Earth Rotation Variations

The Earth's rotation changes on all observable timescales, from subdaily to decadal and longer. The wide range of timescales on which the Earth's rotation changes reflects the wide variety of processes that are causing it to change, including external tidal forces; surficial fluid processes involving the atmosphere, oceans, and hydrosphere; and internal processes acting both within the solid Earth itself and between the fluid core and solid Earth. The changes in the Earth's rotation that are observed and the geophysical models that have been developed to explain them are reviewed here. Sources of Earth rotation data are given in Table 3 and sources of geophysical models are given in Table 4.

Two general approaches are used to investigate why and how the Earth's rotation changes: the torque approach and the angular momentum approach (Munk and MacDonald, 1960; Wahr, 1982). In the torque approach, changes in the Earth's rotation are studied by computing the torques acting on the Earth's crust and mantle by, for example, processes due to the overlying atmosphere and oceans (de Viron and Dehant, 1999; de Viron et al., 1999, 2001a,b; Marcus et al., 2004). In the angular momentum approach, changes in the Earth's rotation are studied by invoking the principle of the conservation of angular momentum. Under this principle, if the angular momentum of, say, the atmosphere changes, then the angular momentum of the crust and mantle must change by an equal but opposite amount in order for the total angular momentum of the solid Earth–atmosphere system to remain the same.

The advantage of the torque approach is that it can be used to elucidate the mechanism by which angular momentum is exchanged between, say, the atmosphere and the solid Earth. In general, there are three different types of torques that act to exchange angular momentum between a surface geophysical fluid like the atmosphere and the solid Earth: (1) a pressure or mountain torque due to normal pressure forces acting on topographic features of the solid Earth, (2) a friction torque due to tangential frictional stresses caused by relative motion between a fluid like the atmosphere and the solid Earth, and (3) a gravitational torque due to gravitational attraction between density heterogeneities in the fluid and solid Earth. For example, Salstein and Rosen (1994) showed that mountain torques, and in particular those over South America, were largely responsible for the rapid exchange of axial angular momentum between the atmosphere and solid Earth that occurred at the beginning of August 1992.

The disadvantage of the torque approach is that the currently available atmospheric, oceanic, and hydrologic models are not suited for the computation of torques. Many parameters of the models, such as friction drag, that are needed to compute the torques are not known very well. And the grid scales of the models are too coarse to allow accurate computation of the torques (de Viron and Dehant, 2003). However, the models are well suited for the computation of angular momentum. So the angular momentum approach is most often used to investigate the causes of the observed changes in the Earth's rotation.

3.09.4.1 UT1 and LOD Variations

Observations of the length of day (Figure 2) show that it consists mainly of (1) a linear trend of rate +1.8 ms cy^{-1}, (2) decadal variations having an amplitude of a few milliseconds, (3) tidal variations having an amplitude of about 1 ms, (4) seasonal variations having an amplitude of about 0.5 ms, and

Table 4 Sources of angular momentum models

Model type	*URL*
Global geophysical fluids	
IERS GGFC	http://geophy.uni.lu/
Atmospheric	
IERS SBA	http://www.aer.com/science-research/earth/earth-mass-and-rotation/special-bureau-atmosphere
TU Vienna	http://ggosatm.hg.tuwien.ac.at/
Oceanic	
IERS SBO	http://euler.jpl.nasa.gov/sbo/
Hydrologic	
IERS SBH	http://www.csr.utexas.edu/research/ggfc/
Consistent AAM, *OAM*, and *HAM*	
GFZ Potsdam	http://www.gfz-potsdam.de/portal/gfz/Struktur/Departments/Department+1/sec13/services

URL, uniform resource locator; IERS, International Earth Rotation and Reference Systems Service; GGFC, Global Geophysical Fluids Center; SBA, Special Bureau for the Atmosphere; TU, Technical University; SBO, Special Bureau for the Oceans; SBH, Special Bureau for Hydrology; AAM, atmospheric angular momentum; OAM, ocean angular momentum; HAM, hydrological angular momentum; GFZ, GeoForschungsZentrum.

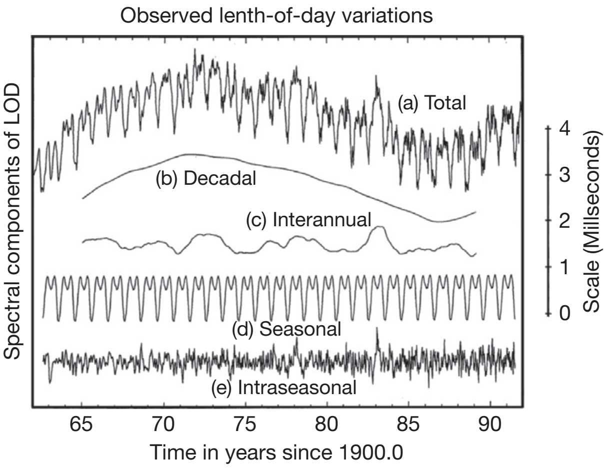

Figure 2 (a) Observed LOD variations during 1962–91 from the COMB91 combined optical astrometric and space-geodetic Earth orientation series and its decomposition into variations on (b) decadal, (c) interannual, (d) seasonal, and (e) intraseasonal timescales. Tidal variations are not shown. Reproduced from Dickey JO, Marcus SL, Hide R, Eubanks TM, and Boggs DH (1994b) Angular momentum exchange among the solid Earth, atmosphere and oceans: A case study of the 1982–83 El Niño event. *Journal of Geophysical Research* 99(B12): 23921–23937.

(5) smaller amplitude variations occurring on all measurable timescales. Here, the LOD variations that are observed and the models that have been developed to explain them are reviewed.

The length of day is the rotational period of the Earth. Changes $\Delta\Lambda(t)$ in the length of day are related to changes in UT and to changes $\Delta\omega_z(t) = \Omega m_z(t)$ in the axial component of the Earth's angular velocity by

$$\frac{\Delta\Lambda(t)}{\Lambda_o} = -m_z(t) = -\frac{\mathrm{d}(\mathrm{UT1} - \mathrm{TAI})}{\mathrm{d}t} \quad [64]$$

By eqns [43] and [58], the length of day will change as a result of changes in the axial component $h_z(t)$ of relative angular momentum and of changes in the axial component $\Omega\Delta I_{zz}(t)$ of angular momentum due to changes in the mass distribution of the Earth. Modeling the observed changes in LOD therefore requires computing both types of changes in angular momentum for the different components of the Earth system. By definition, the angular momentum vector $\mathbf{L}(t)$ of some component of the Earth system such as the atmosphere or oceans is given by

$$\mathbf{L}(t) = \int_{V(t)} \rho(\mathbf{r}, t)\mathbf{r} \times [\boldsymbol{\omega} \times \mathbf{r} + \mathbf{u}(\mathbf{r}, t)]\,\mathrm{d}V \quad [65]$$

where the integral is taken over the, in general, time-dependent volume $V(t)$ of that component of the Earth system under consideration, the first term in the square brackets represents changes in the angular momentum due to changes in mass distribution, the second term represents changes due to relative motion, and $\mathbf{r}$ is the position vector of some mass element of density $\rho(\mathbf{r},t)$ that is moving with the Eulerian velocity $\mathbf{u}(\mathbf{r},t)$ with respect to a terrestrial reference frame that (1) is fixed to the solid Earth; (2) is oriented such that its x- and y-coordinate axes are along the Greenwich and 90°E meridians, respectively, with both axes lying in the equatorial plane; (3) has an origin located at the center of the Earth; and (4) is rotating with angular velocity $\boldsymbol{\omega}$ with respect to an inertial reference frame. In general, $\boldsymbol{\omega}$ is not constant in time but exhibits variations in both magnitude (related to changes in LOD) and direction (related to polar motion). However, these changes are small, and for the purpose of deriving expressions for the angular momentum of different components of the Earth system, it can be assumed that the terrestrial reference frame is uniformly rotating at the rate Ω about the z-coordinate axis: $\boldsymbol{\omega} = \Omega\hat{\mathbf{z}}$. In this case, the axial component of the mass term of the angular momentum can be written as

$$\Omega\Delta I_{zz}(t) = \Omega\int_{V(t)} \rho(\mathbf{r}, t)r^2\cos^2\phi\,\mathrm{d}V \quad [66]$$

where ϕ is the north latitude. Similarly, the axial component of the motion term of the angular momentum can be written as

$$h_z(t) = \int_{V(t)} \rho(\mathbf{r}, t)r\cos\phi\, u(\mathbf{r}, t)\,\mathrm{d}V \quad [67]$$

where $u(\mathbf{r},t)$ is the eastward component of the velocity.

3.09.4.1.1 Secular trend, tidal dissipation, and glacial isostatic adjustment

Tidal dissipation causes the Earth's angular velocity and hence rotational angular momentum to decrease. Since the angular momentum of the Earth–Moon system is conserved, the orbital angular momentum of the Moon must increase to balance the decrease in the Earth's rotational angular momentum. The increase in the orbital angular momentum of the Moon is accomplished by an increase in the radius of the Moon's orbit and a decrease in the Moon's orbital angular velocity. But early observations of the Moon's position showed that it was apparently accelerating, not decelerating, in its orbit. This apparent acceleration of the Moon was a result of assuming that the Earth is rotating with a constant, rather than decreasing, angular velocity when predicting the Moon's position. If the Earth's angular velocity is actually decreasing but is assumed to be constant when predicting the position of the Moon, then the observed position of the Moon will appear to be ahead of its predicted position. That is, the Moon will appear to be accelerating in its orbit. That the Moon is apparently accelerating in its orbit was first noted by Halley (1695). But it was not until 1939 that Spencer Jones (1939) was able to conclusively demonstrate that the angular velocity of the Earth is actually decreasing and that the apparent acceleration of the Moon in its orbit was an artifact of assuming that the angular velocity of the Earth was constant.

Halley (1695) also seemed to have been the first to appreciate the importance of ancient and medieval records of lunar and solar eclipses for determining the apparent acceleration of the Moon and the corresponding decrease in the angular velocity of the Earth over the past few thousand years. The change in the Earth's rate of rotation can be deduced from the discrepancy between when and where eclipses should have been observed if the angular velocity of the Earth were constant to when and where they were actually observed as recorded in Babylonian clay tablets and Chinese, European, and Arabic books and manuscripts (Stephenson, 1997, 1998; Stephenson and Morrison, 2005).

When using eclipse observations to deduce the secular change in the length of day over the past few thousand years, the positions of the Sun and Moon must be accurately known. Of primary importance in this regard is the value for the tidal acceleration $\dot{n}$ of the Moon since it controls the long-term behavior of the Moon's motion. The tidal acceleration of the Moon can be determined from observations of the timings of transits of Mercury (e.g., Morrison and Ward, 1975; Spencer Jones, 1939) and from satellite and lunar laser ranging measurements. Tidal forces distort the figure of the Earth and hence its gravitational field, which in turn perturbs the orbits of artificial satellites. SLR measurements can detect these tidal perturbations in the satellites' orbits and can therefore be used to construct tide models and hence determine the tidal acceleration of the Moon. Using this approach, Christodoulidis et al. (1988) reported a value of -25.27 ± 0.61 arcseconds per century^2 for the tidal acceleration $\dot{n}$ of the Moon due to dissipation by solid Earth and ocean tides. Other SLR-derived values for $\dot{n}$ have been reported by Cheng et al. (1990, 1992), Marsh et al. (1990, 1991), Dickman (1994), Lerch et al. (1994), and Ray (1994). Models of the solid Earth and ocean tides have been used to compute the concomitant secular increase in LOD by Ray et al. (1999), Wu et al. (2003), and Mathews and Lambert (2009).

Like the orbits of artificial satellites, the orbit of the Moon is also perturbed by tidal forces. Since LLR measurements can detect tidal perturbations in the Moon's orbit, they can be used to determine the tidal acceleration of the Moon. In addition to being sensitive to orbital perturbations caused by tides on the Earth, LLR measurements, unlike SLR measurements, are also sensitive to orbital perturbations caused by tides on the Moon. Using LLR measurements, Williams et al. (2001) reported a value of -25.73 ± 0.5 arcseconds per century^2 for the tidal acceleration of the Moon, which by Kepler's law corresponds to an increase of 3.79 ± 0.07 cm year^{-1} in the semimajor axis of the Moon's orbit and which includes a contribution of +0.29 arcseconds per century^2 from dissipation within the Moon itself. Only about half of the discrepancy between the SLR- and LLR-derived values for $\dot{n}$ due to dissipation by just tides on the Earth is currently understood (Williams et al., 2001). Other LLR-derived values for $\dot{n}$ have been reported by Newhall et al. (1988), Dickey et al. (1994a), and Chapront et al. (2002).

By a priori adopting a value for the tidal acceleration $\dot{n}$ of the Moon, lunar and solar eclipse observations can be used to determine the secular increase in the length of day over the past few thousand years. The most recent rereductions of lunar and solar eclipse observations for LOD changes are those of Stephenson and Morrison (1995) and Morrison and Stephenson (2001). Besides using eclipse observations spanning from 700 BC to AD 1600, they also used lunar occultation observations spanning 1600–1955.5 and optical astrometric and space-geodetic measurements spanning 1955.5–90. Adopting a value of -26.0 arcseconds per century^2 for $\dot{n}$, Morrison and Stephenson (2001) found that the length of day has increased at a rate of $+1.80 \pm 0.1$ ms cy^{-1} on average during the past 2700 years (see **Figure 3**). In addition to a secular trend, Stephenson and Morrison (1995) and Morrison and Stephenson (2001) also found evidence for a fluctuation in the length of day that has a peak-to-peak amplitude of about 8 ms and a period of about 1500 years (**Figure 3**), although Dalmau (1997) had questioned its existence.

By the conservation of angular momentum, a tidal acceleration of the Moon of -26.0 arcseconds per century^2 should be accompanied by a +2.3 ms cy^{-1} increase in the length of day (Stephenson and Morrison, 1995). Since the observed increase in the length of day is only +1.8 ms cy^{-1} (Morrison and Stephenson, 2001), some other mechanism or combination of mechanisms must be acting to change the length of day by -0.5 ms cy^{-1}. By eqns [43] and [58], changes both in the axial component of relative angular momentum and in the polar moment of inertia of the Earth can cause LOD to change. A secular trend in the general circulation of fluids like the atmosphere and oceans, and hence in atmospheric and oceanic angular momentum, is unlikely to be sustained over the course of a few thousand years. In fact, using results from a 100-year run of the Hadley Centre general circulation model of the atmosphere, de Viron et al. (2004b) found that the modeled secular trend in atmospheric angular momentum (AAM) during 1870–1997 causes a secular trend in LOD of only +0.08 ms cy^{-1}.

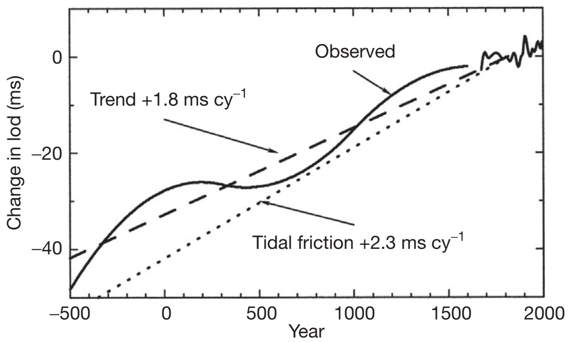

Figure 3 Secular change in the length of day during the past 2500 years estimated from lunar and solar eclipse, lunar occultation, optical astrometric, and space-geodetic observations. The difference between the observed secular trend and that caused by tidal friction is due to the effects of glacial isostatic adjustment and other processes such as ice sheet mass change and the accompanying nonsteric change in sea level. Reprinted from Journal of Geodynamics, 32, Morrison LV and Stephenson FR, Historical eclipses and the variability of the Earth's rotation. 247–265. Copyright (2001), with permission from Elsevier.

