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Causes of fluctuations in the rotation of the Earth

By M. G. Rochester

Department of Earth Sciences, Memorial University of Newfoundland, St. John's, Newfoundland A1B 3X7, Canada

Over the last decade considerable progress has been made in closing the gap between geophysical theory and the observed changes in the Earth's spin rate and polar motion, which are affected by the Earth's internal structure, properties and processes. New observational techniques and data have had a marked impact on understanding the short-term variations in the length of day and pole position, but we can expect between another 5 and 30 years to pass before they permit improved explanations of longer-term phenomena (18.6 year tidal effects, the Markowitz wobble, decade fluctuations in the length of day, etc.) This review summarizes recent advances, problems newly illuminated by recent Earth rotation data (for example mantle anelasticity, core–mantle coupling) and problems still unresolved.

1. Changes in the Earth's rotation: an overview

Some of the most complex and intriguing interplays of theory and observation in planetary physics appear when we study the geophysical effects on, and consequences of, the variability of the Earth's rotation vector $\boldsymbol{\omega}$. The subject is the theme of Lambeck's (1980) monograph and a more recent brief review by Merriam (1983). In this paper I shall outline recent progress towards understanding a few of the more prominent features of the spectrum of changes in $\boldsymbol{\omega}$.

It will be helpful to clarify the relevant terminology. The Conventional Terrestrial System (C.T.S.) is a rigid reference frame with coordinate axes $\boldsymbol{e}_i$ $(i = 1, 2, 3)$, rotating in inertial space with angular velocity $\boldsymbol{\omega}$. The terms *precession* and *nutation* describe changes in the orientation of the C.T.S. with respect to inertial space (in practice, relative to a quasi-inertial frame tied to the ecliptic and equinox at a certain epoch). Such changes are almost entirely *externally forced*, by the gravitational torques of the Moon and Sun on the spinning Earth's equatorial bulge. The $\boldsymbol{e}_3$ axis of the C.T.S cuts the celestial sphere in the Conventional International Origin (C.I.O.). The departure of $\boldsymbol{\omega}$ from alignment with $\boldsymbol{e}_3$ is small (under 0.4″) but of great geophysical interest because it varies. Such *polar motion*, whether periodic (*wobble*) or aperiodic (*polar wander*) reflects primarily events produced *within* the deformable Earth. However a small part of polar motion is the 'dynamical variation of latitude' as $\boldsymbol{\omega}$ traces out a nearly diurnal retrograde circuit in the C.T.S., of amplitude *ca.* 0.02″, due to the forced nutations. The Celestial Ephemeris Pole (C.E.P.) is the rotation pole with these effects removed using Wahr's (1981) theory of the forced nutations of a realistic Earth model. Beginning in January 1984, the pole coordinates x, y disseminated by the Bureau Internationale de l'Heure (B.I.H.) are essentially the $\boldsymbol{e}_1$, $-\boldsymbol{e}_2$ coordinates of the C.E.P. The C.T.S. is currently defined by the adopted latitudes and longitudes of a set of astronomical observatories in 1899, when the International Latitude Service (I.L.S.) began to monitor polar motion.

Fluctuations in the spin rate $|\boldsymbol{\omega}|$ give rise to *changes in the length of day* (l.o.d.), reflected in the non-uniform passage of U.T. (the sidereal time kept by the rotating C.T.S.) with respect to International Atomic Time (T.A.I). Removal of the effects of polar motion, with the Wahr nutation theory, gives U.T.1.

Standard geophysical practice is to define the small quantities $\boldsymbol{m} = \boldsymbol{\omega}/\Omega - \boldsymbol{e}_3$,

$$\boldsymbol{J} = \boldsymbol{I} - \begin{pmatrix} A & 0 & 0 \\ 0 & A & 0 \\ 0 & 0 & C \end{pmatrix},$$

$\boldsymbol{h} = \boldsymbol{H} - \boldsymbol{I}\cdot\boldsymbol{\omega}$, where $\boldsymbol{H}$, $\boldsymbol{I}$ are the total angular momentum and inertia tensor of the deformable Earth, and A, C are the equatorial and polar moments of inertia of the axially symmetric reference Earth model, assumed to be rotating uniformly with spin rate Ω about $\boldsymbol{e}_3$. The net external torque on the Earth is $\boldsymbol{N} = \mathrm{D}\boldsymbol{H}$, where the operator D denotes differentiation with respect to T.A.I. For brevity we define the complex quantities

$$\tilde{m}, \tilde{N}, \tilde{h}, \tilde{J} = (\boldsymbol{e}_1 + \mathrm{i}\boldsymbol{e}_2)\cdot(\boldsymbol{m}, \boldsymbol{N}, \boldsymbol{h}, \boldsymbol{J}\cdot\boldsymbol{e}_3).$$