One of the most important mechanisms acting to cause a secular trend in LOD on timescales of a few thousand years is glacial isostatic adjustment (GIA). The isostatic adjustment of the solid Earth in response to the decreasing load on it following the last deglaciation causes the figure of the Earth to change and hence LOD to change. Since the solid Earth is rebounding in the regions at high latitude where the ice load was formerly located, the figure of the Earth is becoming less oblate, the Earth's rotation is accelerating, and LOD is decreasing. Models of GIA show that its effect on LOD is very sensitive to the assumed value of lower mantle viscosity (e.g., Johnston and Lambeck, 1999; Mitrovica and Milne, 1998; Mitrovica et al., 2006; Peltier and Drummond, 2010; Peltier and Jiang, 1996; Peltier and Luthcke, 2009; Peltier and Wu, 1983; Sabadini and Peltier, 1981; Sabadini and Vermeersen, 2004; Sabadini et al., 1982; Tamisea et al., 2002; Vermeersen et al., 1997; Wu and Peltier, 1984; Yuen et al., 1982). But by deriving a model for the radial viscosity profile of the Earth that fits both postglacial decay times and free-air gravity anomalies associated with mantle convection, Mitrovica and Forte (1997) found that GIA should cause a secular trend in LOD amounting to -0.5 ms cy^{-1}, a value in remarkable agreement with that needed to explain the difference between the observed secular trend in LOD and that caused by tidal dissipation.

However, GIA is not the only mechanism that will cause a secular trend in LOD. The present-day change in glacier and ice sheet mass and the accompanying change in nonsteric sea level will also cause a secular trend in LOD (e.g., James and Ivins, 1995, 1997; Mitrovica and Peltier, 1993; Mitrovica et al., 2006; Nakada and Okuno, 2003; Peltier, 1988; Tosi et al., 2005; Trupin, 1993; Trupin et al., 1992). But the effect of this mechanism on LOD is very sensitive to the still unknown present-day mass change of glaciers and ice sheets, particularly of the Antarctic ice sheet. By adopting various scenarios for the mass change of Antarctica, models predict that its mass change alone should cause a secular trend in LOD ranging anywhere from -0.72 ms cy^{-1} to +0.31 ms cy^{-1} (James and Ivins, 1997).

Other mechanisms that may cause a linear trend in LOD over the last or next century include continental drift (Dickman, 1979), tectonic processes taking place under non-isostatic conditions (Sabadini and Vermeersen, 2004; Vermeersen and Vlaar, 1993; Vermeersen et al., 1994), plate subduction (Alfonsi and Spada, 1998; Greff-Lefftz, 2011; Ricard et al., 1993; Spada et al., 1992), mantle convection (Greff-Lefftz, 2011), upwelling mantle plumes (Greff-Lefftz, 2011), deformation of the mantle caused by pressure variations acting at the core–mantle boundary that are associated with the motion of the fluid core (Dumberry and Bloxham, 2004; Fang et al., 1996; Greff-Lefftz et al., 2004), earthquakes (Chao and Gross, 1987; Gross and Chao, 2006), and climate change (Abarca del Rio, 1999; de Viron et al., 2002; Huang et al., 2001; Landerer et al., 2007; Ponte et al., 2002; Räisänen, 2003; Rosen and Gutowski Jr., 1992; Rosen and Salstein, 2000; Winkelnkemper et al., 2009).

The fluctuation in LOD of 1500-year period found by Stephenson and Morrison (1995) and Morrison and Stephenson (2001) is currently of unknown origin. However, given its large amplitude, which is too large to be caused by atmospheric and oceanic processes but which is comparable in size to the amplitude of the decadal variations, it is probably caused (Dumberry and Bloxham, 2006) by the same core–mantle interactions, such as gravitational coupling (Rubincam, 2003), that are known to cause decadal variations in LOD.

3.09.4.1.2 Decadal variations and core–mantle interactions

While lunar and solar eclipse observations are valuable for studying the secular trend in LOD, they are too sparse and inaccurate to reveal decadal variations. Instead, lunar occultation observations, which are available since the early 1600s (Martin, 1969; Morrison et al., 1981), are used to study decadal variations in LOD. **Figure 4** compares three different LOD series derived from lunar occultation observations. Gross (2001) (black curve) derived an LOD series spanning 1832.5–1997.5 at yearly intervals by analyzing UT1 measurements obtained from lunar occultation observations by Jordi et al. (1994), from the Hipparcos optical astrometric series of Vondrák (1991, 1999) and Vondrák et al. (1992, 1995, 1997, 1998), and from the COMB97 combined optical astrometric and space-geodetic series of Gross (2000a). McCarthy and Babcock (1986) (red curve) derived an LOD series spanning 1657.0–1984.5 at half-yearly intervals by analyzing UT1 measurements obtained from lunar occultation observations by Martin (1969) and Morrison (1979), from the optical astrometric series of McCarthy (1976), and from a series obtained from the BIH. Stephenson and Morrison (1984) (green curve) obtained an LOD series spanning 1630–1980 at 5-year intervals before 1780 and at yearly intervals afterward by analyzing the lunar occultation and solar eclipse observations cataloged by Morrison (1978) and Morrison et al. (1981), combining them with an LOD series derived from optical astrometric measurements of UT1 obtained from the BIH. As can be seen from **Figure 4**, during their common time span, these three different LOD series are consistent with each other to within the 1σ standard error of the Gross (2001) series. Other LOD series that have been derived from lunar occultation observations are those of Morrison (1979), Jordi et al. (1994), and Liao and Greiner-Mai (1999).

The peak-to-peak variation in LOD seen in **Figure 4** is about 7 ms, a variation far too large to be caused by atmospheric and oceanic processes (Gross et al., 2005). That atmospheric processes cannot cause such large decadal-scale LOD variations can be easily demonstrated by considering the change in LOD that would be caused if the motion of the atmosphere were to stop entirely. Because of the pole-to-equator temperature gradient, the atmosphere superrotates with respect to the solid Earth at an average rate of about 7 m s^{-1}. If the superrotation of the atmosphere were to stop so that the atmosphere just passively corotates with the solid Earth, then by conservation of angular momentum, the length of day would decrease by about 3 ms (Hide et al., 1980). Since stopping the atmospheric motion entirely does not cause an LOD change as large as that observed in **Figure 4**, some other mechanism or combination of mechanisms must be acting to cause the large decadal-scale LOD variations that are observed.

The most important mechanism acting to cause decadal variations in LOD is core–mantle coupling. While it has been recognized for quite some time that the core is the only viable

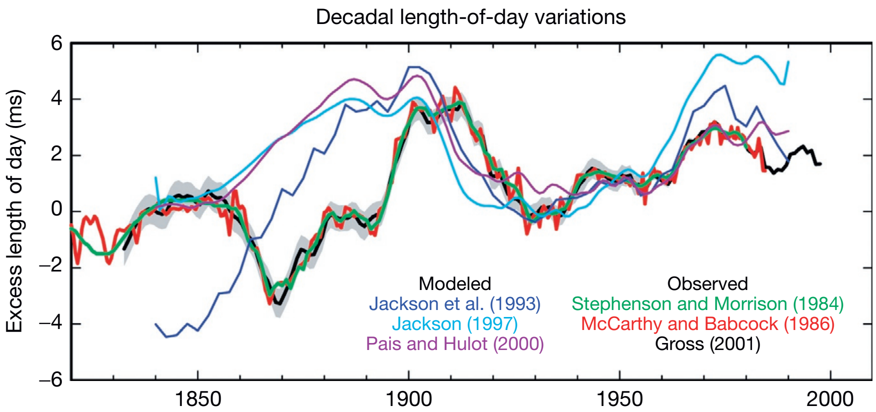

Figure 4 Plots of observed and modeled LOD variations on decadal timescales. The observed LOD series are those of Gross (2001) (black curve with gray shading representing $\pm 1\sigma$ standard error), McCarthy and Babcock (1986) (red curve), and Stephenson and Morrison (1984) (green curve). Note that after about 1955, the uncertainties of the Gross (2001) LOD values are less than the width of the black line. The modeled core angular momentum series are those of Jackson et al. (1993) (dark blue curve), the *uvm-s* model of Jackson (1997) (light blue curve), and the PH-inversion model of Pais and Hulot (2000) (purple curve). A secular trend of +1.8 ms cy^{-1} has been added to the modeled core angular momentum series to match the observed secular trend (see **Section 3.09.4.1.1**). An arbitrary bias has also been added to the modeled series in order to facilitate their comparison with the observed series. Reprinted from Physics of the Earth and Planetary Interiors, 123, Gross RS, A combined length-of-day series spanning 1832–1997: LUNAR97, 65–76. Copyright (2001), with permission from Elsevier.

source of the large decadal LOD variations that are observed (e.g., Lambeck, 1980; Munk and MacDonald, 1960), it was not until 1988 that Jault et al. (1988) were able to model the core angular momentum (CAM) and show that it causes decadal LOD variations that agree reasonably well with those observed.

The flow of the fluid at the top of the core can be inferred from surface observations of the secular variation of the magnetic field by assuming that (1) the mantle is a perfect insulator so that the magnetic field can be expressed as the gradient of a potential, which facilitates the downward continuation of the magnetic field to the top of the core; (2) the core is a perfect conductor so that the magnetic field is 'frozen' into the core fluid and is thus advected by the horizontal flow at the top of the core; (3) the flow at the top of the core is tangentially geostrophic so that it is governed by the balance between the horizontal components of just the pressure gradient and the Coriolis force at the top of the core (Bloxham and Jackson, 1991; Whaler and Davis, 1997); and (4) the flow is large scale. The last two assumptions are required in order to reduce the inherent nonuniqueness of core surface flow determinations. Other assumptions about the core surface flow fields that have been made (Finlay et al., 2010; Holme, 2007) are that the flow is purely toroidal so that it has no radial component (Bloxham, 1990; Whaler, 1980, 1986), that the flow is steady in time (Gubbins, 1982; Voorhies, 1986; Voorhies and Backus, 1985), that the flow is steady within a drifting reference frame (Davis and Whaler, 1996; Holme and Whaler, 2001), that the flow includes a helical component (Amit and Olson, 2004, 2006), that the flow is tangentially magnetostrophic (Asari and Lesur, 2011), or that the flow is quasi-geostrophic (Canet et al., 2009; Gillet et al., 2009; Pais and Jault, 2008).

While surface magnetic field observations can be used to infer the flow at the top of the core, the flow everywhere within the core must be known in order to compute the angular momentum of the core and hence the effect of core motion on the length of day. Jault et al. (1988) realized from dynamical considerations that for geostrophic flows on decadal timescales, the axisymmetric, or zonal, component of the core flow can be described by relative motion of nested cylinders that are coaxial with the rotation axis. This model for the motion of the core at depth allows the axial angular momentum of the core, and hence the effect of core motion on LOD, to be computed from the flow fields at the top of the core that are inferred from surface magnetic field observations (Amit and Olson, 2006; Asari et al., 2009; Gillet et al., 2009; Hide et al., 2000; Holme and Olsen, 2006; Holme and Whaler, 2001; Jackson, 1997; Jackson et al., 1993; Jault et al., 1988; Olsen and Mandea, 2008; Pais and Hulot, 2000; Pais and Jault, 2008; Pais et al., 2004; Wardinski and Lesur, 2012; Wardinski et al., 2008; Whaler and Holme, 2011). **Figure 4** compares the observed decadal LOD variations with the modeled results obtained by Jackson et al. (1993), Jackson (1997), and Pais and Hulot (2000). As can be seen, while the agreement is not perfect, CAM models produce decadal LOD variations that have about the same amplitude and phase as those observed.

In the coaxial nested cylinder model of the zonal core flow, the rotation of adjacent cylinders is coupled because of the magnetic field lines that thread through them. If for some reason the rotation of the cylinders is disturbed, then the magnetic field provides a restoring force that will cause the cylinders to oscillate. As first proposed by Braginsky (1970), it is the exchange with the mantle of the angular momentum associated with these torsional oscillations of the fluid core that causes the length of day to change (Buffett, 1998; Buffett and Mound, 2005; Buffett et al., 2009; Dickey and de Viron, 2009; Hide et al., 2000; Mound and Buffett, 2005, 2007; Pais and Hulot, 2000; Zatman and Bloxham, 1997, 1998, 1999).

While the exchange of CAM with the solid Earth can clearly cause decadal LOD variations of approximately the right amplitude and phase, the mechanism or mechanisms by which the angular momentum is exchanged between the core and solid Earth are less certain. Possible core–mantle coupling mechanisms are viscous torques, topographic torques, electromagnetic torques, and gravitational torques (Bloxham, 1998; Buffett, 2007; Jault, 2003; Kuang and Chao, 2003; Ponsar et al., 2003; Roberts and Aurnou, 2012). Viscous coupling is caused by the drag of the core flow on the core–mantle boundary, with the strength of the coupling depending on the viscosity of the core fluid. Given the current estimates of core viscosity, it is generally agreed that viscous torques are too weak to be effective in coupling the core to the mantle (Rochester, 1984).

If the core–mantle boundary is not smooth but exhibits undulations or 'bumps,' then the flow of the core fluid can exert a torque on the mantle due to the fluid pressure acting on the boundary topography (Asari et al., 2006; Buffett, 1998; Hide, 1969, 1977, 1989, 1993, 1995a; Hide et al., 1993; Hinderer et al., 1990; Jault and Le Mouël, 1989, 1990, 1991; Jault et al., 1996; Kuang and Bloxham, 1993; Kuang and Chao, 2001; Mound and Buffett, 2005). The strength of this topographic coupling, a mechanism first suggested by Hide (1969), depends on the amplitude of the topography at the core–mantle boundary. Because of uncertainties in the size of this topography and a controversy about how the topographic torque should be computed (Bloxham and Kuang, 1995; Hide, 1995b, 1998; Jault and Le Mouël, 1999; Kuang and Bloxham, 1997), there is as yet no consensus on the importance of topographic coupling as a mechanism for exchanging angular momentum between the core and mantle.

Electromagnetic torques arise from the interaction between the magnetic field within the core and the flow of electric currents in the weakly conducting mantle that are induced both by time variations of the magnetic field and by diffusion of electric currents from the core to the mantle (Bullard et al., 1950; Dumberry and Mound, 2008; Holme, 1998a,b, 2000; Jault and Le Mouël, 1991; Love and Bloxham, 1994; Mound and Buffett, 2005; Nakada, 2006, 2009a, 2011; Roberts, 1972; Rochester, 1960, 1962; Roden, 1963; Stewart et al., 1995; Stix and Roberts, 1984; Wicht and Jault, 1999, 2000). The strength of this electromagnetic torque, a mechanism first suggested by Bullard et al. (1950), depends both on the conductivity of the mantle and on the strength of the magnetic field crossing the core–mantle boundary. If the conductivity of the mantle, or of a narrow layer at the base of the mantle, is sufficiently large, then electromagnetic torques can produce decadal LOD variations as large as those observed. But because of uncertainties in the conductivity at the base of the mantle, the importance of electromagnetic coupling, like that of topographic coupling, as a mechanism for exchanging angular momentum between the core and mantle remains unclear.

Gravitational attraction between density heterogeneities in the fluid core and mantle can exert a torque on the mantle, leading to changes in the length of day (Buffett, 1996a; Jault and Le Mouël, 1989; Xu et al., 2000). The strength of the gravitational torque depends upon the size of the mass anomalies in the core and mantle, which are poorly known. As a result, there have been few quantitative estimates of the magnitude of the gravitational torque. However, Buffett (1996a,b) suggested that the inner core may be gravitationally locked to the mantle. If so, then any rotational disturbance of the inner core, possibly caused by electromagnetic torques acting on the inner core, will be transmitted to the mantle, causing LOD variations. While Buffett (1998) and Mound and Buffett (2005) considered this last mechanism to be the most viable mechanism for exchanging angular momentum between the core and mantle, Zatman (2003) found that it is inconsistent with a model of inner core rotation rate determined from the core flow within the tangent cylinder. In any case, the amplitude of the observed decadal LOD variations has been used to place limits on the strength of the gravitational coupling between the inner core and the mantle (Aubert and Dumberry, 2011; Buffett and Creager, 1999; Dumberry, 2007; Dumberry and Mound, 2010).

Geomagnetic jerks are sudden changes in the otherwise smoothly changing secular variation of the geomagnetic field. They manifest themselves either as changes in the slope of the secular variation or, equivalently, as steplike changes in the secular acceleration (Mandea et al., 2010). The occurrence of geomagnetic jerks appears to be correlated with sudden changes in the time rate of change of LOD (e.g., Holme and de Viron, 2005; Mandea et al., 2010; Olsen and Mandea, 2007), with changes in the phase of the Chandler wobble (Bellanger et al., 2001, 2002a; Gibert and Le Mouël, 2008; Gibert et al., 1998), and with changes in the phase of the free core nutation (Shirai et al., 2005). While the origin of geomagnetic jerks is currently unknown, these correlations suggest that their cause is related to the same processes in the core that cause the angular momentum of the core, and hence the rotation of the Earth, to change such as torsional oscillations (Bloxham et al., 2002).