If only first-order terms in the small quantities are retained, the observational data distributed by the B.I.H. are related to the m_i of geophysical interest as follows:

$$x - \mathrm{i}y = (\tilde{m})_{\tilde{N}=0}, \quad \mathrm{D}(\mathrm{U.T.1} - \mathrm{T.A.I.}) = m_3;$$

and the latter are governed by the linearized Liouville equations

$$A(\mathrm{D} - \mathrm{i}\sigma_r)\,\tilde{m} = \tilde{N}/\Omega - (\mathrm{D} + \mathrm{i}\Omega)\,(\tilde{J} + \tilde{h}/\Omega),$$

$$C\mathrm{D}m_3 = N_3/\Omega - \mathrm{D}(J_{33} + h_3/\Omega),$$

where $\sigma_r = (C-A)\,\Omega/A$ is the Euler frequency of the torque-free ($\boldsymbol{N} = 0$) wobble of a rigid Earth ($\boldsymbol{J} = \boldsymbol{h} = 0$). By appropriately modifying the parameters, similar Liouville equations can be written for subsystems of the whole Earth (e.g. the liquid core, the atmosphere) if it seems easier to calculate the torques between them and the solid Earth than their relative angular momenta. The Liouville equations display explicitly the effects on the pole position and spin rate of external torques and the rearrangement of both mass (earthquakes, sea level changes, seasonal atmospheric mass redistribution, tidal deformation) and angular momentum (winds, ocean currents, motions of the liquid core) within the C.T.S.

The principal features of the spectrum of changes in the spin rate and polar motion are shown in figures 1–3. I shall refrain from more than passing comment on: (*a*) potential effects not yet unambiguously observed, e.g. the 460-day principal free core nutation (Sasao & Wahr 1981; Wahr & Larden 1983); (*b*) phenomena now apparently well understood, e.g. the effects of core fluidity and mantle elasticity on the forced nutation amplitudes, modelled adequately for current observational accuracy by Wahr's (1981) theory; (*c*) features of the rotation spectrum discussed in other papers presented here: long-term changes in the l.o.d. due to tidal friction and deglaciation (Stephenson & Morrison, this symposium); short-term variations in spin rate and polar motion due to the exchange of angular momentum between the solid Earth and the global atmospheric wind system (Hide, this symposium); rotation of the solid inner core (Smylie & Szeto, this symposium).

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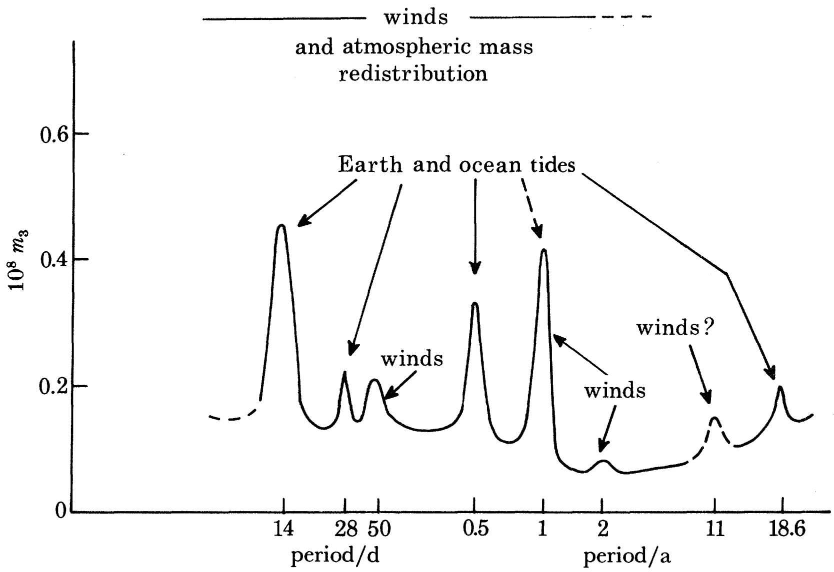

FIGURE 1. Spectrum of short-term changes in length of day.