3.09.4.1.3 Tidal variations and solid Earth, oceanic, and atmospheric tides

Tidal forces due to the gravitational attraction of the Sun, Moon, and planets deform the solid and fluid parts of the Earth, causing the Earth's inertia tensor to change and hence the Earth's rotation to change. Jeffreys (1928) was the first to predict that the periodic displacement of the solid and fluid masses of the Earth associated with the tides should cause periodic changes in the Earth's rate of rotation at the tidal frequencies. While Markowitz (1955, 1959) reported observing such periodic variations at the fortnightly and monthly tidal frequencies from observations taken at two photographic zenith tubes, these observations were later thrown into doubt by the error analysis of Fliegel and Hawkins (1967). By combining observations from about 55 instruments, Guinot (1970) was able to detect variations in the Earth's rate of rotation at the fortnightly and monthly tidal frequencies that had amplitudes significantly greater than the level of observation noise. Today, the high accuracy of the space-geodetic measurement systems allows long-period tidal effects on UT1 and LOD to be unambiguously observed (e.g., Bellanger et al., 2002b; Benjamin et al., 2006; Chao et al., 1995a; Dickman and Nam, 1995; Gross, 2009a; Hefty and Capitaine, 1990; McCarthy and Luzum, 1993; Nam and Dickman, 1990; Ray and Egbert, 2012; Robertson et al., 1994; Schastok et al., 1994; Schuh and Schmitz-Hübsch, 2000).

In a seminal paper, Yoder et al. (1981) derived a model for the long-period tidal variations in the rotation of the Earth, assuming that the crust and mantle of the Earth are elastic, that the core is decoupled from the mantle, and that the ocean tides are in equilibrium. **Table 5** gives their results for the long-period tidal variations in UT1 and LOD, where the LOD results have been derived here from their UT1 results by using eqn [64]. On intraseasonal timescales, the largest effects are found to be at the fortnightly and monthly tidal periods. The large tidal variations at annual and semiannual periods are obscured in Earth rotation observations by meteorologic effects (see **Section 3.09.4.1.4**), while those having periods of 9.3 and 18.6 years are obscured by the decadal variations in the Earth's rotation (see **Section 3.09.4.1.2**).

The effects of mantle anelasticity on the tidal variations in the Earth's rate of rotation have been discussed by Merriam (1984, 1985), Wahr and Bergen (1986), Defraigne and Smits (1999), Benjamin et al. (2006), and Ray and Egbert (2012). Dissipation associated with mantle anelasticity causes the deformational and hence rotational response of the Earth to lag behind the forcing tidal potential. As a result, not only does mantle anelasticity modify the in-phase rotational response of the Earth to the tidal potential, but also out-of-phase terms are introduced. Anelastic effects are found to modify the elastic rotational response of the Earth by a few percent.

Defraigne and Smits (1999) also considered the effects of nonhydrostatic structure within the Earth on the tidal variations in the Earth's rotation. A nonhydrostatic initial state of the Earth was determined by computing the buoyancy-driven flow in the mantle due to the seismically observed mass anomalies there while accounting for the associated flow-induced boundary deformation, potential readjustment, and mass readjustment in the outer and inner cores. Their results indicate that nonhydrostatic structure within the Earth modifies the rotational response of the Earth to the zonal tide generating potential by less than 0.1%.

Dynamic effects of long-period ocean tides on the Earth's rotation using ocean tide models based upon Laplace's tidal equations have been computed by Brosche et al. (1989), Seiler (1990, 1991), Wünsch and Busshoff (1992), Dickman (1993), Gross (1993), Seiler and Wünsch (1995), Benjamin et al. (2006), Weis (2006), and Dickman and Gross (2010). But the accuracy of ocean tide models greatly improved when TOPEX/Poseidon (T/P) sea surface height measurements became available (Gross, 2009a). Dynamic effects of long-period ocean tides on the Earth's rotation using tide models based upon T/P sea surface height measurements have been computed by Kantha et al. (1998), Desai and Wahr (1999), and Ray and Egbert (2012). **Table 6** gives the results obtained by Kantha et al. (1998) from their ocean tide model that assimilates T/P-derived tides. As expected, dynamic tide effects are seen to be larger at the fortnightly tidal frequency than they are at the monthly frequency, with the amplitude of the

Table 5 Modeled variations in UT1 and LOD caused by elastic solid body and equilibrium ocean tides

Argument					*Period*	*ΔUT1*	*ΔΛ(t)*	*Argument*					*Period*	*ΔUT1*	*ΔΛ(t)*
l	l′	F	D	Ω	*(days)*	*sin (μs)*	*cos (μs)*	l	l′	F	D	Ω	*(days)*	*sin (μs)*	*cos (μs)*
1	0	2	2	2	5.64	−2.35	2.62	1	0	0	0	−1	27.44	53.39	−12.22
2	0	2	0	1	6.85	−4.04	3.71	1	0	0	0	0	27.56	−826.07	188.37
2	0	2	0	2	6.86	−9.87	9.04	1	0	0	0	1	27.67	54.43	−12.36
0	0	2	2	1	7.09	−5.08	4.50	0	0	0	1	0	29.53	4.70	−1.00
0	0	2	2	2	7.10	−12.31	10.90	1	−1	0	0	0	29.80	−5.55	1.17
1	0	2	0	0	9.11	−3.85	2.66	−1	0	0	2	−1	31.66	11.75	−2.33
1	0	2	0	1	9.12	−41.08	28.30	−1	0	0	2	0	31.81	−182.36	36.02
1	0	2	0	2	9.13	−99.26	68.29	−1	0	0	2	1	31.96	13.16	−2.59
3	0	0	0	0	9.18	−1.79	1.22	1	0	−2	2	−1	32.61	1.79	−0.34
−1	0	2	2	1	9.54	−8.18	5.38	−1	−1	0	2	0	34.85	−8.55	1.54
−1	0	2	2	2	9.56	−19.74	12.98	0	2	2	−2	2	91.31	−5.73	0.39
1	0	0	2	0	9.61	−7.61	4.98	0	1	2	−2	1	119.61	3.29	−0.17
2	0	2	−2	2	12.81	2.16	−1.06	0	1	2	−2	2	121.75	−188.47	9.73
0	1	2	0	2	13.17	2.54	−1.21	0	0	2	−2	0	173.31	25.10	−0.91
0	0	2	0	0	13.61	−29.89	13.80	0	0	2	−2	1	177.84	117.03	−4.13
0	0	2	0	1	13.63	−320.82	147.86	0	0	2	−2	2	182.62	−4824.74	166.00
0	0	2	0	2	13.66	−775.69	356.77	0	2	0	0	0	182.63	−19.36	0.67
2	0	0	0	−1	13.75	2.16	−0.99	2	0	0	−2	−1	199.84	4.89	−0.15
2	0	0	0	0	13.78	−33.84	15.43	2	0	0	−2	0	205.89	−54.71	1.67
2	0	0	0	1	13.81	1.79	−0.81	2	0	0	−2	1	212.32	3.67	−0.11
0	−1	2	0	2	14.19	−2.44	1.08	0	−1	2	−2	1	346.60	−4.51	0.08
0	0	0	2	−1	14.73	4.70	−2.00	0	1	0	0	−1	346.64	9.21	−0.17
0	0	0	2	0	14.77	−73.41	31.24	0	−1	2	−2	2	365.22	82.81	−1.42
0	0	0	2	1	14.80	−5.26	2.24	0	1	0	0	0	365.26	−1535.87	26.42
0	−1	0	2	0	15.39	−5.08	2.07	0	1	0	0	1	386.00	−13.82	0.22
1	0	2	−2	1	23.86	4.98	−1.31	1	0	0	−1	0	411.78	3.48	−0.05
1	0	2	−2	2	23.94	10.06	−2.64	2	0	−2	0	0	−1095.17	−13.72	−0.08
1	1	0	0	0	25.62	3.95	−0.97	−2	0	2	0	1	1305.47	42.11	−0.20
−1	0	2	0	0	26.88	4.70	−1.10	−1	1	0	1	0	3232.85	−4.04	0.01
−1	0	2	0	1	26.98	17.67	−4.11	0	0	0	0	2	−3399.18	789.98	1.46
−1	0	2	0	2	27.09	43.52	−10.09	0	0	0	0	1	−6798.38	−161726.81	−149.47

The tabulated coefficients are derived using a constant value of $k/C = 0.94$ as recommended by Yoder et al. (1981). Terms with UT1 amplitudes less than 2 μs are not tabulated. l, l', F, D, and Ω are the Delaunay arguments, expressions for which are given in Simon et al. (1994). The period, given in solar days, is the approximate period of the term as viewed in the terrestrial reference frame.

Source: Yoder CF, Williams JG, and Parke ME (1981) Tidal variations of Earth rotation. *Journal of Geophysical Research* 86: 881–891.

Table 6 Modeled variations in UT1 and LOD caused by long-period dynamic ocean tides

Tide	*Argument*					*Period (days)*	*ΔUT1 mass (μs)*		*ΔUT1 motion (μs)*		*ΔΛ(t) mass (μs)*		*ΔΛ(t) motion (μs)*	
	l	l′	F	D	Ω		*sin*	*cos*	*sin*	*cos*	*cos*	*sin*	*cos*	*sin*
M_f	0	0	2	0	2	13.66	−102.8	33.0	−1.4	15.4	47.28	15.18	0.64	7.08
M_m	1	0	0	0	0	27.56	−119.1	8.8	5.7	10.7	27.15	2.01	−1.30	2.44

The tabulated coefficients are from the assimilated (A) model of Kantha et al. (1998). l, l', F, D, and Ω are the Delaunay arguments, expressions for which are given in Simon et al. (1994). The period, given in solar days, is the approximate period of the term as viewed in the terrestrial reference frame.

Source: Kantha LH, Stewart JS, and Desai SD (1998) Long-period lunar fortnightly and monthly ocean tides. *Journal of Geophysical Research* 103(C6): 12639–12647.

out-of-phase mass and motion terms (the cosine coefficients for UT1 and the sine coefficients for LOD) each being larger for the fortnightly tide than they are for the monthly tide.

Ocean tides in the diurnal and semidiurnal tidal bands also affect the Earth's rate of rotation. While Yoder et al. (1981) were the first to predict these effects using theoretical ocean tide models based upon Laplace's tidal equations, they were not actually observed until Dong and Herring (1990) detected them in VLBI measurements. Subdaily variations in UT1 and LOD at the diurnal and semidiurnal tidal frequencies have now been unambiguously observed in measurements taken by VLBI (Artz et al., 2010, 2011; Böhm et al., 2012; Brosche et al., 1991; Gipson, 1996; Haas and Wünsch, 2006; Herring, 1993; Herring and Dong, 1991, 1994; Sovers et al., 1993; Wünsch and Busshoff, 1992), SLR (Watkins and Eanes, 1994), GPS (Freedman et al., 1994b; Grejner-Brzezinska and Goad, 1996;

Hefty et al., 2000; Lichten et al., 1992; Malla et al., 1993; Rothacher et al., 2001; Steigenberger et al., 2006), and in combinations of GPS and VLBI (Artz et al., 2012; Thaller et al., 2005, 2006, 2007).

Following Yoder et al. (1981), predictions of subdaily tidal variations in the Earth's rate of rotation using theoretical ocean tide models have been made by Brosche (1982), Baader et al. (1983), Brosche et al. (1989), Seiler (1990, 1991), Wünsch and Busshoff (1992), Dickman (1993), Gross (1993), Seiler and Wünsch (1995), and Weis (2006). But as with long-period ocean tide models, the accuracy of diurnal and semidiurnal tide models greatly improved when T/P sea surface height measurements became available. Dynamic effects of diurnal and semidiurnal ocean tides on the Earth's rotation computed from tide models incorporating T/P sea surface height measurements have been given by Ray et al. (1994), Chao et al. (1995b, 1996a), and Chao and Ray (1997). Extensive tables of coefficients for the effect of diurnal and semidiurnal ocean tides on UT1 and LOD are given by Petit and Luzum (2010) (**Chapter 1.08**).

Comparisons of observations with models show the dominant role that ocean tides play in causing subdaily UT1 and LOD variations (**Figure 5(a)**), with as much as 90% of the observed UT1 variance being explained by diurnal and semidiurnal ocean tides (Chao et al., 1996a; Thaller et al., 2006). Apart from errors in observations and models, the small difference that remains (e.g., Schindelegger et al., 2011; Schuh and Schmitz-Hübsch, 2000) may be due to nontidal atmospheric and oceanic effects.

The diurnally varying solar heating of the atmosphere excites diurnal and semidiurnal tidal waves in the atmosphere that travel westward with the Sun (Chapman and Lindzen, 1970; Dai and Wang, 1999; Haurwitz and Cowley, 1973; Volland, 1988, 1997). These radiational tides are much larger than the gravitational tides in the atmosphere, with the amplitude of the surface pressure variations due to the radiational tides being about 20 times larger than the amplitude due to the gravitational tides. While gravitational tides in the atmosphere have no discernible effect on the Earth's rotation, the radiational tides do have an effect (Brzezinski et al., 2002a; de Viron et al., 2005; Zharov and Gambis, 1996). Using the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) reanalysis wind and pressure fields, Brzezinski et al. (2002a) predicted that the UT1 variations caused by the diurnal S_1 radiational tide should have an amplitude of about 0.5 μs and that the LOD variations should have an amplitude of about 3.3 μs. Since the NCEP/NCAR reanalysis wind and pressure fields are given every 6 h, the semidiurnal tidal frequency band is incompletely sampled, making it difficult to estimate the effect on UT1 and LOD of the semidiurnal S_2 radiational tide. In addition, since the oceans will respond dynamically to the tidal variations in the atmospheric wind and pressure fields, the oceans will also contribute to the excitation of UT1 and LOD by the radiational tides. In fact, the effect of radiational tides on UT1 and LOD is typically included in the

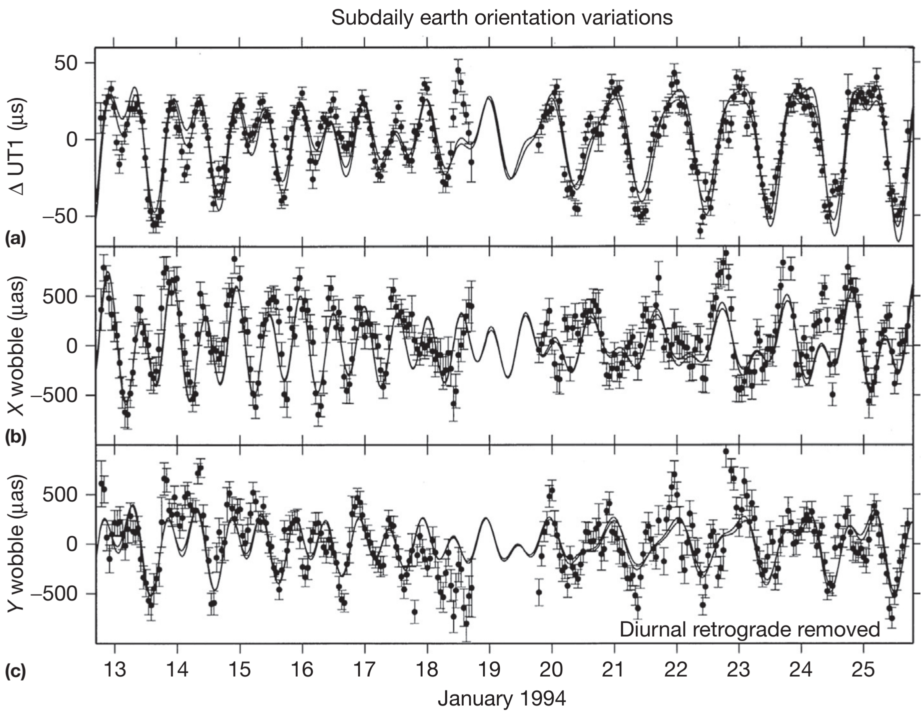

Figure 5 Plots of observed and modeled (a) UT1, (b) x-component of polar motion, and (c) y-component of polar motion during the Cont94 measurement campaign of 12–26 January 1994. The dots with 1σ error bars are the hourly VLBI observations. Polar motion variations in the retrograde nearly diurnal frequency band have been removed from the observed series and are not included in the modeled series. The solid lines are the predicted effects from the diurnal and semidiurnal T/P ocean tide models B and C of Chao et al. (1996a). Each tide model explains about 90% of the observed UT1 variance and about 60% of the observed polar motion variance. Reproduced from Chao BF, Ray RD, Gipson JM, Egbert GD, and Ma C (1996a) Diurnal/semidiurnal polar motion excited by oceanic tidal angular momentum. *Journal of Geophysical Research* 101(B9): 20151–20163.

tables of the effects of diurnal and semidiurnal ocean tides on the Earth's rate of rotation (see, e.g., Petit and Luzum, 2010, Table 8.3a and 8.3b).