2. FLUCTUATIONS IN SPIN RATE

(a) *Tidal terms in the l.o.d.*

Atomic timekeeping has made possible the detection of the forced changes in the l.o.d. due to the Earth's deformation in response to the fortnightly, monthly, and semi-annual zonal tides. Merriam (1980, 1982), Wahr *et al.* (1981) and Yoder *et al.* (1981) have given improved theoretical descriptions of the effects on m_3 of this deformation (measured by a linear combination of second-degree Love numbers called the zonal response coefficient κ) by taking into account the non-global ocean's tidal response and loading, and the apparent decoupling of the fluid core from the mantle for axial perturbations at these frequencies. As Wahr *et al.* remark, the core may be coupled strongly enough at the 18.6 year period (§2(*c*)) for this assumption to require modification for that tidal term. Introduction of the Wahr nutation series into the B.I.H. processing of Earth rotation data, together with the advent of high-quality data for the angular momentum of the atmosphere during 1976–1982 (Rosen & Salstein 1983), permits highly accurate estimates of κ from U.T.1 after zonal wind effects are removed (Merriam 1984*a*). If attributable to dispersion (the effect of anelasticity on rigidity) the differences between the resulting values of κ for the fortnightly and monthly tidal terms in U.T.1 can be used to constrain the frequency dependence of mantle anelasticity, measured by the dissipation factor Q. The physical principle involved is interesting. If Q varies with frequency as σ^{α} then for large Q, and low frequency, the effect of dispersion on the amplitude of a forced oscillation (such as an Earth tide) appears as an increment proportional to $\sigma^{-\alpha}$, whereas the phase lag varies as $\sigma^{1-\alpha}$. At periods long compared with those at which seismic rigidity is measured, amplitude (i.e. κ) is a more sensitive indicator of α than is lag. Merriam (1984*a*) finds $0 < \alpha < 0.3$.

The Earth's deformation in response to a zonal tide changes the second-degree coefficient J_2 in the harmonic expansion of its gravitational potential, and causes a corresponding periodic change in the node of an Earth satellite orbit. The change in gravitational potential of the solid Earth plus oceans is measured by a second-degree tidal effective Love number $\bar{k}_2$. Laser tracking of the LAGEOS satellite over a 5.5 year interval after its launching in 1976 reveals an acceleration of the node with 0.5 year period due to the solar semi-annual tide (Yoder *et al.* 1983). Merriam (1984*b*) shows that the value of $\bar{k}_2$ deduced from the amplitude of this fluctuation in the node also requires an upper limit of 0.3 on the mantle anelasticity parameter α.

(*b*) *Long-term non-tidal changes in the l.o.d.*

Laser ranging to LAGEOS shows a residual acceleration of the node, after tidal effects are removed, which Yoder *et al.* (1983) interpret as evidence for a secular decrease in J_2 and therefore in the Earth's polar moment of inertia C. Their value for the accompanying non-tidal acceleration of the Earth, $\mathrm{D}m_3 = 7 \times 10^{-11}\ \mathrm{a}^{-1}$, is in good agreement with that found by Morrison & Stephenson (1982) from ancient Babylonian lunar eclipse observations and lunar occultations since 1620, especially if the most recent determination of the Moon's tidal acceleration by lunar laser ranging (Dickey *et al.* 1983) is used. No allowance is made in these calculations for a changing gravitational constant. If this decrease in J_2 is indeed a snapshot of the secular decrease predicted to accompany the Pleistocene deglaciation (Peltier 1982), it provides an additional strong constraint on lower mantle viscosity (Peltier 1983).