3.09.4.1.4 Seasonal variations

Seasonal variations in LOD were first detected by Stoyko (1937). Numerous studies have since shown that the observed annual and semiannual variations in LOD are primarily caused by annual and semiannual changes in the angular momentum of the zonal winds (e.g., Aoyama and Naito, 2000; Eubanks et al., 1985; Eubanks, 1993; Hide and Dickey, 1991; Höpfner, 1998, 2000b; Lambeck, 1980, 1988; Lambeck and Hiopgood, 1981; Munk and MacDonald, 1960; Neef and Matthes, 2012; Rosen, 1993; Yu et al., 1999; Zhou et al., 2008). In fact, within the uncertainty of the earlier LOD measurements, seasonal variations in LOD could be accounted for solely by seasonal variations in the zonal winds (Dickey et al., 1993a; Naito and Kikuchi, 1990, 1991; Rosen, 1993; Rosen and Salstein, 1985, 1991). But as measurement accuracies improved, discrepancies between the observed and modeled variations became noticeable.

With the advent of models of the general circulation of the oceans, it became possible to evaluate the effect of the oceans on seasonal LOD variations (Brosche and Sündermann, 1985; Brosche et al., 1990, 1997; Dickey et al., 1993b; Frische and Sündermann, 1990; Gross, 2004; Gross et al., 2004; Johnson et al., 1999; Kouba and Vondrák, 2005; Marcus et al., 1998; Ponte and Stammer, 2000; Ponte et al., 2001, 2002; Segschneider and Sündermann, 1997; Yan et al., 2006). For example, Table 7 gives the results obtained by Gross et al. (2004). They found that during 1992–2000, the amplitude of the observed annual LOD variation was 369.0 ± 6.4 μs with that caused by the effect of zonal winds integrated to a height of 10 hectoPascals (hPa) being 414.8 ± 5.0 μs, by atmospheric surface pressure variations being 37.4 ± 0.8 μs, by oceanic currents being 7.6 ± 0.3 μs, and by ocean-bottom pressure variations being 7.9 ± 0.3 μs. The sum of the effects of atmospheric winds to 10 hPa, surface pressure, oceanic currents, and bottom pressure had an annual amplitude of 363.0 ± 5.2 μs, the same as that observed to within measurement uncertainty, and a phase difference of only 4.5°. But the effects of the winds above 10 hPa have not yet been taken into account.

Although only 1% of the atmospheric mass is located in the region of the atmosphere above 10 hPa, the strength of the zonal winds there is great enough that they have a noticeable effect on seasonal LOD variations (Dickey et al., 1993a; Höpfner, 2001; Rosen, 1993; Rosen and Salstein, 1985, 1991). Gross et al. (2004) found that during 1992–2000, the annual LOD variation caused by winds above 10 hPa had an amplitude of 20.5 ± 0.3 μs, larger than the sum of the effects of oceanic currents and bottom pressure. Since the annual stratospheric winds are out of phase with the annual tropospheric winds, including the effect of the winds above 10 hPa brings the amplitude of the modeled annual LOD variations down to 343.5 ± 5.2 μs. Thus, when the effects of stratospheric winds are included, the annual LOD budget is not closed.

Table 7 also gives the results obtained by Gross et al. (2004) for the semiannual and terannual (3 cpy) LOD variations. Like the annual LOD budget, the semiannual budget is also not closed. But within measurement uncertainty, LOD variations at the terannual frequency are completely accounted for by the effects of the atmosphere and oceans.

Apart from errors in observations and models, the residual that remains after modeled atmospheric and oceanic effects

Table 7 Observed and modeled nontidal LOD variations at seasonal frequencies during 1992–2000

Excitation process	*Annual*		*Semiannual*		*Terannual*	
	Amplitude (μs)	*Phase (°)*	*Amplitude (μs)*	*Phase (°)*	*Amplitude (μs)*	*Phase (°)*
Observed	369.0 ± 6.4	31.6 ± 1.0	294.3 ± 6.3	−116.5 ± 1.2	52.4 ± 6.4	20.9 ± 7.0
Atmospheric						
Winds (ground to 10 hPa)	414.8 ± 5.0	34.7 ± 0.7	244.3 ± 5.0	−110.0 ± 1.2	54.1 ± 5.0	30.4 ± 5.3
Winds (10–0.3 hPa)	20.5 ± 0.3	−161.0 ± 0.9	29.4 ± 0.3	−122.7 ± 0.6	3.5 ± 0.3	−165.7 ± 5.1
All winds (ground to 0.3 hPa)	395.1 ± 5.0	35.5 ± 0.7	273.1 ± 5.0	−111.3 ± 1.0	50.7 ± 5.0	31.4 ± 5.6
Surface pressure (IB)	37.4 ± 0.8	−154.6 ± 1.3	9.2 ± 0.8	113.3 ± 5.1	2.6 ± 0.8	−48.8 ± 18.1
All winds and surface pressure	358.3 ± 5.1	36.5 ± 0.8	266.7 ± 5.1	−112.7 ± 1.1	51.2 ± 5.1	28.6 ± 5.7
Oceanic						
Currents	7.6 ± 0.3	−165.5 ± 2.3	0.9 ± 0.3	115.5 ± 18.7	1.6 ± 0.3	35.9 ± 11.1
Bottom pressure	7.9 ± 0.3	−148.4 ± 2.2	3.1 ± 0.3	176.7 ± 5.5	1.0 ± 0.3	−67.3 ± 16.4
Currents and bottom pressure	15.3 ± 0.5	−156.8 ± 2.0	3.6 ± 0.5	163.4 ± 8.6	1.7 ± 0.5	−0.6 ± 18.4
Atmospheric and oceanic						
All winds and currents	388.0 ± 5.0	35.9 ± 0.7	272.5 ± 5.0	−111.5 ± 1.1	52.3 ± 5.0	31.6 ± 5.5
Surface and bottom pressure	45.2 ± 0.9	−153.5 ± 1.2	10.9 ± 0.9	127.9 ± 4.9	3.6 ± 0.9	−54.1 ± 15.1
Total of all atmospheric and oceanic						
Without winds above 10 hPa	363.0 ± 5.2	36.1 ± 0.8	238.1 ± 5.2	−112.4 ± 1.3	56.1 ± 5.2	26.9 ± 5.3
With winds above 10 hPa	343.5 ± 5.2	37.1 ± 0.9	267.1 ± 5.2	−113.5 ± 1.1	52.7 ± 5.2	27.7 ± 5.7

IB, inverted barometer. The amplitude A and phase α of the observed and modeled seasonal LOD variations are defined by $\Delta\Lambda(t) = A\cos[\sigma(t - t_o) - \alpha]$ where σ is the annual, semiannual, or terannual frequency and the reference date t_o is 1 January 1990.

Source: Gross RS, Fukumori I, Menemenlis D, and Gegout P (2004) Atmospheric and oceanic excitation of length-of-day variations during 1980–2000. *Journal of Geophysical Research* 109: B01406, http://dx.doi.org/10.1029/2003JB002432.

have been removed from the observations may be caused by hydrologic processes (Chao and O'Conner, 1988; Chen, 2005; Chen et al., 2000a; Yan and Chao, 2012; Zhong et al., 2003). For example, Chen (2005) found that during 1993.0–2004.3, modeled hydrologic processes cause an annual change in LOD of amplitude 17.2 μs, which when added to oceanic effects is reduced to 12.5 μs. However, when the effects of balancing mass within the atmosphere, ocean, and hydrologic system are included, this is further reduced to 1.4 μs, which is not large enough to close the annual LOD budget. His results for the semiannual LOD variations are similar. However, Yan and Chao (2012) found that the annual budget can be nearly closed when global mass balance is enforced and when, in addition, the effect of atmospheric winds is inflated by about 7% (Chao and Yan, 2010).

Since meteorologic processes are the predominant cause of seasonal LOD variations and since these processes can change from year to year, there is no reason to expect that the seasonal LOD variations should be the same from year to year, either in amplitude or in phase. In fact, Feissel and Gavoret (1990) showed that the amplitude of the annual LOD oscillation was about twice as large as normal during the 1982–83 El Niño/Southern Oscillation (ENSO) event, with the amplitude of the semiannual LOD oscillation being about half as large as normal. Gross et al. (1996a, 2002) extended this study, showing that during 1962–2000, the amplitudes of the seasonal LOD and wind-driven AAM variations at both annual and semiannual frequencies have not been constant but have fluctuated by as much as 50%. They also showed that the changing amplitudes of the annual and semiannual LOD and AAM variations are significantly correlated with the Southern Oscillation Index (SOI), an index defined to be the normalized difference in surface pressure between Darwin and Tahiti. Since the SOI is also correlated with LOD and AAM variations occurring on interannual timescales (see **Section 3.09.4.1.5**), the significant correlation they observed between changes in the amplitude of the seasonal cycle and the SOI is evidence of a linkage between the seasonal cycle and interannual LOD and AAM variations, a linkage that can only occur through nonlinear interactions.

3.09.4.1.5 Interannual variations and ENSO

Like seasonal variations in LOD, variations on interannual timescales are also predominantly caused by changes in the angular momentum of the zonal winds (e.g., Eubanks, 1993; Hide and Dickey, 1991; Lambeck and Hopgood, 1981; Neef and Matthes, 2012; Rosen, 1993; Yu et al., 1999). The most prominent feature of the climate system on these timescales is the ENSO phenomenon. ENSO is a global-scale oscillation of the coupled atmosphere–ocean system characterized by fluctuations in atmospheric surface pressure and ocean temperatures in the tropical Pacific (e.g., Philander, 1990). During an ENSO event, in which the SOI decreases, the tropical easterlies collapse causing the AAM to increase. By the conservation of angular momentum, as the AAM increases, the solid Earth's angular momentum decreases and the length of day increases. Numerous studies (e.g., Abarca del Rio et al., 2000; Chao, 1984, 1988, 1989; Dickey et al., 1992a, 1993a, 1994b, 1999; Eubanks et al., 1986; Feissel and Gavoret, 1990; Gambis, 1992; Jordi et al., 1994; Rosen et al., 1984; Salstein and Rosen, 1986; Stefanick, 1982; Zheng et al., 2003; Zhou et al., 2001) have shown that observed LOD variations on interannual timescales, as well as interannual variations in the angular momentum of the zonal winds, are (negatively) correlated with the SOI, reflecting the impact on the length of day of changes in the zonal winds associated with ENSO. For example, Chao (1984) reported finding a correlation between the interannual LOD variations and the SOI during 1957–83 with a maximum negative value for the correlation coefficient of −0.56 obtained when the SOI leads the interannual LOD by 1 month; Eubanks et al. (1986) reported a maximum negative correlation coefficient during 1962–84 of about −0.5 when the SOI leads the interannual LOD by 3 months; Chao (1988) reported a maximum negative value for the correlation coefficient during 1972–86 of −0.68 for a 2-month lead time; and Dickey et al. (1993a) reported a maximum negative correlation coefficient during 1964–89 of −0.67 for a 1-month lead time. Since thermal winds are largely responsible for interannual LOD variations, Dickey et al. (2007), following Su et al. (2005), showed that the 1–2-month lag of interannual LOD variations behind the SOI can be explained by the 1–2-month lag of the tropical atmospheric temperature gradient behind the Niño-3.4 sea surface temperature index (another ENSO index that is highly correlated with the SOI).

In a detailed study of the 1982–83 ENSO event, Dickey et al. (1994b) showed that up to 92% of the observed interannual LOD variance could be explained by atmospheric wind and pressure fluctuations. They suggested that variations in oceanic angular momentum could explain the remaining signal. But studies (Gross et al., 2004; Johnson et al., 1999; Ponte et al., 2002) of the effects of oceanic processes show that they are only marginally effective in causing interannual LOD variations. For example, Gross et al. (2004) found that during 1980–2000, atmospheric winds were the dominant mechanism causing the length of day to change on interannual timescales, explaining 85.8% of the observed variance and having a correlation coefficient of 0.93 with the observations. The effect of atmospheric surface pressure changes explained only 2.6% of the observed variance and was not significantly correlated with the observations. However, including the effect of surface pressure changes with that of the winds increased the observed variance explained from 85.8% to 87.3%. Oceanic currents and bottom pressure changes were found to have only a marginal effect on interannual LOD variations, each explaining less than 1% of the observed variance and neither being significantly correlated with the observations. However, including their effects with those of atmospheric winds and surface pressure changes increased the observed variance explained from 87.3% to 87.9% and increased the correlation coefficient with the observations from 0.93 to 0.94.

The interannual LOD signal that remains after atmospheric and oceanic effects are removed may be caused by hydrologic processes (Chen, 2005; Chen et al., 2000a). For example, Chen (2005) found a significant correlation between the sum of the mass-balanced oceanic and hydrologic angular momenta and the observed interannual LOD variations from which atmospheric effects have been removed, although the modeled variations are of smaller amplitude than those observed. Like seasonal variations, better atmospheric, oceanic, and hydrologic models are needed to close the LOD budget on interannual timescales.

There is also a persistent oscillation in LOD over the last century with a period of about 6 years and an amplitude of about 0.12 ms that is not caused by atmospheric or oceanic fluctuations (Abarca del Rio et al., 2000; Djurovic and Paquet, 1996; Vondrák, 1977). Mound and Buffett (2003, 2006) suggested that it is the signature of a free mode of oscillation caused by the exchange of angular momentum between the mantle and core arising from gravitational coupling between the mantle and inner core. On the other hand, Gillet et al. (2010) suggested that it is the result of fast torsional waves in the core similar to the torsional oscillations that have been discussed as being the cause of decadal LOD variations (**Section 3.09.4.1.2**) but that now occur at much higher frequencies.

3.09.4.1.6 Intraseasonal variations and the Madden–Julian oscillation

Like the seasonal and interannual variations in the length of day, variations on intraseasonal timescales are also predominantly caused by changes in the angular momentum of the zonal winds (e.g., Dickey et al., 1992b, 2010; Eubanks, 1993; Hide and Dickey, 1991; Lambeck and Hopgood, 1981; Neef and Matthes, 2012; Rosen, 1993; Rosen et al., 1990; Yu et al., 1999; Zhou et al., 2008). The Madden–Julian oscillation (Madden and Julian, 1971, 1972, 1994, 2005) with a period of 30–60 days is the most prominent feature in the atmosphere on these timescales and a number of studies have shown that fluctuations in the zonal winds associated with this oscillation cause the length of day to change (Anderson and Rosen, 1983; Chao and Salstein, 2005; Dickey et al., 1991; Feissel and Gambis, 1980; Feissel and Nitschelm, 1985; Hendon, 1995; Langley et al., 1981b; Madden, 1987; Marcus et al., 2001).

Studies of the effects of oceanic processes show that they are only marginally effective in causing intraseasonal LOD variations (Gross et al., 2004; Johnson et al., 1999; Kouba and Vondrák, 2005; Ponte, 1997; Ponte and Ali, 2002; Ponte and Stammer, 2000). For example, Gross et al. (2004) found that during 1992–2000, atmospheric winds were the dominant mechanism causing the length of day to change on intraseasonal timescales, explaining 85.9% of the observed intraseasonal variance and having a correlation coefficient of 0.93 with the observations. Atmospheric surface pressure, oceanic currents, and ocean-bottom pressure were found to have only a minor effect on intraseasonal LOD changes, each explaining only about 3–4% of the observed variance. However, including the effect of surface pressure changes with that of the winds increased the variance explained from 85.9% to 90.2% and increased the correlation coefficient with the observations from 0.93 to 0.95. Additionally including the effects of changes in oceanic currents and bottom pressure further increased the variance explained from 90.2% to 92.2% and further increased the correlation coefficient from 0.95 to 0.96. Thus, although the impact of the oceans is relatively minor, closer agreement with the observations in the intraseasonal frequency band is obtained when the effects of oceanic processes are added to that of atmospheric.

While the overall contribution of the oceans to exciting intraseasonal LOD variations may be relatively minor, the effect on LOD of specific rapid changes in the oceans' circulation is detectable. For example, after accounting for the effect of the atmosphere, Marcus et al. (2012) showed that rapid changes in the LOD residual that occurred during 8–21 November 2009 were caused by rapid changes in the Antarctic Circumpolar Current as diagnosed from a data assimilating ocean model.

Hydrologic effects on intraseasonal LOD variations are thought to be relatively insignificant (Chen, 2005; Chen et al., 2000a).

3.09.4.2 Polar Motion

Observations of polar motion (**Figure 6**) show that it consists mainly of (1) a forced annual wobble having a nearly constant amplitude of about 100 mas, (2) the free Chandler wobble having a period of about 433 days and a variable amplitude ranging from about 100 to 200 mas, (3) quasiperiodic variations on decadal timescales having amplitudes of about 30 mas known as the Markowitz wobble, (4) a linear trend having a rate of about 3.5 mas year^{-1} and a direction toward 79°W longitude, and (5) smaller amplitude variations occurring on all measurable timescales. Here, the polar motion variations that are observed and the models that have been developed to explain them are reviewed.

The motion of the pole is usually described by giving the *x*- and *y*-components of its location in the terrestrial reference frame. But since polar motion is inherently a two-dimensional quantity, periodic motion of the pole can also be described by giving the amplitude A and phase α of its prograde and retrograde components defined by

$$\mathbf{p}(t) = p_x(t) - ip_y(t) = A_p e^{i\alpha_p} e^{i\sigma(t-t_o)} + A_r e^{i\alpha_r} e^{-i\sigma(t-t_o)} \qquad [68]$$

where the subscript p denotes prograde, the subscript r denotes retrograde, σ is the strictly positive frequency of motion, and t_o is the reference date. Prograde motion of the pole is circular

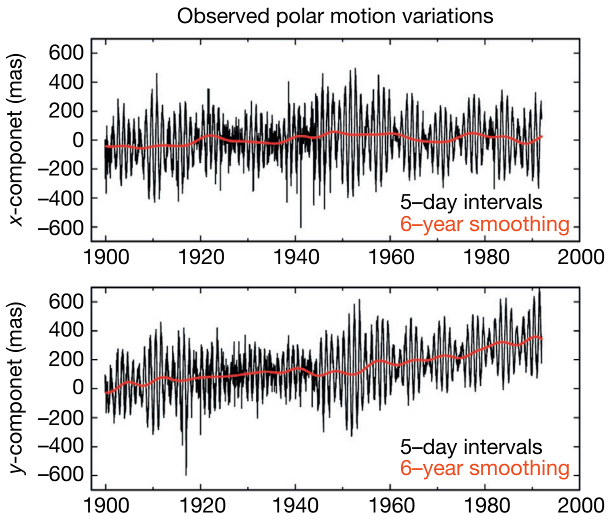

Figure 6 Observed polar motion variations from the Hipparcos optical astrometric series. The low-frequency variations shown in red were obtained by applying to the 5-day Hipparcos series a low-pass boxcar filter having a 6-year cutoff period. The beating between the 12-month annual wobble and the 14-month Chandler wobble is readily apparent.

motion in a counterclockwise direction; retrograde motion is circular motion in a clockwise direction. In general, the sum of prograde and retrograde circular motion is elliptical motion, with linear motion resulting when the amplitudes of the prograde and retrograde components are the same.

By eqns [41], [42], and [57], the observed polar motion variations are excited by changes in the equatorial components $h_x(t)$ and $h_y(t)$ of relative angular momentum and by changes in the equatorial components $\Omega\Delta I_{xz}(t)$ and $\Omega\Delta I_{yz}(t)$ of angular momentum due to changes in the mass distribution of the Earth. Like modeling the observed LOD changes, modeling the observed polar motion excitation requires computing both types of changes in angular momentum for the different components of the Earth system. From eqn [65] and again assuming for this purpose that the terrestrial reference frame is uniformly rotating at the rate Ω about the z-coordinate axis, the equatorial components of the mass term of the angular momentum can be written as

$$\Omega\Delta I_{xz}(t) = -\Omega\int_V \rho(\mathbf{r}, t) r^2 \sin\phi\cos\phi\cos\lambda\, \mathrm{d}V \qquad [69]$$

$$\Omega\Delta I_{yz}(t) = -\Omega\int_V \rho(\mathbf{r}, t) r^2 \sin\phi\cos\phi\sin\lambda\, \mathrm{d}V \qquad [70]$$

where ϕ is north latitude and λ is east longitude. Similarly, the equatorial components of the motion term of the angular momentum can be written as

$$h_x(t) = \int_V \rho(\mathbf{r}, t)[r\sin\lambda v(\mathbf{r}, t) - r\sin\phi\cos\lambda u(\mathbf{r}, t)]\,\mathrm{d}V \qquad [71]$$

$$h_y(t) = \int_V \rho(\mathbf{r}, t)[-r\cos\lambda v(\mathbf{r}, t) - r\sin\phi\sin\lambda u(\mathbf{r}, t)]\,\mathrm{d}V \qquad [72]$$

where $u(\mathbf{r},t)$ is the eastward component of the velocity and $v(\mathbf{r},t)$ is the northward component.

3.09.4.2.1 True polar wander and glacial isostatic adjustment

Determining an unbiased estimate of the linear trend in the observed path of the pole over the past 100 years is complicated by the presence of the annual, Chandler, and Markowitz wobbles. Various approaches have been taken to estimate the trend in the presence of these large-amplitude periodic and quasiperiodic variations (see **Table 8** for the resulting trend estimates). The annual wobble has been removed both by a least-squares fit (McCarthy and Luzum, 1996; Schuh et al., 2000, 2001; Wilson and Gabay, 1981) and by a seasonal adjustment of the polar motion series (Wilson and Vicente, 1980). The Chandler wobble has been removed both by a least-squares fit for periodic terms (Dickman, 1981; McCarthy and Luzum, 1996; Schuh et al., 2000, 2001) and by deconvolution (Vicente and Wilson, 2002; Wilson and Gabay, 1981; Wilson and Vicente, 1980). Smoothing has also been used to remove the annual and Chandler wobbles (Höpfner, 2004; Okamoto and Kikuchi, 1983). The decadal-scale variations have been removed by modeling them as being strictly periodic at a single frequency of 1/31 cpy and then least-squares fitting a sinusoid at this single frequency (Dickman, 1981; McCarthy and Luzum, 1996).

Rather than modeling the decadal variations as being strictly periodic at a single frequency, Gross and Vondrák (1999) devised a method to account for their quasiperiodic nature when estimating the linear trend. The annual and Chandler wobbles were first removed from the observations by applying to the polar motion series a low-pass boxcar filter having a 6-year cutoff period. A spectrum of the resulting low-pass filtered polar motion series was then computed and the linear trend was estimated by a simultaneous weighted least-squares fit for a mean, a trend, and periodic terms at the frequencies of all the peaks evident in the spectrum. The estimates for the linear trend that they obtained by applying this technique to the ILS, Hipparcos, and SPACE96 polar motion series are given in **Table 8**. **Figure 7** shows the observed low-pass filtered polar motion series they used (solid red lines); the modeled series they obtained by the simultaneous weighted

Table 8 Observed linear trend in the path of the pole

Rate (mas year⁻¹)	*Direction (°W)*	*Data span*	*Source*
Hipparcos polar motion			
3.39	78.5	1899.7–1992.0	(a)
3.51 ± 0.01	**79.2 ± 0.20**	**1900.0–1992.0**	**(b)**
3.31 ± 0.05	76.1 ± 0.80	1899.7–1992.0	(c)
Hipparcos polar motion excitation			
2.84	73.03	1899–1992	(d)
ILS polar motion			
3.4	78	1900–1977	(e)
3.521 ± 0.094	80.1 ± 1.6	1899.8–1979.0	(f)
3.52	79.4	1899.8–1979.0	(g)
3.456	80.56	1899.0–1979.0	(h)
3.81 ± 0.07	75.5 ± 1.0	1899.8–1979.0	(b)
ILS polar motion excitation			
3.4	66	1900–1977	(e)
3.3	65	1901–1970	(i)
3.49	79.5	1899–1979	(d)
Latitude observations			
3.62	89	1900–1978	(j*)
3.51	79	1900–1978	(j†)
3.24	84.9	1899.8–1979.0	(k†)
2.97	77.7	1899.8–1979.0	(k‡)
Space-geodetic polar motion			
3.39 ± 0.53	85.4 ± 4.0	1976–1994	(l)
4.123 ± 0.002	73.9 ± 0.03	1976.7–1997.1	(b)
4.5 ± 0.1	68 ± 8	1976–1992	(m)
1.8 ± 0.4	58 ± 9	1992–2008	(m)
Combined astrometric and space-geodetic polar motion			
3.29	78.2	1900.0–1984.0	(n)
3.33 ± 0.08	75.0 ± 1.1	1899.8–1994.1	(l)
3.901 ± 0.022	65.17 ± 0.22	1891.0–1999.0	(o)
Combined astrometric and space-geodetic excitation			
3.35	76.3	1900–1999	(d)
3.54	69.92	1900–1999	(d)

The recommended estimate is given in bold.

Sources: (a) Vondrák et al. (1998), (b) Gross and Vondrák (1999), (c) Schuh et al. (2000, 2001), (d) Vicente and Wilson (2002), (e) Wilson and Vicente (1980), (f) Dickman (1981), (g) Chao (1983), (h) Okamoto and Kikuchi (1983), (i) Wilson and Gabay (1981), (j*) Zhao and Dong (1988) based on measurements taken at nine latitude observing stations, (j†) Zhao and Dong (1988) based on measurements taken at the five ILS observing stations located at Mizusawa, Kitab, Carloforte, Gaithersburg, and Ukiah, (k†) Vondrák (1994) based on measurements taken at the five ILS latitude observing stations located at Mizusawa, Kitab, Carloforte, Gaithersburg, and Ukiah, (k‡) Vondrák (1994) based on measurements taken at the four ILS latitude observing stations located at Mizusawa, Kitab, Carloforte, and Gaithersburg, (l) McCarthy and Luzum (1996), (m) Roy and Peltier (2011), (n) Vondrák (1985), (o) Höpfner (2004).

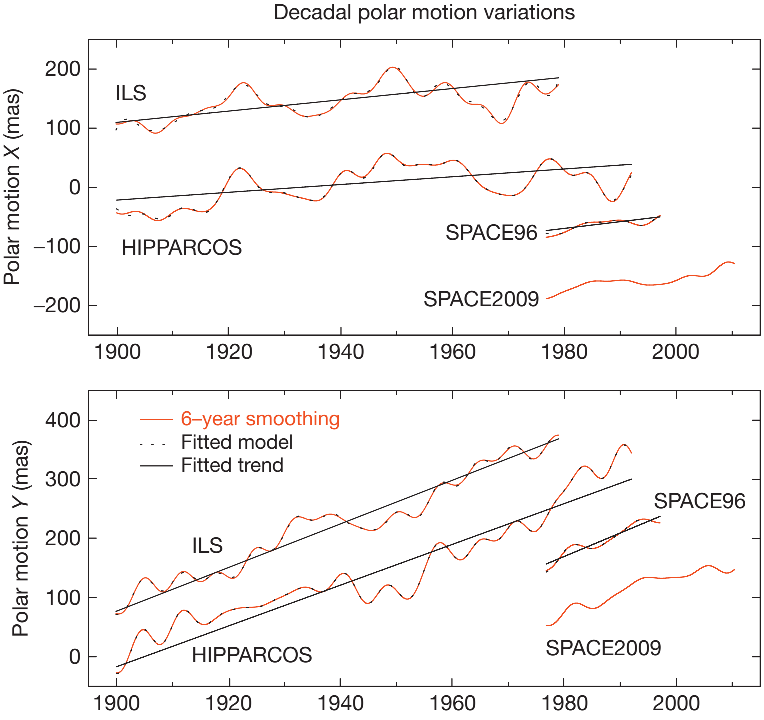

Figure 7 Observed decadal-scale polar motion variations (solid red lines) from the ILS optical astrometric series, the Hipparcos optical astrometric series, and the SPACE96 and SPACE2009 combined space-geodetic series. The decadal-scale variations were obtained by applying to the original series a low-pass boxcar filter having a 6-year cutoff period. The solid black lines show the linear trends in the pole path that were estimated from the ILS, Hipparcos, and SPACE96 series by Gross and Vondrák (1999). The dotted black lines show the model of the decadal variations that they used when estimating the trend. For clarity of display, the curves have been offset from each other by an arbitrary amount. Adapted from Gross RS and Vondrák J (1999) Astrometric and space-geodetic observations of polar wander. *Geophysical Research Letters* 26(14): 2085–2088.

least-squares fit for a mean, a trend, and all periodic terms evident in the spectrum (dotted black lines); and the resulting linear trend estimates (solid black lines). As can be seen, their model is an excellent fit to the observations, and hence, their estimates for the linear trend should be unbiased by the presence of the quasiperiodic decadal-scale polar motion variations. For this reason and because the Hipparcos series spans a greater length of time than the other series that they studied, the recommended estimate for the linear trend in the pole path is the estimate that they obtained for the Hipparcos series, namely, a trend of rate 3.51 ± 0.01 mas year^{-1} toward $79.2 \pm 0.20°$W longitude.

The estimates given in **Table 8** for the observed linear trend in the path of the pole are with respect to a terrestrial reference frame that has been attached to the mean lithosphere in such a manner that within it, the tectonic plates exhibit no net rotation. Argus and Gross (2004) argued that ideally, the motion of the pole should be given with respect to a reference frame that is attached to the mean solid Earth. They argued that it would therefore be better to use a reference frame that is attached to hotspots rather than the mean lithosphere because hotspots move slower than the mean lithosphere with respect to the mean solid Earth and hence better represent the mean solid Earth. Transforming the Hipparcos trend results of Gross and Vondrák (1999) to a hotspot reference frame, they found that during the past century, the linear trend in the path of the pole relative to hotspots was 4.03 mas year^{-1} toward 68.4° W longitude or about 15% faster and in a more eastward direction than the trend relative to the mean lithosphere.

One of the most important mechanisms acting to cause a linear trend in the path of the pole on timescales of a few thousand years is GIA. The isostatic adjustment of the solid Earth as it responds to the decreasing load on it following the last deglaciation causes the figure of the Earth to change and hence the pole to drift. Models of GIA show that its effect on the pole path is sensitive to the assumed value of lower mantle viscosity, to the assumed thickness and rheology of the lithosphere, to the treatment of the density discontinuity at 670 km depth, and to the assumed compressibility of the Earth model (e.g., Cambiotti et al., 2010; Johnston and Lambeck, 1999; Matsuyama et al., 2010; Mitrovica and Milne, 1998; Mitrovica and Wahr, 2011; Mitrovica et al., 2005, 2006, 2007; Nakada, 2000, 2002, 2009b; Peltier, 2007; Peltier and Drummond, 2010; Peltier and Jiang, 1996; Peltier and Luthcke, 2009; Peltier and Wu, 1983; Sabadini and Peltier, 1981; Sabadini and Vermeersen, 2002, 2004; Sabadini et al., 1982, 2002; Tamisea et al., 2002; Tsai and Stevenson, 2007; Vermeersen and Sabadini, 1996, 1999; Vermeersen et al., 1996, 1997, 1998; Wu and Peltier, 1984; Yuen et al., 1982).

However, GIA is not the only mechanism that will cause a trend in the pole path. The present-day change in glacier and ice sheet mass and the accompanying change in nonsteric sea level will also cause a linear trend in polar motion (Gasperini et al., 1986; James and Ivins, 1995, 1997; Mitrovica et al., 2006;

Nakada and Okuno, 2003; Peltier, 1988; Trupin, 1993; Trupin et al., 1992). But the effect of this mechanism is very sensitive to the still unknown present-day mass change of glaciers and ice sheets, particularly of the Antarctic ice sheet. By adopting various scenarios for the mass change of Antarctica, models predict that its mass change alone should cause a linear trend in the pole path ranging anywhere in rate from 0.31 to 4.46 mas year^{-1} and in direction from 101°W to 281°W longitude (James and Ivins, 1997).