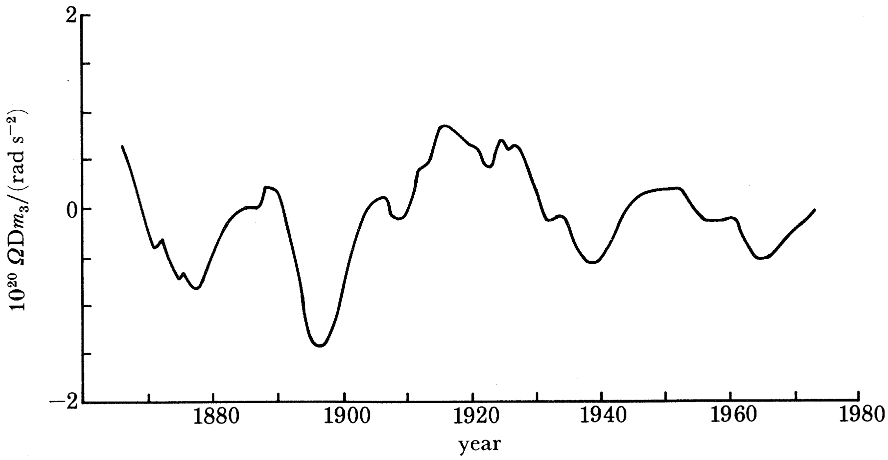

FIGURE 2. Decade fluctuations in the length of day.

(*c*) *Decade fluctuations in the l.o.d.*

Between the short-term (under approximately 5 years) changes in m_3, apparently almost entirely due to varying global atmospheric angular momentum (Hide *et al.* 1980; Lambeck & Hopgood 1982; Barnes *et al.* 1983; Hide, this symposium), and the long-term accelerations largely attributable to tidal friction and melting of the Pleistocene ice sheets (Morrison & Stephenson 1982; Stephenson & Morrison, this symposium), lie the 'decade' fluctuations in the l.o.d. (figure 2). These are so large (e.g. a decrease in m_3 of 7×10^{-8} between 1870 and 1900) that only the liquid core is both massive and mobile enough to provide the requisite source or sink of angular momentum. The maximum torque required to act across the core–mantle boundary (c.m.b.) is 10^{18} Nm (Morrison 1979), and changes in torque comparable with this must take place in a decade to account for the change in m_3 around 1900. Such torques correspond to an average tangential stress of 2×10^{-3} N m^{-2} on the c.m.b.

Viscous friction will give rise to a stress $\mathscr{S}_\mathrm{v} \approx \rho\nu u/\delta$ where ρ, ν are the density and kinematic viscosity of the core, u is the speed with which the outer core slips past the mantle, and $\delta \approx (\nu/\Omega)^{\frac{1}{2}}$ is the boundary layer thickness (Hide 1977). The westward drift of the geomagnetic field yields an estimate for $u \approx 4\times10^{-4}$ m s^{-1}. Physically based estimates of ν range over values $\nu \approx 4\times10^{-7}$ m^2 s^{-1} (Gubbins 1976), 10^{-5} m^2 s^{-1} (Gans 1972), and 10^{-4} m^2 s^{-1} (Bukowinski & Knopoff 1976). Only with the last value is there some indication that viscous friction may play a more than negligible role in core–mantle coupling. Electromagnetic and topographic coupling have seemed much more effective mechanisms for exciting such large variations in m_3.

Electromagnetic coupling arises as the lower mantle (a semiconductor at the prevailing temperatures) is penetrated by changing magnetic field from the liquid core. The associated tangential stress $\mathscr{S}_\mathrm{M} \approx B_\mathrm{r} B_\mathrm{t}/\mu_0$ where B_r, B_t are the radial and tangential components of the magnetic flux density and $\mu_0 = 4\pi\times10^{-7}$ H m^{-1}. B_t can be poloidal or toroidal. The latter can be produced either as part of the dynamo operation in the core interior, or at the c.m.b. as the outer core slips past the mantle. Thus electromagnetic coupling is both direct (because of leakage of the geomagnetic secular variation (g.s.v.) into the lower mantle) and passive (because of the generation of toroidal magnetic field as the outer core accelerates past the mantle). By Lenz's law the latter will tend to reduce such accelerations. The critical coupling parameter is the radial distribution of electrical conductivity $\lambda(r)$ in the conducting part (the lower 2000 km) of the mantle. Estimates of $\lambda(r)$ are based on the g.s.v. spectrum observed at the Earth's surface after being filtered by diffusion through the conducting lower mantle, but the inversion is far from unique because nothing is known about the input spectrum at the c.m.b. ($r = b = 3485$ km). Obviously, however, the higher the average electrical conductivity $\langle\lambda\rangle$ in the deep mantle, the more the high-frequency end of the g.s.v. will be screened from view at magnetic observatories.