Other mechanisms that may cause a linear trend in the path of the pole over the last or next century include continental drift (Dickman, 1979; Nakada, 2007, 2008), tectonic processes taking place under nonisostatic conditions (Sabadini and Vermeersen, 2004; Vermeersen and Vlaar, 1993; Vermeersen et al., 1994), plate subduction (Alfonsi and Spada, 1998; Greff-Lefftz, 2011; Ricard and Sabadini, 1990; Ricard et al., 1992, 1993; Richards et al., 1997; Rouby et al., 2010; Spada et al., 1992; Steinberger and Torsvik, 2010), mantle convection (Cambiotti et al., 2011; Chan et al., 2011a, 2011b; Greff-Lefftz, 2011; Nakada, 2008, 2009b; Richards et al., 1999; Schaber et al., 2009; Steinberger and O'Connell, 1997), upwelling mantle plumes (Greff-Lefftz, 2004, 2011; Rouby et al., 2010; Steinberger and O'Connell, 2002), earthquakes (Alfonsi and Spada, 1998; Chao and Gross, 1987; Chao et al., 1996b; Gross and Chao, 2006; Soldati and Spada, 1999; Spada, 1997; Zhou et al., 2013), and climate change (Landerer et al., 2009; Ponte et al., 2002).

As noted by Roy and Peltier (2011), the rate and direction of the linear trend in the pole path appear to have changed starting about 1992 (see **Table 8**). They speculated that this change may be caused by present-day changes in glacier and ice sheet mass, although changes in continental water storage may also be important. However, given the known presence of decadal-scale variations in the pole path, it is uncertain whether the pole will continue along this new path in the future or whether it will revert back to its pre-1992 long-term average rate and direction.

3.09.4.2.2 Decadal variations, the Markowitz wobble, and core–mantle interactions

By analyzing ILS measurements taken during 1900–59, Markowitz (1960, 1961) found that at long periods, the motion of the pole with respect to the Earth's crust and mantle includes a periodic component superimposed on a linear drift. He found that the periodic component, now known as the Markowitz wobble in his honor, had a period of 24 years and an amplitude of 22 mas and that the linear drift had a rate of 3.2 mas year^{-1} and a direction toward 60° W longitude, remarkably close to the recent determination of 3.51 mas year^{-1} toward 79.2° W longitude (Gross and Vondrák, 1999) that is recommended here.

In more recent rereductions of the ILS and other optical astrometric measurements, which extend more than three decades past those analyzed by Markowitz (1960, 1961), rather than appearing as a strictly periodic phenomenon of well-defined frequency, the Markowitz wobble now appears as a quasiperiodic variation on decadal timescales having an amplitude of about 30 mas (see **Figure 7**). However, since optical astrometric measurements are known to be corrupted by systematic errors, there has always been some doubt about the reality of the decadal variations seen in the ILS and Hipparcos series. Since the highly accurate space-geodetic measurements are less susceptible to systematic error than are optical astrometric measurements, any decadal variations seen in the space-geodetic measurements can be considered to be reliable. **Figure 7** compares the decadal variations seen in the SPACE96 and SPACE2009 combined space-geodetic series with those seen in the ILS and Hipparcos optical astrometric series. While decadal variations are seen to be present in the SPACE96 and SPACE2009 series and hence can be considered to be real, they have a smaller amplitude and a different phase than those seen in either the ILS or Hipparcos series. Thus, space-geodetic measurements both confirm the reality of decadal polar motion variations and demonstrate that the decadal variations seen in the ILS and Hipparcos series are unreliable and should be used only to place an upper bound on the size of the decadal variations.

The cause of the decadal-scale polar motion variations is currently unknown. Gross et al. (2005) found that redistribution of mass within the atmosphere and oceans cannot be the main excitation source of decadal polar motion variations during 1949–2002 since it amounts to only 20% (x-component) and 38% (y-component) of that observed and with the modeled excitation being 180° out of phase with that observed. However, the ocean model used in their study was not forced by mass changes associated with precipitation, evaporation, or runoff from rivers including that from glaciers and ice sheets and so had a constant total mass. Thus, their study did not address the question of the excitation of decadal polar motion by processes that change the total mass of the oceans, such as a nonsteric sea level height change associated with glacier and ice sheet mass change.

Wilson (1993) noted that an oscillation in global sea level on decadal timescales would excite decadal polar motion variations with a polarization similar to that observed, implying that mass change of the oceans is responsible for exciting the observed decadal polar motions. In fact, using a climate model, Celaya et al. (1999) showed that changes in Antarctic snow pack are capable of inducing decadal polar motion variations of nearly the same amplitude as that observed. But realistic estimates of mass change in glaciers and the Antarctic and other ice sheets, along with estimates of the accompanying nonsteric change in sea level, are required to further evaluate this possible source of decadal polar motion variations.

Changes in continental water storage may also be an important excitation source of decadal polar motion variations. But as with changes in present-day glacier and ice sheet mass, realistic estimates of changes in continental water storage are needed before this possible source of decadal polar motion variations can be accurately evaluated.

Since core–mantle processes are known to cause decadal variations in the length of day, they may also excite decadal variations in polar motion. But electromagnetic coupling between the core and mantle appears to be two to three orders of magnitude too weak (Greff-Lefftz and Legros, 1995) and topographic coupling appears to be too weak by a factor of three to ten (Asari et al., 2006; Dumberry, 2010; Greff-Lefftz and Legros, 1995; Hide et al., 1996; Hulot et al., 1996). In addition, the modeled decadal polar motion variations resulting from these studies show little agreement in phase with the observed variations.

Following the study of Jochmann (1989), Greiner-Mai et al. (2000, 2003) and Greiner-Mai and Barthelmes (2001) suggested that irregular motion of a tilted oblate inner core may excite decadal polar motion variations, although they did not account for the dynamic effects of the fluid core in their model. Dumberry and Bloxham (2002), who did account for the dynamic effects of the fluid core, suggested that a tilt of the inner core with respect to the mantle of only 0.07°, perhaps caused by an electromagnetic torque acting on the inner core, would generate gravitational and pressure torques on the mantle strong enough to excite decadal polar motion variations having amplitudes as large as those observed and, like the observed variations, having a preferred direction of motion. But when Mound (2005) examined this mechanism, he concluded that while torsional oscillations may excite both decadal LOD and polar motion variations, when the oscillations are constrained to match the observed LOD amplitude, the resulting electromagnetic torque on the inner core is too weak to excite decadal polar motions to their observed level.

Dumberry (2008) further studied the influence of the inner core on polar motion by examining the gravitational coupling between the inner core and density heterogeneities in the mantle. He found that his model is capable of producing decadal polar motions having amplitudes as large as those observed and that are polarized about a longitudinal axis. But when applying constraints to his model based on observed LOD variations, the amplitudes are found to be reduced to less than 1 mas, much less than those observed. While tantalizingly close results are obtained when the inner core is included in models of the effect of core–mantle processes on polar motion, such models have not yet explained all of the observed properties of the Markowitz wobble.

In closing, it is worth noting that the interaction between the core flow and a realistic nonaxial, nondipolar magnetic field will produce both axial and equatorial torques on the inner core and mantle, leading to LOD and polar motion variations that are coupled (Dumberry and Bloxham, 2002; Mound, 2005; Nakada, 2006). While such coupled motions have not yet been observed, it is worth searching the observations for them since it would provide additional insight into core–mantle processes.

3.09.4.2.3 Tidal wobbles and oceanic and atmospheric tides

Tidally induced deformations of the solid Earth caused by the second-degree zonal tide raising potential cause long-period changes in the Earth's rate of rotation (see **Section 3.09.4.1.3**). But since this potential is symmetrical about the polar axis, tidal deformations of the axisymmetric solid Earth cannot excite polar motion. However, due to the nonaxisymmetric shape of the coastlines, the second-degree zonal tide raising potential acting on the oceans can generate polar motion via the exchange of nonaxial oceanic tidal angular momentum with the solid Earth. Yoder et al. (1981) discussed the possible existence of long-period tidal variations in polar motion and tables of such variations predicted from theoretical ocean tide models have been given by Seiler (1990, 1991), Dickman (1993), Gross (1993), Brosche and Wünsch (1994), Seiler and Wünsch (1995), Weis (2006), and Dickman and Gross (2010).

Table 9 (also see Petit and Luzum, 2010, Table 8.4) gives the results obtained by Dickman and Gross (2010) from the theoretical ocean tide model of Dickman (1993) as revised by Dickman and Nam (1995). These results were shown by Dickman and Gross (2010) to fit the observed long-period tidal variations in polar motion better than any of the other models that they studied.

The results of Dickman and Gross (2010) for the effect of long-period ocean tides on polar motion were obtained from a theoretical ocean tide model that was not constrained by any type of data. As with UT1 and LOD, more accurate results for the effect of long-period ocean tides on polar motion can be expected to be obtained from tide models that are constrained

Table 9 Modeled variations in polar motion and polar motion excitation caused by long-period ocean tides

Tide	Argument					Period (days)	Polar motion				Polar motion excitation			
							Prograde		Retrograde		Prograde		Retrograde	
	l	l′	F	D	Ω		Amp (μas)	Phase (°)	Amp (μas)	Phase (°)	Amp (μas)	Phase (°)	Amp (μas)	Phase (°)
$M_{t'}$	1	0	2	0	1	9.12	4.43	−112.62	5.57	21.33	205.83	67.21	269.95	21.17
M_t	1	0	2	0	2	9.13	10.72	−112.56	13.48	21.30	497.59	67.27	652.59	21.14
$M_{f'}$	0	0	2	0	1	13.63	27.35	−91.42	30.59	13.31	841.32	88.42	1002.12	13.15
M_f	0	0	2	0	2	13.66	66.09	−91.31	73.86	13.27	2028.73	88.53	2414.94	13.11
M_{Sf}	0	0	0	2	0	14.77	5.94	−87.13	6.42	11.75	168.13	92.70	194.74	11.60
M_m	1	0	0	0	0	27.56	43.74	−56.70	31.12	−0.91	643.61	123.13	520.16	−1.06
M_{Sm}	−1	0	0	2	0	31.81	8.85	−51.11	5.42	−4.21	111.62	128.72	79.23	−4.36
S_{sa}	0	0	2	−2	2	182.62	86.48	−20.30	99.77	175.57	118.56	159.42	336.32	175.46
S_a	0	1	0	0	0	365.26	17.96	−17.38	152.15	170.60	3.33	161.60	332.53	170.51
M_n	0	0	0	0	1	−6798.38	208.17	166.89	186.98	166.67	221.43	166.88	175.07	166.68

l, l', F, D, and Ω are the Delaunay arguments, expressions for which are given in Simon et al. (1994). The period, given in solar days, is the approximate period of the term as viewed in the terrestrial reference frame. The amplitude (amp) and phase of the prograde and retrograde components of polar motion are defined by eqn [68]. The amplitude and phase of the prograde and retrograde components of polar motion excitation are similarly defined but with $\boldsymbol{\chi}(t) = \chi_x(t) + i\chi_y(t)$.

Source: Dickman SR and Gross RS (2010) Rotational evaluation of a long-period spherical harmonic ocean tide model. *Journal of Geodesy* 84: 457–464, http://dx.doi.org/10.1007/s00190-010-0383-5.

by altimetric sea surface height data. Unfortunately, very few such data-constrained model results are available, and those that are available are either erroneous or incomplete. The results of Desai and Wahr (1995), as discussed by Gross et al. (1997), did not converge and are thus suspect even though data through T/P cycle 130 were used. And Ray and Egbert (2012) gave results for only the fortnightly ocean tide. More complete results for the effect of long-period ocean tides on polar motion based on altimetric-constrained ocean tide models are desirable.

Long-period tidal variations in polar motion were first observed by Chao (1994) and have been discussed by Gross et al. (1996b, 1997), Gross (2009b), and Ray and Egbert (2012). **Table 10** gives the results of Gross (2009b) for the observed variations that were determined by him by fitting periodic terms at the tabulated tidal frequencies to polar motion excitation observations spanning from 2 January 1980 to 8 September 2006 from which atmospheric and nontidal oceanic effects had been removed. While these results fully explain the observed long-period tidal variations during this time interval, they will most likely not be able to fully explain the observed variations outside this time interval.

Ocean tides in the diurnal and semidiurnal tidal bands also cause polar motion variations. While Yoder et al. (1981) discussed the possible existence of such polar motion variations, they were not actually observed until Dong and Herring (1990) detected them in VLBI measurements. Subdaily variations in polar motion at the diurnal and semidiurnal tidal frequencies have now been observed in measurements taken by VLBI (Artz et al., 2010, 2011; Böhm et al., 2012; Gipson, 1996; Haas and Wünsch, 2006; Herring, 1993; Herring and Dong, 1994; Sovers et al., 1993), SLR (Watkins and Eanes, 1994), GPS (Hefty et al., 2000; Rothacher et al., 2001; Steigenberger et al., 2006), and in combinations of GPS and VLBI measurements (Artz et al., 2012; Thaller et al., 2005, 2006, 2007). Predictions of subdaily tidal variations in polar motion using theoretical ocean tide models have been given by Seiler (1990, 1991), Dickman (1993), Gross (1993), Brosche and Wünsch (1994), Seiler and Wünsch (1995), and Weis (2006). Predictions of subdaily tidal variations in polar motion using tide models based upon T/P sea surface height measurements have been given by Chao et al. (1996a) and Chao and Ray (1997). Extensive tables of coefficients for the effect of diurnal and semidiurnal ocean tides on polar motion are given by Petit and Luzum (2010) (**Chapter 1.08**).

Comparisons of observations with models show the major role that ocean tides play in causing subdaily polar motion variations (**Figure 5(b)** and **5(c)**), with as much as 60% of the observed polar motion variance being explained by diurnal and semidiurnal ocean tides (Chao et al., 1996a; Thaller et al., 2006). Apart from errors in observations and models, the difference that remains (e.g., Schuh and Schmitz-Hübsch, 2000) may be due to nontidal atmospheric (Weber et al., 2002) and oceanic (Nastula et al., 2007a) effects, although Schindelegger et al. (2011) found that nontidal atmospheric effects can explain only about 25% of the residual polar motion signal.

The effects on polar motion of the radiational tides in the atmosphere have been studied by Zharov and Gambis (1996), Brzezinski et al. (2002a), and de Viron et al. (2005). However, since the oceans will respond dynamically to subdaily tidal variations in the atmospheric wind and pressure fields, the oceans will also contribute to the excitation of polar motion by the radiational tides. Brzezinski et al. (2004) predicted the effects on polar motion of the subdaily radiational tides in the atmosphere and oceans using a barotropic ocean model forced by the 6-h NCEP/NCAR reanalysis wind and pressure fields (also see Gross, 2005a). **Table 11** gives their results at prograde nearly diurnal frequencies. They found the largest effect to be at the prograde S_1 tidal frequency with an amplitude of 9 μas that is primarily excited by ocean-bottom pressure variations. Their predictions at retrograde nearly diurnal frequencies affect the nutations and are therefore not reproduced here. Since the NCEP/NCAR reanalysis wind and pressure fields are given every 6 h, the semidiurnal tidal frequency band is incompletely sampled, making it difficult to estimate atmospheric and oceanic effects on polar motion at these frequencies. More complete models for the effects of the radiational tides at nearly semidiurnal frequencies must await the availability of atmospheric fields sampled more often than every 6 h.

Table 10 Observed variations in polar motion and polar motion excitation caused by long-period ocean tides

Tide	Argument					Period (days)	Polar motion				Polar motion excitation			
							Prograde		Retrograde		Prograde		Retrograde	
	l	l′	F	D	Ω		Amp (μas)	Phase (°)	Amp (μas)	Phase (°)	Amp (μas)	Phase (°)	Amp (μas)	Phase (°)
$M_{t'}$	1	0	2	0	1	9.12	3.88	−128.58	1.65	−26.83	180.15	51.26	80.18	−26.98
M_t	1	0	2	0	2	9.13	9.36	−128.58	4.00	−26.83	434.55	51.26	193.41	−26.98
$M_{f'}$	0	0	2	0	1	13.63	28.62	−105.95	36.81	39.42	880.22	73.88	1205.93	39.27
M_f	0	0	2	0	2	13.66	69.18	−105.95	88.97	39.42	2123.44	73.88	2909.16	39.27
M_m	1	0	0	0	0	27.56	38.56	−111.41	83.24	34.77	567.37	68.42	1391.36	34.62

l, l', F, D, and Ω are the Delaunay arguments, expressions for which are given in Simon et al. (1994). The period, given in solar days, is the approximate period of the term as viewed in the terrestrial reference frame. The amplitude (amp) and phase of the prograde and retrograde components of polar motion are defined by eqn [68]. The amplitude and phase of the prograde and retrograde components of polar motion excitation are similarly defined but with $\boldsymbol{\chi}(t) = \chi_x(t) + i\chi_y(t)$.

Source: Gross RS (2009) An improved empirical model for the effect of long-period ocean tides on polar motion. *Journal of Geodesy* 83: 635–644, http://dx.doi.org/10.1007/s00190-008-0277-y.