Recent discussions of $\lambda(r)$ have focussed on the 1969–1970 worldwide 'jerk' or step-change in the second time derivative, in the magnetic field from the core, accomplished in less than 2 years. The rapidity with which this phenomenon penetrated the mantle led Achache *et al.* (1980) to an upper bound of 150 Ω^{-1} m^{-1} on $\langle\lambda\rangle$. A far more sophisticated analysis by Backus (1983) suggests that the magnetic 'jerk' permits $\lambda(r) = (b/r)^\gamma\lambda(b)$, with $\gamma = 16$ and $\lambda(b) = 3\times10^3$ Ω^{-1} m^{-1}, in the same region. This gives $\langle\lambda\rangle = L^{-1}\int_b^{b+L}\lambda(r)\,\mathrm{d}r \approx 400$ Ω^{-1} m^{-1} with $L \approx 2000$ km. These values are still far enough below the core conductivity (greater than 10^5 Ω^{-1} m^{-1}) for Hide's (1978) method of magnetically locating the c.m.b. radius to work, but they permit significantly tighter coupling than the electrical conductivity distributions favoured in the 1960s, for which $\langle\lambda\rangle \approx 100$ Ω^{-1} m^{-1}.

Electromagnetic coupling is governed by two timescales (Roberts 1972): τ_1, screening time for diffusion of a g.s.v. step through the lower mantle, and τ_2, the time constant for passive coupling. For a mantle conductivity profile which steepens rapidly towards the c.m.b., i.e. for $\gamma \gg 1$, $\tau_1 \approx \mu_0\lambda(b)\,\gamma^{-2}b^2$ and

$$\tau_2 \approx 4\pi C_\mathrm{m}\,C_\mathrm{c}\,\gamma/C\lambda(b)\,b^3\int_{S_\mathrm{c}} B_\mathrm{r}^2\,|\boldsymbol{e}_3\times\boldsymbol{r}\,|^2\,\mathrm{d}S$$

(Braginskii & Fishman 1976). Here $C_\mathrm{m} = C - C_\mathrm{c}$ is the polar moment of inertia of the mantle and $\boldsymbol{r}$ is the unit normal to the surface S_c of the c.m.b. The rise time of a change in m_3 following a perturbation in the leakage coupling torque is $\tau_1^{\frac{1}{3}}\tau_2^{\frac{2}{3}}$ (Roberts 1972), so it is shortened to

8 years when Backus's conductivity distribution is used. Also an increase in $\langle\lambda\rangle \approx b\lambda(b)/\gamma L$ permits greater leakage torque Γ_L just as reduced τ_2 increases passive coupling torque Γ_P (which varies as $1/\tau_2$). The higher harmonics of the geomagnetic field are more rapidly attenuated geometrically as r increases from the c.m.b., i.e. the spatial spectrum of the geomagnetic field is whiter there than at the Earth's surface. Since the higher harmonics contribute proportionately more to the coupling torque the steeper the gradient in λ, and since they are subject to more rapid g.s.v. than the lower harmonics, the Backus conductivity profile leads not only to much greater average Γ_L and Γ_P than for a uniform conductivity distribution, but also to greater fluctuations in Γ_L and Γ_P in a given time interval.

By considering harmonics up to degree 12 for the 1975 International Geomagnetic Reference Field, Stix (1982) found $\Gamma_P = 5\times10^{18}$ N m for $\gamma = 12$, $\lambda(b) = 1.8\times10^3\ \Omega^{-1}\ \mathrm{m}^{-1}$. For the Backus conductivity profile, the corresponding value of $\Gamma_P = 7\times10^{18}$ N m. On average we expect a balance between Γ_L and Γ_P. It seems reasonable to conclude that if Backus's estimate of the lower mantle conductivity distribution is acceptable, an unbalanced electromagnetic coupling torque of about 15% of Γ_P, i.e. 10^{18} N m, can be brought to bear within a decade by the g.s.v., as necessary to account for the largest recorded decade variation in the l.o.d.

The excellent agreement between short-term fluctuations in spin rate and the simultaneous changes in axial angular momentum of the atmosphere (Hide *et al.* 1980; Lambeck & Hopgood 1982; Rosen & Salstein 1983) suggests that core–mantle coupling becomes ineffective on time scales shorter than 5 years. This might have been anticipated from Morrison's (1979) spectral analysis of the l.o.d. variations during 1861–1978, which showed that the power in the period range longer than 10 years is an order of magnitude greater than that in the period range from 2 to 10 years.

Irregularities in topography at the c.m.b., supposed to explain the spatial correlations between the Earth's gravity and magnetic fields, could provide the equivalent of the 'mountain torque' exerted on the Earth's surface by the zonal winds. As the core fluid flows past

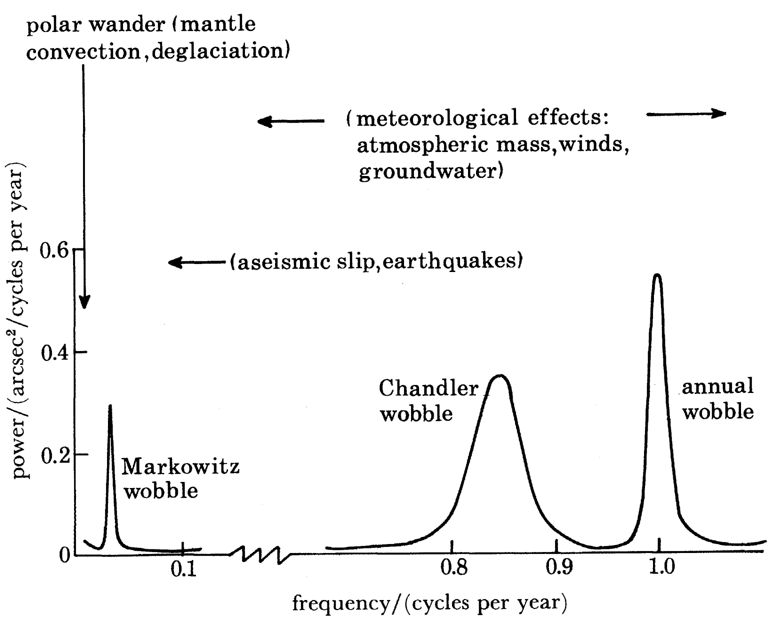

FIGURE 3. Polar motion spectrum.

undulations of great horizontal extent (*ca.* 10^6 m) and height (or depth) h, dynamic pressure forces will produce a mean tangential stress $\mathscr{S}_T = C_T \rho \Omega u (h - \delta)$ on the c.m.b., provided $h > \delta$ (Hide 1977). The drag coefficient C_T depends in a complicated way on the geometry of the undulations and on the relative strengths of toroidal and poloidal magnetic field in the core, but Hassan & Eltayeb (1982) find that bumps of a few kilometres in height could produce $C_T \approx 0.001$, i.e. $\mathscr{S}_T$ of order 10^{-3} N m^{-2}, provided that, below the c.m.b., the ratio of toroidal to poloidal field strength does not greatly exceed unity. Whether such stresses integrate to give a torque of the size needed to explain the decade fluctuations in the l.o.d. will depend of course on their distribution over the c.m.b. (well over 100 of this lateral scale could be located there). Unfortunately irregularities of this amplitude are below the resolving power of seismology at this time.

3. Geographical motion of the rotation pole

(*a*) *Long-period polar motion*

The homogeneous I.L.S. data set (Yumi & Yokoyama 1980) has been analysed by Wilson & Vicente (1980) and Dickman (1981), who found that the rotation pole has moved relative to the C.I.O. on a zigzag path towards 80° W since the beginning of this century, at an average rate of 0.003″ a^{-1}. Because of Mizusawa's longitude and uninterrupted operation this polar wander is heavily dependent on data from that I.L.S. observatory, which is located in a tectonically active region. Fedorov & Rasulov (1981) have argued that even a null secular drift of the pole during this time cannot be excluded on statistical grounds. Assuming the drift is real, a series of recent studies (Nakiboglu & Lambeck 1980; Yuen *et al.* 1983; Wu & Peltier 1984) shows that it can be explained by rearrangement of the inertia tensor of the solid Earth and oceans, as the melting of the late Pleistocene ice sheets causes changes in sea level and redistribution of the surface load on the visco-elastic Earth.