Table 11 Modeled variations in polar motion caused by the diurnal radiational tide in the atmosphere and oceans

Tide	*Fundamental argument*						*Period (solar days)*	$p_x(t)$ (μas)		$p_y(t)$ (μas)	
	γ	l	l′	F	D	Ω		*sin*	*cos*	*sin*	*cos*
P_1	1	0	0	−2	2	−2	1.00275	0.1 ± 0.4	0.0 ± 0.4	-0.0 ± 0.4	0.1 ± 0.4
S_1	1	0	−1	0	0	0	1.00000	8.3 ± 0.4	-3.4 ± 0.4	3.4 ± 0.4	8.3 ± 0.4
K_1	1	0	0	0	0	0	0.99727	-0.1 ± 0.5	0.5 ± 0.5	-0.5 ± 0.5	-0.1 ± 0.5

γ is GMST reckoned from the lower culmination of the vernal equinox (GMST + π). l, l', F, D, and Ω are the Delaunay arguments, expressions for which are given in Simon et al. (1994). The period, given in solar days, is the approximate period of the term as viewed in the terrestrial reference frame. Since the diurnal radiational tide is seasonally modulated, it also affects polar motion at the P_1 and K_1 tidal frequencies.

Source: Brzezinski A, Ponte RM, and Ali AH (2004) Nontidal oceanic excitation of nutation and diurnal/semidiurnal polar motion revisited. *Journal of Geophysical Research* 109: B11407, http://dx.doi.org/10.1029/2004JB003054.

3.09.4.2.4 The Chandler wobble and its excitation

Any irregularly shaped solid body rotating about some axis that is not aligned with its figure axis will freely wobble as it rotates (Euler, 1765). The Eulerian free wobble of the Earth is known as the Chandler wobble in honor of Seth Carlo Chandler, Jr., who first observed it (Chandler, 1891). Unlike the forced wobbles of the Earth, such as the annual wobble, whose periods are the same as the periods of the forcing mechanisms, the period of the free Chandler wobble is a function of the internal structure and rheology of the Earth (see Section 3.09.2.1) and its decay time constant, or quality factor Q, is a function of the dissipation mechanisms acting to dampen it. The observed values for the period and Q of the Chandler wobble can therefore be used to better understand the internal structure of the Earth and the dissipation mechanisms, such as mantle anelasticity, that dampen the Chandler wobble causing its amplitude to decay in the absence of excitation.

Determining an unbiased estimate for the period and Q of the Chandler wobble is complicated by the relatively short duration of the observational record and by incomplete and inaccurate models of the mechanisms acting to excite it. In the absence of any knowledge of its excitation, statistical models of the excitation can be adopted (e.g., Jeffreys, 1972; Ooe, 1978; Wilson and Haubrich, 1976; Wilson and Vicente, 1980, 1990). Since atmospheric, oceanic, and hydrologic processes are thought to be major sources of Chandler excitation, its period and Q have also been estimated by using AAM data (Furuya and Chao, 1996; Kuehne et al., 1996), atmospheric and oceanic angular momentum data (Gross, 2005b; Seitz and Kutterer, 2005; Seitz et al., 2012), and atmospheric, oceanic, and hydrologic angular momentum data (Wilson and Chen, 2005). By fitting a small set of basic Earth parameters to observed nutation amplitudes and precession rate, semianalytic estimates of the period and Q of the Chandler wobble have been determined by Mathews et al. (2002); a recent update of the estimates has been given by Chen and Shen (2010). Table 12 gives the resulting values for the period and Q of the Chandler wobble.

Table 12 Estimated period and Q of the Chandler wobble

Period (solar days)	*Q*	*Data span (years)*	*Source*
Statistical excitation			
433.2 ± 2.2	63 (36, 192)	67.6	(a)
434.0 ± 2.6	100 (50, 400)	70	(b)
434.8 ± 2.0	96 (50, 300)	76	(c)
433.3 ± 3.1	170 (47, 1000)	78	(d)
433.0 ± 1.1	**179 (74, 789)**	**86**	**(e)**
433.1 ± 1.7	–	93	(f)
Atmospheric excitation			
439.5 ± 2.1	72 (30, 500)	8.6	(g)
433.7 ± 1.8	49 (35, 100)	10.8	(h)
430.8	41	10	(i)
Atmospheric and oceanic excitation			
429.4	107	10	(i)
431.9	83	51	(i)
432.98	97	60	(j)
Semianalytic			
430.3	88.4	20	(k)
433.03	100.20	20	(l)

The recommended estimate is given in bold. The 1σ confidence interval for the Q estimates is given in parentheses.

Sources: (a) Jeffreys (1972), (b) Wilson and Haubrich (1976), (c) Ooe (1978), (d) Wilson and Vicente (1980), (e) Wilson and Vicente (1990), (f) Vicente and Wilson (1997), (g) Kuehne et al. (1996), (h) Furuya and Chao (1996), (i) Gross (2005b), (j) Seitz et al. (2012), (k) Mathews et al. (2002), (l) Chen and Shen (2010).

Gross (2005b) discussed the sensitivity of the estimated period and Q of the Chandler wobble to the length of the data sets that he analyzed, concluding that data sets spanning at least 31 years are needed to obtain stable estimates. He also noted the need for Monte Carlo simulations to determine corrections to the bias of the estimated Q values. Since Gross (2005b) did not do this, but since it was done by Wilson and Vicente (1990) who also used data spanning 86 years to estimate the period and Q of the Chandler wobble, the recommended estimate is that determined by them, namely, a period of 433.0 ± 1.1 (1σ) solar days and a Q of 179 with a 1σ range of 74–789.

With these recommended values for the period T and quality factor Q of the Chandler wobble, in the absence of excitation, it would freely decay to the minimum rotational energy state of rotation about the figure axis with an e-folding amplitude decay time constant $\tau=2QT/2\pi$ of about 68 years. But a damping time of 68 years is short on a geologic timescale and since the amplitude of the Chandler wobble has at times been observed to actually increase, some mechanism or mechanisms must be acting to excite it. Since its discovery, many possible excitation mechanisms have been studied, including core–mantle interactions (Gire and Le Mouël, 1986; Hinderer et al., 1987, 1990; Jault and Le Mouël, 1993; Nakada, 2009a, 2011), earthquakes (Gross, 1986; Souriau and Cazenave, 1985), continental water storage (Brzezinski et al., 2012; Chao et al., 1987; Hinnov and Wilson, 1987; Jin et al., 2010; Kuehne and Wilson, 1991; Liao et al., 2004), atmospheric wind and surface pressure variations (Aoyama, 2005; Aoyama and Naito, 2001; Aoyama et al., 2003; Furuya et al., 1996, 1997; Liao et al., 2007; Stuck et al., 2005; Wahr, 1983; Wilson and Haubrich, 1976), and oceanic current and bottom pressure variations (Bizouard et al., 2011; Brzezinski and Nastula, 2002; Brzezinski et al., 2002b; Gross, 2000b; Gross et al., 2003; Liao, 2005; Liao et al., 2003; Nastula et al., 2012; Ponte, 2005; Ponte and Stammer, 1999; Seitz and Schmidt, 2005; Seitz et al., 2004, 2005; Thomas et al., 2005; Zotov and Bizouard, 2012).

While there is growing agreement that the Chandler wobble is excited by a combination of atmospheric, oceanic, and hydrologic processes, the relative contribution of each process to its excitation is still being debated. For example, Gross et al. (2003) studied the excitation of the Chandler wobble during 1980–2000, finding that atmospheric winds and surface pressure and oceanic currents and bottom pressure combined are significantly coherent with and have enough power to excite the Chandler wobble. They found that during this time interval, the observed power in the Chandler band is 1.90 mas^2, with the sum of all atmospheric and oceanic excitation processes having more than enough power, at 2.22 mas^2, to excite the Chandler wobble. Ocean-bottom pressure variations were found to be the single most effective process exciting the Chandler wobble, with atmospheric surface pressure variations having about 2/3 as much power as ocean-bottom pressure variations and with the power of winds and currents combined being less than 1/3 the power of the combined effects of surface and bottom pressure. The conclusion that oceanic processes are more effective than atmospheric in exciting the Chandler wobble has also been reached by Gross (2000b), Brzezinski and Nastula (2002), Brzezinski et al. (2002b, 2012), and Liao et al. (2003, 2004). Studies of the excitation of the Chandler wobble using coupled atmosphere–ocean climate models also confirm that atmospheric and oceanic processes have enough power to maintain the Chandler wobble and indicate that the relative contribution of individual atmospheric and oceanic processes changes with time (Celaya et al., 1999; Leuliette and Wahr, 2002; Ponte et al., 2002). However, other studies have concluded that atmospheric processes alone have enough power to excite the Chandler wobble (Aoyama, 2005; Aoyama and Naito, 2001; Aoyama et al., 2003; Furuya et al., 1996, 1997). The resolution of the debate about the relative importance of atmospheric, oceanic, and hydrologic processes to exciting the Chandler wobble awaits the availability of more accurate models of these processes and, because the power needed to maintain the Chandler wobble depends upon its damping (Gross, 2000b), of more accurate estimates of its period and Q.

3.09.4.2.5 Seasonal wobbles

While continuing to analyze variation of latitude measurements, Chandler (1892) soon discovered that the wobbling motion of the Earth includes an annual component in addition to the 14-month component. The annual wobble is a forced wobble of the Earth that is caused largely by the annual appearance of a high atmospheric pressure system over Siberia every winter (e.g., Eubanks, 1993; Lambeck, 1980, 1988; Munk and MacDonald, 1960).

Chao and Au (1991) showed that during 1980–88, the amplitude of the prograde annual polar motion excitation can be accounted for by atmospheric wind and pressure fluctuations, with equatorial winds contributing about 25% and pressure fluctuations contributing about 75% to the total atmospheric excitation. But even though the amplitude of the observed prograde annual polar motion excitation is consistent with atmospheric wind and pressure fluctuations, there is a rather large phase discrepancy of about 30°. Furthermore, no agreement was found by Chao and Au (1991) between the observed retrograde annual polar motion excitation and that due to atmospheric processes, with atmospheric excitation being about twice as large as the observed excitation. Discrepancies as large as a factor of two in amplitude were also found by Chao and Au (1991) between observed semiannual polar motion excitation and that due to atmospheric processes (also see Aoyama and Naito, 2000; Barnes et al., 1983; Chao, 1993; Dobslaw et al., 2010; King and Agnew, 1991; Kolaczek et al., 2003; Merriam, 1982; Nastula and Kolaczek, 2002; Nastula and Salstein, 2012; Nastula et al., 2005, 2009; Neef and Matthes, 2012; Stuck et al., 2005; Wahr, 1983; Wilson and Haubrich, 1976; Zhou et al., 2006, 2008).

Since the agreement with atmospheric excitation is poor except for the prograde annual amplitude, other processes must be contributing to the excitation of seasonal wobbles. Recently, near-global general circulation models of the oceans have been used to investigate the contribution that nontidal oceanic processes make to exciting the seasonal wobbles of the Earth (Brzezinski et al., 2005; Chen et al., 2004a; Dobslaw et al., 2010; Furuya and Hamano, 1998; Gross, 2004; Gross et al., 2003; Johnson, 2005; Johnson et al., 1999; Liu et al., 2007; Nastula et al., 2000, 2003, 2012; Ponte and Stammer, 1999; Ponte et al., 1998, 2001; Seitz and Schmidt, 2005; Seitz et al., 2004; Wünsch, 2000; Zhong et al., 2006; Zhou et al., 2005). These studies have shown that adding nontidal oceanic excitation to atmospheric improves the agreement with the observed excitation. For example, **Table 13** gives the results obtained by Gross et al. (2003) for the atmospheric excitation and oceanic excitation of polar motion at the annual and semiannual frequencies. They found that atmospheric processes were more effective than oceanic in exciting the annual and semiannual wobbles. Atmospheric surface pressure variations were found to be the single most important mechanism exciting the annual and semiannual wobbles, with the sum of surface and ocean-bottom pressure variations being about two to three times as effective as the sum of winds and currents.

A rather large residual remains after the effects of the atmosphere and oceans are removed from the observed seasonal polar motion excitation. This residual not only is probably at least partly due to errors in the atmospheric and oceanic models but also could be due to the neglect of other excitation processes such as hydrologic processes (Chao and O'Connor, 1988; Chen and Wilson, 2005; Chen et al., 2000a; Dobslaw et al., 2010; Hinnov and Wilson, 1987; Kuehne and Wilson, 1991; Nastula and Kolaczek, 2005; Nastula and Salstein, 2012; Wünsch, 2002). Wünsch (2002) summarized the contribution

Table 13 Observed and modeled nontidal polar motion excitation at annual and semiannual frequencies

Excitation process	*Annual*				*Semiannual*			
	Prograde		*Retrograde*		*Prograde*		*Retrograde*	
	Amplitude (mas)	*Phase (°)*	*Amplitude (mas)*	*Phase (°)*	*Amplitude (mas)*	*Phase (°)*	*Amplitude (mas)*	*Phase (°)*
Observed excitation	14.52 ± 0.33	−62.77 ± 1.30	7.53 ± 0.33	−120.09 ± 2.50	5.67 ± 0.33	107.56 ± 3.32	5.80 ± 0.33	123.66 ± 3.24
Atmospheric excitation								
Winds	2.97 ± 0.12	−34.92 ± 2.30	2.04 ± 0.12	12.06 ± 3.36	0.36 ± 0.12	71.76 ± 19.0	0.55 ± 0.12	−134.43 ± 12.4
Surface pressure (IB)	15.12 ± 0.17	−101.92 ± 0.66	15.05 ± 0.17	−105.30 ± 0.66	2.60 ± 0.17	47.45 ± 3.81	4.75 ± 0.17	103.77 ± 2.08
Winds and surface pressure	16.51 ± 0.23	−92.38 ± 0.81	14.23 ± 0.23	−98.00 ± 0.94	2.93 ± 0.23	50.36 ± 4.58	4.49 ± 0.23	109.77 ± 2.99
Oceanic excitation								
Currents	2.31 ± 0.09	39.72 ± 2.13	2.11 ± 0.09	50.70 ± 2.33	1.32 ± 0.09	176.16 ± 3.71	1.41 ± 0.09	−143.46 ± 3.49
Bottom pressure	3.45 ± 0.11	63.18 ± 1.87	3.42 ± 0.11	110.24 ± 1.89	0.77 ± 0.11	133.89 ± 8.41	1.48 ± 0.11	−136.46 ± 4.35
Currents and bottom pressure	5.64 ± 0.16	53.81 ± 1.61	4.84 ± 0.16	88.22 ± 1.87	1.96 ± 0.16	160.89 ± 4.62	2.89 ± 0.16	−139.87 ± 3.14
Atmospheric and oceanic excitation								
Winds and currents	4.22 ± 0.15	−3.09 ± 2.08	3.91 ± 0.15	31.71 ± 2.24	1.28 ± 0.15	160.38 ± 6.83	1.96 ± 0.15	−140.92 ± 4.48
Surface and bottom pressure	11.82 ± 0.20	−97.61 ± 0.99	12.43 ± 0.20	−114.51 ± 0.94	2.75 ± 0.20	63.61 ± 4.26	4.22 ± 0.20	121.55 ± 2.78
Total of all modeled excitation	12.23 ± 0.28	−77.50 ± 1.32	9.43 ± 0.28	−101.18 ± 1.71	2.90 ± 0.28	89.70 ± 5.57	4.41 ± 0.28	147.63 ± 3.66

IB, inverted barometer. The amplitude and phase of the prograde and retrograde components of polar motion excitation are defined by eqn [68] but with $\boldsymbol{\chi}(t) = \chi_x(t) + i\chi_y(t)$. The reference date to the phase is 1 January 1990.
Source: Gross RS, Fukumori I, and Menemenlis D (2003) Atmospheric and oceanic excitation of the Earth's wobbles during 1980–2000. *Journal of Geophysical Research* 108(B8): 2370. http://dx.doi.org/10.1029/2002JB002143.

of soil moisture and snow load to exciting the annual and semiannual wobbles. Although the available soil moisture models exhibit large differences, Wünsch (2002) concluded that soil moisture and snow load effects are important contributors to exciting the annual and semiannual wobbles. Climate models have also been used to study atmospheric, oceanic, and hydrologic excitation of seasonal polar motion (Celaya et al., 1999; Neef and Matthes, 2012; Ponte et al., 2002; Zhong et al., 2003).