Another controversial feature of the polar motion spectrum (figure 3) is the retrograde long-period wobble originally discovered in the I.L.S. record by Markowitz, and found by Wilson & Vicente (1980) and Dickman (1981) to have period *ca.* 30 years and amplitude *ca.* 0.03″. Although the alignment of the semi-major axis of its highly elliptical path is poorly determined (10°–60° E), observations at the Ukiah I.L.S. station seem to have weighed heavily in its determination, and Okamoto & Kikuchi (1983) have suggested that this 'Markowitz wobble' may really reflect poorly modelled local phenomena at Ukiah, which is very near the San Andreas fault. The problem of how to excite such a free mode, if it exists, remains unsolved. Dickman (1983) has proposed that it may be a natural mode of the coupled mantle–ocean system.

(*b*) *Chandler wobble*

The most prominent feature of polar motion is the Chandler wobble, with r.m.s. amplitude *ca.* 0.15″ and a period of around 435 days. From time to time, the breadth of the Chandler peak has been attributed to a time-varying period (Carter 1982) or two or more nearly equal periods (Chao 1983). By numerical experiments Okubo (1982*a*) has confirmed that these interpretations are artefacts of the variety of spectral analyses applied to the geophysically more reasonable model of the Chandler band, regarded as the result of random excitation of a single-period damped oscillator. Statistical estimates derived from spectral analyses of the polar motion yield values for Q_w, where $2\pi/Q_w$ is the fraction of wobble energy dissipated per Chandler cycle. The reliability of these estimates is questionable because of the relatively short length of the

I.L.S. record, the non-stationary (and still apparently largely unknown) nature of the excitation processes, and the absence of a convincing criterion for choosing the prediction error filter length when using the maximum entropy method to detect periodic constituents. Probably the best estimate so far is by Okubo (1982*b*), who finds $Q_w \approx 50$–100, in agreement with other recent values (Lambeck 1980, p. 102).

Smith & Dahlen (1981) show that the discrepancy between the observed Chandler period and that for a rigid Earth ($2\pi/\sigma_r$) is reduced to about 8 days by the combined effects of core fluidity (largely decoupling the core from the mantle, thus reducing the period), mantle elasticity (which lengthens the period by permitting a partial adjustment of the Earth's equatorial bulge to a rotation axis not aligned with $\boldsymbol{e}_3$), and the non-global ocean (assumed to respond in equilibrium fashion to the wobbling mantle). Dickman (1979) suggested that a non-equilibrium enhancement of the ocean pole tide accompanying wobble would account for this discrepancy. From an examination of I.L.S. and B.I.H. data, together with more recent observations of polar motion by Doppler and laser satellite tracking, Guinot (1982) concluded that the ellipticity of the Chandler pole path is several times larger than an equilibrium pole tide in the non-global ocean would predict. However, numerical solutions of the appropriate Laplace tidal equations show that departures from equilibrium should be negligible at the Chandler period (Carton & Wahr 1983; O'Connor & Starr 1983).

Smith & Dahlen (1981) argue that a more attractive explanation of the 8-day discrepancy is dispersion, the decrease in mantle rigidity with increasing period, due to anelasticity. If the mantle Q varies as σ^α, they find that to account for the Chandler period $0.04 \lesssim \alpha \lesssim 0.19$. Molodenskii & Zharkov (1982) independently confirm this calculation. However, the relation of mantle Q to Q_w is another controversial subject. If mantle anelasticity is to damp the Chandler wobble, the ratio Q_w/Q must be large enough to take into account that the elastic strain energy being dissipated by friction in the mantle during wobble is smaller than the total (largely kinetic) wobble energy. Several recent calculations (Smith & Dahlen 1981; Okubo 1982*b*; Ben-Menahem 1982) use different energetic or dynamic arguments to conclude that $Q_w/Q \approx 1.8$–2.1, in sharp contradiction to the ratio 8.8 obtained by Merriam & Lambeck (1979). The reasons for such contrasting results have been discussed by Smith & Dahlen and by Okubo but are not yet clear. The question is important because the larger the ratio Q_w/Q the larger either α or Q_w must be. The argument is irrelevant if the oceans contribute significantly to damping of the Chandler wobble, but that now seems less likely than it did (Carton & Wahr 1983).

The identification of mechanisms capable of exciting the Chandler wobble is still beset with difficulties. Although the widespread displacement fields accompanying a major earthquake (modelled as an elastic dislocation) appear able to change the Earth's inertia tensor enough to shift the pole by 0.01–0.02″ (Mansinha *et al.* 1979), very large earthquakes are sufficiently infrequent that their cumulative capacity to maintain the Chandler wobble against damping seems inadequate even with $Q_w = 100$. Seismic excitation of wobble may be more effective if the seismic moments of the largest earthquakes (empirically determined from earthquake magnitude) are underestimated, or if there are significant pre- or post-seismic mass shifts (aseismic slip). The problem of estimating seismic moment could be bypassed if observations of the second-degree spheroidal oscillations produced by an earthquake could be used to infer directly the static second-degree displacement field. Visco-elasticity may enhance seismic generation of wobble by a factor of four or so, but not on a short enough timescale unless there is a thin (under about 100 km) layer of viscosity 10^3–10^4 times lower than that deduced from observations of postglacial rebound (Dragoni *et al.* 1983). Wilson & Gabay (1981) compared the homogeneous I.L.S. and B.I.H. polar motion records with the earthquake excitation series *EQ* for 1901–1970 estimated by O'Connell & Dziewonski (1976). The noisiness of the polar motion data, and perhaps the failure of *EQ* to incorporate aseismic slip, probably explain why they found poor coherence between these time series, and prevent any conclusive estimate of the tectonic contribution to wobble excitation.

The other major contender for a Chandler wobble power source is suggested by the considerable departure from strictly annual periodicity shown by the geographical distribution of atmospheric mass, and the proximity of the annual period to the Chandler resonance. Wahr (1982, 1983) has examined the meteorological and oceanographic contributions to wobble during 1900–1973, and judges that on average only 20–25% of the required power is available from atmospheric mass redistribution and the accompanying pressure-driven ocean loading of the Earth. However, Barnes *et al.* (1983) and Hide, this symposium, conclude by using daily values of the equatorial components of the atmospheric angular momentum vector over about one and later about three Chandler periods and by introducing a new theoretical treatment of the wobble problem, that the observed polar motion might be accounted for without invoking significant contributions from non-meteorological processes.

Runcorn (1982) argues for the possibility that the Chandler wobble can be sustained by electromagnetic torques on the mantle, produced as 'impulses' of g.s.v. that penetrate the c.m.b. on a timescale significantly less than the Chandler period. The essence of his argument is that the impulsive torque required to change the radius of the pole path by 0.04″ in a few months (as consistent with some B.I.H. data) is about 2×10^{-7} of the Earth's total angular momentum, and thus only five times bigger than the impulsive torque of $4 \times 10^{-8}\ C\Omega$ presumably available from the core to generate the largest recorded decade fluctuation in the l.o.d. The gravest difficulty confronting Runcorn's suggestion appears to be that excitation of Chandler wobble by impulsive torques requires most of the power in these torques to be at periods much shorter than the resonant period (1.2 years) whereas Lambeck & Hopgood (1982) have shown that the spectrum of changes in the l.o.d. over 1958–1980 has significant power only at periods over five years once the excitation by zonal winds is removed, and Morrison (1979) found that the spectral power of the l.o.d. variations in the period range of 2 to 10 years is an order of magnitude less than that at the periods appropriate to the timescale of the largest decade fluctuation.

4. Concluding remarks

The relatively recent advent of high-precision space observational techniques (Wilkins, this symposium) and of excellent meteorological, as well as geodynamic, data has greatly improved understanding of the short-term features of the Earth's rotation spectrum (especially the l.o.d.) and allowed inferences to be made about such diverse geophysical phenomena as mantle anelasticity and core–mantle coupling. As the atmospheric data series becomes longer it should be possible to make much more precise statements about the role of core–mantle coupling in the l.o.d. fluctuations. Perhaps in the next decade or so we will see at what periods the rotational inertia of the core becomes entrained by the accelerating mantle.

Some features of interest, such as the change in the l.o.d. with 18.6 year period driven by a zonal tide, polar wander, or the 30 year period Markowitz wobble, will surely emerge from obscurity when we have two to four decades of first-class data on U.T.1, polar motion and atmospheric angular momentum. Other problems (seismic excitation and damping of the Chandler wobble, the role of topographic coupling between core and mantle) require progress to be made in seismology. Superconducting gravimetry may become stable enough to discern changes in J_2 and C directly. Magnetic observations better distributed globally (as by the MAGSAT program, but sustained for a much longer time) may enable the mantle conductivity profile, and hence the contribution of electromagnetic core–mantle coupling to decade variations in the l.o.d., to be more firmly delineated.

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