3.09.4.2.6 Nonseasonal wobbles

Like the seasonal wobbles, the wobbling motion of the Earth on interannual timescales is a forced response of the Earth to its excitation mechanisms. Abarca del Rio and Cazenave (1994) compared the observed excitation of the Earth's wobbles to atmospheric excitation during 1980–91, finding similar fluctuations both in components on timescales between 1 and 3 years and in the *y*-component on timescales between 1.2 and 8 years, but only when the atmospheric excitation is computed assuming the oceans fully transmit the imposed atmospheric pressure variations to the floor of the oceans (rigid ocean approximation). Chao and Zhou (1999) studied the correlation of the observed polar motion excitation functions on interannual timescales during 1964–94 with the SOI and the North Atlantic Oscillation Index (NAOI). Although little agreement was found with the SOI, a significant agreement was found with the NAOI, especially for the *x*-component of the observed excitation function, indicating a possible meteorologic origin of the interannual wobbles (also see Nastula et al., 2009; Neef and Matthes, 2012; Zhou et al., 1998). The lack of an ENSO signal in polar motion, as evidenced by the lack of a correlation between the SOI and interannual polar motion variations found by Chao and Zhou (1999), was explained by Marcus et al. (2010) as being a consequence of the mitigation of atmospheric ellipsoidal torques by the geometric distribution of the land and oceans.

Johnson et al. (1999) compared the observed polar motion excitation functions on interannual timescales during 1988–98 with atmospheric and oceanic excitation functions, finding only weak agreement between the observed and modeled excitation functions. For periods between 1 year and 6 years excluding the annual cycle, oceanic processes were found by Gross et al. (2003) to be much more important than atmospheric in exciting interannual polar motions, explaining 33% of the observed variance compared with 6% for atmospheric processes and having a correlation coefficient of 0.59 with the observations compared with 0.28 for atmospheric processes. Ocean-bottom pressure variations were found to be the single most important excitation mechanism, explaining 30% of the observed variance and having a correlation coefficient of 0.59 with the observations. Although the other processes were not nearly as effective as ocean-bottom pressure variations in exciting interannual polar motions, when all the processes were combined, they were found to explain 40% of the observed variance and to have a correlation coefficient of 0.64 with the observations (also see Brzezinski et al., 2005; Chen and Wilson, 2005; Johnson, 2005).

Like the seasonal and interannual wobbles, the wobbling motion of the Earth on intraseasonal timescales is a forced response of the Earth to its excitation mechanisms. Eubanks et al. (1988) were the first to study the Earth's wobbles on timescales between 2 weeks and several months, concluding that these rapid polar motions during 1983.75–86.75 are at least partially driven by atmospheric surface pressure changes and suggesting that the remaining excitation may be caused by dynamic ocean-bottom pressure variations. Subsequent studies confirmed the importance of atmospheric processes in exciting rapid polar motions (di Leonardo and Dickman, 2004; Feldstein, 2008; Gross and Lindqwister, 1992; Kolaczek et al., 2000a,b; Kosek et al., 1995; Kuehne et al., 1993; Nastula, 1992, 1995, 1997; Nastula and Salstein, 1999; Nastula et al., 1990, 1997, 2009; Neef and Matthes, 2012; Salstein, 1993; Salstein and Rosen, 1989; Schuh and Schmitz-Hübsch, 2000; Stieglitz and Dickman, 1999; Zhou et al., 2006, 2008), although the existence of significant discrepancies indicates that nonatmospheric processes must also play an important role.

The contribution of oceanic processes to exciting rapid polar motions has been studied using both barotropic (Nastula and Ponte, 1999; Ponte, 1997) and baroclinic (Bizouard and Seoane, 2010; Brzezinski et al., 2005; Chen et al., 2004a; de Viron et al., 2004a; Gross et al., 2003; Johnson et al., 1999; Lambert et al., 2006; Nastula and Kolaczek, 2005; Nastula et al., 2000; Ponte et al., 1998, 2001; Zhou et al., 2005) models of the oceans. Such studies have shown that while better agreement with the observations is obtained when oceanic excitation is added to that of the atmosphere, significant discrepancies still remain. For example, Gross et al. (2003) found that on intraseasonal timescales, with periods between 5 days and 1 year excluding the seasonal cycles, atmospheric processes were found to be more effective than oceanic in exciting polar motion, explaining 45% of the observed variance compared with 19% for oceanic processes and having a correlation coefficient of 0.67 with the observations compared with 0.44 for oceanic processes. Of these processes, atmospheric surface pressure variations were found to be the single most effective process exciting intraseasonal polar motions, with winds being the next most important process; ocean-bottom pressure was about half as effective as atmospheric surface pressure, and currents were about half as effective as winds. Atmospheric winds and surface pressure and oceanic currents and bottom pressure combined explained 65% of the observed variance and had a correlation coefficient of 0.81 with the observations.

On the shortest intraseasonal timescales, Ponte and Ali (2002) used the NCEP/NCAR reanalysis atmospheric model and a barotropic ocean model forced by the NCEP/NCAR reanalysis surface pressure and wind stress fields to study atmospheric and oceanic excitation of polar motion from July 1996 to June 2000 on timescales of 2–20 days. They confirmed the importance of oceanic processes in exciting rapid polar motions, finding that about 60% of the observed polar motion excitation variance between periods of 6 and 20 days is explained by atmospheric excitation, increasing to about 80% when oceanic excitation is added to that of the atmosphere (see also Bizouard and Seoane, 2010; Johnson, 2008; Kouba, 2005; Lambert et al., 2006; Nastula and Kolaczek, 2007; Nastula et al., 2002).

A spectra of subdaily atmospheric and oceanic excitation functions (Brzezinski et al., 2002a; Gross, 2005a) reveal the presence of atmospheric normal modes: (1) a peak at about −0.83 cycles per day (cpd) due to the $\psi_1{}^1$ atmospheric normal mode (Masaki, 2008; Sidorenkov, 2003); (2) a peak of smaller amplitude at about +1.8 cpd due to the $\xi_2{}^1$ atmospheric normal mode; and (3) slight enhancement in power at about −0.12 cpd due to the $\psi_3{}^1$ atmospheric normal mode (Brzezinski, 1987; Feldstein, 2006). Because of their small amplitudes, the effects of these normal modes on polar motion are difficult to detect, although the effect of the ψ_3^1 mode has been detected (Bizouard and Seoane, 2010; Eubanks et al., 1988; Feldstein, 2008).

Apart from errors in observations and models, the residual that remains on broad intraseasonal to interannual timescales after modeled atmospheric and oceanic effects have been removed from the observations may be caused by hydrologic processes (e.g., Fernández et al., 2007; Göttl and Seitz, 2009).

3.09.5 Mass Redistribution, Gravity, and Earth Rotation

Redistribution of mass within the Earth system causes the Earth's gravitational field to change and, by changing the Earth's inertia tensor, causes the Earth's rotation to change. Because the elements of the Earth's inertia tensor are related to the degree-2 coefficients of the Earth's gravitational field, measurements of these coefficients can be used to study changes in the Earth's rotation caused by mass redistribution. The relations between the degree-2 coefficients of the gravitational field and the elements of the inertia tensor are derived here and the results of using gravitational field measurements to study changes in the Earth's rotation are reviewed.

In geodesy, the gravitational potential $U(\mathbf{r}_o)$ of the Earth evaluated at some external field point $\mathbf{r}_o(r_o, \theta_o, \phi_o)$ is usually given in the form (e.g., Kaula, 1966)

$$U(\mathbf{r}_o) = \frac{GM}{r_o}\sum_{l=0}^{\infty}\sum_{m=0}^{l}\left(\frac{a}{r_o}\right)^l (C_{lm}\cos m\phi_o + S_{lm}\sin m\phi_o)\widetilde{P}_{lm}(\cos\theta_o) \quad [73]$$

where G is the gravitational constant, M is the mass of the Earth, a is some reference radius that is less than r_o and that is usually taken to be the radius of some spherically symmetrical Earth model, and $\widetilde{P}_{lm}$ are the normalized associated Legendre functions of degree l and order m. The dimensionless expansion coefficients C_{lm} and S_{lm}, known as Stokes coefficients, are the normalized multipole moments of the Earth's density field $\rho(\mathbf{r})$:

$$C_{lm} + iS_{lm} = \frac{N_{lm}}{Ma^l}\int_{V_o} r^l\, Y_{lm}(\theta,\phi)\,\rho(\mathbf{r})\,dV \quad [74]$$

where the integral extends over the volume V_o of the Earth, the Y_{lm} are the fully normalized surface spherical harmonic functions, and the N_{lm} are normalization factors (Chao and Gross, 1987):

$$N_{lm} = (-1)^m \frac{2}{2l+1}\sqrt{(2-\delta_{m0})\pi} \quad [75]$$

In Cartesian coordinates, the degree-2, order-0, and order-1 Stokes coefficients can be written as

$$C_{20} = \frac{1}{\sqrt{20}Ma^2}\int_{V_o}\left[2r^2 - 3\left(x^2+y^2\right)\right]\rho(\mathbf{r})\,dV \quad [76]$$

$$C_{21} + i\,S_{21} = \sqrt{\frac{3}{5}}\frac{1}{Ma^2}\int_{V_o}(xz + i\,yz)\,\rho(\mathbf{r})\,dV \quad [77]$$

The inertia tensor $\mathbf{I}$ is defined as

$$\mathbf{I} = \int_{V_o}\left(r^2\mathbf{I} - \mathbf{rr}\right)\rho(\mathbf{r})\,dV \quad [78]$$

where $\mathbf{I}$ is the identity tensor. In Cartesian coordinates, the elements of the inertia tensor that, when they change, affect the Earth's rotation (see eqns [38]–[43]) can be written as

$$I_{xz} = -\int_{V_o} xz\,\rho(\mathbf{r})\,dV \quad [79]$$

$$I_{yz} = -\int_{V_o} yz\,\rho(\mathbf{r})\,dV \quad [80]$$

$$I_{zz} = \int_{V_o}\left(x^2+y^2\right)\rho(\mathbf{r})\,dV \quad [81]$$

Thus, the desired relations between the degree-2 Stokes coefficients and those elements of the inertia tensor that affect the Earth's rotation are

$$C_{20} = \frac{1}{\sqrt{20}Ma^2}(T - 3I_{zz}) \quad [82]$$

$$C_{21} + iS_{21} = -\sqrt{\frac{3}{5}}\frac{1}{Ma^2}\left(I_{xz} + iI_{yz}\right) \quad [83]$$

where T is the trace of the inertia tensor:

$$T = I_{xx} + I_{yy} + I_{zz} = \int_{V_o} 2r^2\rho(\mathbf{r})\,dV \quad [84]$$

Rochester and Smylie (1974) had shown that the trace of the inertia tensor T does not change for many types of mass redistribution including those for which the total mass of the system does not change. For example, the trace of the inertia tensor does not change if the system consists of the entire atmospheric, oceanic, and land–hydrologic system because the mass of this entire system does not change as water in its various phases moves between the atmosphere, oceans, and land. However, if the system consists of just, say, the atmosphere, then the trace of the inertia tensor does change because the mass of the atmosphere changes as water evapotranspires into and precipitates out of the atmosphere.

Equations [82] and [83] are quite general. But applying them to processes that load the solid Earth is a bit cumbersome because the integrals in eqns [76], [77], and [79]–[81] extend over the entire volume of the Earth and thus include the deformation of the solid Earth caused by the loading process. That is, eqns [82] and [83] apply to the sum of the loading process and the solid Earth's response to the load. It is usually more convenient to separate these and account for the deformational response of the solid Earth to the load by using load Love numbers. In this case, eqn [74] for the degree-2 Stokes coefficients can be written as

$$C_{2m} + iS_{2m} = \left(1 + k_2' + \Delta k_{an}'\right)\frac{N_{2m}}{Ma^2}\int_{V_l} r^2\, Y_{2m}(\theta,\phi)\,\rho(\mathbf{r})\,dV \quad [85]$$

where the integral now extends over just the volume V_l of the load, k'_2 is the second-degree load Love number, and $\Delta k'_{an}$

accounts for the effects of mantle anelasticity on the second-degree load Love number. Applying eqns [79]–[81] to just the load, the desired relations between time-dependent changes in the degree-2 Stokes coefficients and the relevant elements of the inertia tensor of the load are

$$\Delta C_{20} = \frac{1 + k_2' + \Delta k_{\mathrm{an}}'}{\sqrt{20} M a^2} \left(\Delta T^l - 3 \Delta \mathbf{I}_{zz}^l \right) \qquad [86]$$

$$\Delta C_{21} + i \Delta S_{21} = -\left(1 + k_2' + \Delta k_{an}'\right) \frac{\sqrt{3/5}}{M a^2} \left(\Delta \mathbf{I}_{xz}^L + i \Delta \mathbf{I}_{yz}^L \right) \qquad [87]$$

where the time dependence has been suppressed for brevity and the superscript l indicates that the quantity is for just the load, not the entire Earth.

Load Love numbers are also used to account for the effect on the Earth's rotation of the deformation of the solid Earth caused by surficial loading processes (see eqns [41]–[43]). Thus, as in eqns [86] and [87], the changes in the elements of the inertia tensor in eqns [41]–[43] are for just the load. By combining eqns [41]–[43] and [86] and [87] after setting the relative angular momenta terms $h_i(t)$ in eqns [41]–[43] to zero, the following relation between the Earth rotation excitation functions and the degree-2 Stokes coefficients can be derived:

$$\Delta C_{20}(t) = \frac{1 + k_2' + \Delta k_{\mathrm{an}}'}{\sqrt{20} M a^2} \left[\Delta T^l(t) - \frac{3 C_m / 0.997}{1 + \alpha_3 \left(k_2' + \Delta k_{an}'\right)} \chi_z(t) \right] \qquad [88]$$

$$\Delta C_{21}(t) + i \Delta S_{21}(t) = -\frac{\sqrt{3/5}}{M a^2} \frac{C - A'}{1.608} \left[\chi_x(t) + i \chi_y(t) \right] \qquad [89]$$

Equations [88]–[89] can be used to relate observed changes in the degree-2 Stokes coefficients to changes in the Earth rotation excitation functions caused by redistribution of mass within the Earth's surface geophysical fluids. Observations of the time-dependent degree-2 Stokes coefficients are currently being provided by SLR and the Gravity Recovery and Climate Experiment (GRACE) twin satellite mission. The use of these gravitational field observations to study changes in the Earth's rotation is discussed in the succeeding text.

3.09.5.1 GRACE and SLR Measurements of Earth Rotation Excitation

Spatial and temporal variations in the Earth's gravitational field cause perturbations to the motion of satellites as they orbit the Earth. Such orbit perturbations can be determined from ranging measurements, either ranging between ground stations and the satellites as in the case of SLR (see Section 3.09.3.3.3) or ranging between satellites as in the case of the GRACE twin satellites (Tapley et al., 2004b). Variations in the Earth's gravitational field are routinely determined from such range measurements, and numerous studies have been conducted that compare the temporal variations in the degree-2 coefficients of the gravitational field to Earth rotation measurements from which motion effects have been modeled and removed.

In a series of studies, J. Chen and colleagues compared degree-2 gravitational field coefficients from SLR and GRACE to Earth rotation measurements and models of surface geophysical fluid mass redistribution (Chen and Wilson, 2003, 2008, 2010, 2012; Chen et al., 2000b, 2004b, 2005, 2012). The availability of three independent estimates of the degree-2 Stokes coefficients (from SLR, GRACE, and EOP via eqns [88] and [89]) and models of surface mass redistribution is essential for verifying and validating the measurements and models and for improving our understanding of the large-scale mass redistribution within and between the atmosphere, oceans, and water stored on land (also see, e.g., Bourda, 2008; Gross et al., 2009; Hancock and Moore, 2007; Jin et al., 2011; Nastula et al., 2007b; Seoane et al., 2009, 2012).

The relatively high spatial resolution of the GRACE gravitational field measurements allows individual Earth rotation excitation processes to be isolated. For example, the individual effects on the Earth's rotation of changes in the mass distribution of the oceans (e.g., Göttl et al., 2012; Jin et al., 2010) and of water stored on land (e.g., Brzezinski et al., 2009; Fernández, 2009; Jin et al., 2010, 2012; Nastula and Salstein, 2012; Seoane et al., 2011) can be studied with GRACE measurements after atmospheric effects are removed from them using a model. The availability of GRACE measurements, in addition to models, of individual Earth rotation excitation processes is leading to major advancements in our knowledge of the relative importance of those processes in causing Earth rotation variations.

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